

Climatic Impacts of Stratospheric Geoengineering with Sulfate, Black Carbon and Titania Injection

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Abstract

In this paper, we examine the potential climatic effects of geoengineering by sulfate, black carbon and titania injection against a baseline RCP8.5 scenario. We use the HadGEM2-CCS model to simulate scenarios in which the top-of-the-atmosphere radiative imbalance due to rising greenhouse gas concentrations is offset by sufficient aerosol injection throughout the 2020-2100 period. We find that the global-mean temperature is effectively maintained at historical levels for the entirety of the period for all 3 aerosol-injection scenarios, though there are a wide range of side-effects which are discussed in detail. The most prominent conclusion is that although the BC injection rate necessary to produce an equivalent global mean temperature-response is much lower, the severity of stratospheric temperature changes ($> +70$ °C) and precipitation impacts effectively exclude BC from being a viable option for geoengineering. Additionally, while it has been suggested that titania would be an effective particle because of its high scattering efficiency, it also efficiently absorbs solar ultraviolet radiation producing a significant stratospheric warming ($> +20$ °C). As injection rates and climatic impacts for titania are close to those for sulfate, there appears to be little benefit in terms of climatic influence of using titania when compared to the injection of sulfur dioxide, which has the added benefit of being well modelled through extensive research that has been carried out on naturally occurring explosive volcanic eruptions.

1 **1 Introduction**

2 The climatic impacts of continued greenhouse gas (GHG) emissions are likely to be severe
3 which has prompted countenance of new strategies for tackling GHG-induced global warming
4 [e.g Collins et al., 2013]. Geoengineering strategies, or large-scale climate interventions that
5 aim to reduce global warming, include strategies to sequester atmospheric carbon dioxide –
6 Carbon Dioxide Removal (CDR) methods, and strategies to reduce solar irradiance at Earth’s
7 surface – Solar Radiation Management (SRM) methods [Shepherd et al., 2009]. Stratospheric
8 Aerosol Injection (SAI), an SRM scheme which has received significant attention, involves
9 the enhancement of the stratospheric aerosol layer in order to reflect more sunlight back to
10 space. This scheme mimics large volcanic eruptions such as Mt Pinatubo in 1991, which
11 injected approximately 15-20 Tg of sulfur dioxide (SO₂) into the tropical stratosphere and
12 induced a globally averaged surface cooling of around -0.3 °C for the following two years
13 [Stenchikov et al., 2002].

14 Sulfate (SO₄) aerosols have featured predominantly in SAI research because of the volcanic
15 analogue (e.g. in the Geoengineering Model Intercomparison Project, GeoMIP [Kravitz et al.,
16 2013]). General Circulation Model (GCM) simulations suggest that, while sufficient sulfate
17 injection could effectively reduce global-mean temperature, possible side effects include
18 changes to regional precipitation [e.g. Bala et al., 2008; Tilmes et al., 2013], ozone [e.g.
19 Tilmes et al., 2009; Pitari et al., 2014], stratospheric dynamics [Aquila et al., 2014] and sea-
20 ice extent [Berdahl et al., 2014]. Precipitation changes could result from changes to the moist
21 static stability of the atmosphere and a concomitant weakening of the hydrological cycle
22 [Bala et al., 2008], and the regional precipitation changes under GeoMIP simulations have
23 been shown to be reasonably consistent across a range of climate models [Tilmes et al., 2013].
24 Ozone concentrations could change as a result of enhanced heterogeneous chemistry on the
25 surface of sulfate aerosols or indirectly by changes to the stratospheric dynamics and
26 chemistry [e.g. Tilmes et al., 2009]. Stratospheric dynamical changes could occur as the result
27 of tropical heating in the sulfate layer and by changes to wave propagation from the
28 troposphere [e.g. Aquila et al., 2014].

29 In order to ameliorate the known side-effects of sulfate injection, some authors have proposed
30 alternative aerosols to sulfate [e.g. Teller et al., 1997]. Crutzen (2006) suggested the possible
31 injection of black carbon (BC), which would mimic hypothetical nuclear winter scenarios.
32 One advantage of BC over sulfate is that less mass would be needed for an equivalent

1 radiative forcing [Crutzen, 2006]. BC particles efficiently absorb solar radiation, unlike
2 sulfate which primarily reflects solar radiation [Ferraro et al., 2011]. Alternatively, minerals
3 such as titania (TiO_2), silica (SiO_2) and alumina (Al_2O_3), which have a high refractive index at
4 wavelengths of peak solar radiative flux (~ 550 nm), have also been suggested [Pope et al.,
5 2012]. Although the use of alternative aerosols is not a new suggestion [e.g. Teller et al.,
6 1997], comparatively little research has been conducted on their potential utility. Kravitz et al
7 (2012) simulated a constant BC injection scenario of 1 Tg/yr in the tropics for small radius
8 ($0.03 \mu\text{m}$) and large radius ($0.15 \mu\text{m}$) aerosols. They found that the small particle BC aerosol
9 scenario produced a global surface cooling of -9.45 °C, but also induced stratospheric
10 warming $> +60$ °C and global ozone loss of 50%. The large particle BC aerosol scenario had a
11 negligible climatic impact. Using a fixed dynamical heating (FDH) code, Ferraro et al (2011)
12 compared the stratospheric heating of sulfate, titania, and BC layers for an equivalent
13 instantaneous radiative forcing. Their results showed a tropical stratospheric warming signal
14 for all the aerosols, though much greater in the case of BC. To date, no work has used a
15 comprehensive fully coupled atmosphere-ocean GCM to directly compare the possible
16 climatic impacts of SAI with alternative aerosols to sulfate, which is the motivation for this
17 research.

18 In this work, we simulate the stratospheric injection of sulfate, titania and BC against a
19 baseline RCP8.5 concentrations scenario using a fully-coupled GCM. Titania is selected to
20 represent an efficient light-scattering aerosol and BC is selected as a light-absorbing aerosol.
21 RCP8.5, which is the high-end carbon-intensive CMIP5 scenario, is selected to give a
22 significant greenhouse effect against which to employ geoengineering, in order to distinguish
23 the climatic impacts specific to each aerosol. Observations have shown that the current global
24 GHG emissions exceed the emissions inherent in RCP8.5 [Peters et al., 2013]; therefore our
25 work could be considered as geoengineering against a business-as-usual scenario.
26 Additionally, the next generation of GeoMIP simulations (GeoMIP6) will utilise a carbon-
27 intensive scenario [Kravitz et al., 2015], hence our work will provide a useful supplement to
28 those results. We chose to inject aerosol at a sufficient rate to counterbalance the Top Of the
29 Atmosphere (TOA) global/annual-mean radiative flux imbalance caused by increasing
30 atmospheric GHGs. Our simulation design is similar to the G3 scenario of the Geoengineering
31 Model Intercomparison Project (GeoMIP), which instead used the RCP4.5 concentrations
32 scenario as its baseline and injected sulfate at a sufficient rate to counterbalance GHG
33 radiative forcing [Kravitz et al., 2011]. We analyse the climate changes in the 2090s with

1 respect to a simulated historical period and discuss impacts on a wide range of meteorological
2 parameters.

3

4 **2 Model**

5 *2.1. The HadGEM2-CCS model*

6 For this investigation, we use the HadGEM2-CCS climate model in a fully coupled
7 atmosphere-ocean mode. HadGEM2-CCS is the high-top configuration of the HadGEM2
8 family of models, and includes a well-resolved stratosphere that is capable of internally
9 generating a realistic Quasi-Biennial Oscillation (QBO) [The HadGEM2 Development Team,
10 2011]. The atmosphere component comprises 60 vertical levels extending to 84km and a
11 horizontal resolution of $1.25^\circ \times 1.875^\circ$ latitude by longitude respectively. The 40-level ocean
12 component has a horizontal resolution of 1° by 1° from the poles to 30°N/S , with the
13 latitudinal resolution then increasing smoothly to 0.33° at the equator [The HadGEM2
14 Development Team, 2011]. For this investigation, GHG concentrations, stratospheric ozone,
15 anthropogenic aerosols and aerosol precursor gases are prescribed following the Coupled
16 Model Intercomparison Project phase 5 (CMIP5) [Taylor et al., 2012] protocol, with historical
17 data from 1860-2005 and RCP8.5 concentrations from 2005-2100. HadGEM2-CCS contains
18 the aerosol module Coupled Large-scale Aerosol Simulator for Studies in Climate
19 (CLASSIC). The module's sulfur cycle is described in detail in Bellouin et al (2011). Briefly,
20 it includes the oxidation of sulfur dioxide (SO_2) to sulfate aerosol in aqueous and gas phase
21 reactions. Sulfate is represented by Aitken, accumulation and dissolved modes, with
22 hygroscopic growth in the accumulation mode following d'Almeida et al (1991). Aerosol size
23 modes are represented by lognormal size-distributions with a prescribed dry-mode median
24 radius (r_m) and geometric standard deviation (σ).

25

26 *2.2 Stratospheric aerosol microphysical and optical properties*

27 For this investigation, stratospheric sulfate is modelled using the *volc2* size-distribution from
28 Rasch et al (2008) for the sulfate accumulation mode, with $r_m = 0.376 \mu\text{m}$ and $\sigma = 1.25$; the
29 relatively large r_m is chosen to reflect the high concentrations of SO_2 injected in this
30 experiment.

1 CLASSIC includes a tropospheric BC scheme with fresh, aged and in-cloud modes [Bellouin
2 et al., 2011]. We introduce an additional non-hygroscopic stratospheric BC component and
3 prescribe a lognormal size-distribution with $r_m = 0.0118 \mu\text{m}$ and $\sigma = 2.0$, which is taken from
4 tropospheric BC observations [Deepak and Gerber, 1983]. We prescribe a density for BC of
5 1000 kg/m^3 and take refractive indices from a World Meteorological Organisation report
6 [Deepak and Gerber, 1983].

7 For stratospheric titania, we assume the non-hygroscopic lognormal size distribution of Pope
8 et al. (2012) with $r_m = 0.045 \mu\text{m}$ and $\sigma = 1.8$. This size-distribution was selected to give the
9 titania aerosol a high scattering efficiency, as shown by Pope et al (2012). We prescribe a
10 density for titania of 4230 kg/m^3 [Pope et al, 2012], and for the refractive indices we follow
11 Ferraro et al (2011) and use the average of the extra-ordinary and ordinary values from
12 Ribarsky (1985).

13 The specific absorption (k_{abs}) and scattering (k_{sca}) coefficients for sulfate (accumulation/dry-
14 mode), titania and BC are plotted in Fig. 1 as a function of wavelength. For sulfate, the
15 specific extinction coefficient (k_{ext}) at 500nm of $3200 \text{ m}^2/\text{kg}$ and single scattering albedo (ω_o)
16 of 1 reflects the non-absorbing properties of sulfate. Although titania's 500nm scattering
17 efficiency ($k_{\text{sca}} = 3850 \text{ m}^2/\text{kg}$) is greater than sulfate's in this instance, titania additionally
18 absorbs SW radiation ($k_{\text{abs}} = 2000 \text{ m}^2/\text{kg}$ at 250 nm, and $k_{\text{abs}} = 600 \text{ m}^2/\text{kg}$ at 500 nm) which
19 can be explained by the band-theory of solids [Yang et al., 2003]. Thus titania is partially
20 absorbing. Our modelled BC efficiently absorbs SW radiation ($k_{\text{abs}} = 8300 \text{ m}^2/\text{kg}$ at 500nm)
21 but also produces a non-negligible SW scattering effect ($k_{\text{sca}} = 2500 \text{ m}^2/\text{kg}$ at 500nm) which is
22 comparable in magnitude to the equivalent scattering efficiency of both titania and sulfate.
23 Therefore, to describe titania as an efficient light-scatterer and/or BC as an efficient light-
24 absorber is an over-simplification.

25 Our choice of particle size and density will impact the aerosol's gravitational sedimentation
26 rate and therefore its atmospheric residence time (the sedimentation rate is also a property of
27 the local atmospheric conditions) [Rasch et al., 2008]. To determine the importance of our
28 choice of aerosol properties, we have calculated the respective gravitational sedimentation
29 rates by using the method of Pruppacher and Klett (1979) (which utilises Stoke's law) and
30 incorporating temperature and pressure values from the International Standard Atmosphere
31 [ICAO, 1993] (Fig. S1 in the Supplement). We find that the average sedimentation rates
32 between 18-26 km altitude for our prescribed sulfate, titania, and BC are 23, 9.5 and 0.75

1 m/day respectively, and the equivalent rates between 26-30 km are 52, 22, and 1.8 m/day.
2 Therefore, one would expect BC to be advected to much higher altitudes than sulfate in these
3 simulations. For perspective, Schoeberl et al (2008) deduced from observations that the
4 atmospheric tropical vertical velocity between 18-26 km has an upper limit of 35 m/day, and
5 the equivalent velocity between 26-30 km is below 61 m/day.

6

7 **3 Method**

8 We first validated the model's stratospheric sulfate scheme by simulating the Mt Pinatubo
9 eruption and then comparing the results with observations. These simulations comprised a 10-
10 member ensemble in which 20 Tg[SO₂] is injected between 16-18 km over a single day in
11 June 1991, following the method of Aquila et al (2012). Figure 2a shows the global/annual-
12 mean sulfate aerosol optical depth (AOD) anomaly for the HadGEM2-ensemble and for
13 AVHRR and SAGE-II observations. The model clearly captures the peak AOD from the
14 AVHRR data, and the exponential decline thereafter. Figures 2b-d show the zonal-mean AOD
15 anomaly for the same time period. The agreement between the model and observed AOD is
16 reasonable. Some differences in the temporal evolution of the AODs in the model and the
17 observations are due to the almost concurrent eruption of Cerro Hudson which injected
18 approximately 3.3Tg[SO₂] into the southern hemisphere [Deshler and Anderson-Sprecher,
19 2006]. This relatively close agreement between observations and HadGEM2 estimates,
20 together with other modelling studies of other volcanic eruptions [Haywood et al., 2010]
21 suggests that the model is a useful tool for stratospheric geoengineering simulations.

22 The geoengineering investigation was based on a 240-year Pre-Industrial Control simulation
23 (forced by constant 1860's GHGs and aerosol emissions) and historical simulations for the
24 period 1860-2005 following CMIP5 [Taylor et al., 2012] protocol followed by RCP8.5
25 emission specified from 2005-2019. Leading on from these simulations, we performed 3-
26 member ensembles for the period 2020-2100 for: RCP8.5 only, RCP8.5 with SO₂ injection
27 (geoSulf), RCP8.5 with TiO₂ injection (geoTiO₂), and RCP8.5 with BC injection (geoBC).
28 Aerosol (or gaseous SO₂ for the geoSulf scenario) was injected at a constant rate between 23-
29 28 km altitude in a single vertical column at the equator. The injection altitude and location
30 were chosen to ensure a long stratospheric lifetime of the aerosol, which is transported
31 poleward by the upper branch of the Brewer-Dobson circulation [Niemeier et al., 2011], and
32 therefore make the geoengineering approach reasonably efficient.

1 We inject aerosol at such a rate as to maintain the top-of-the-atmosphere (TOA) net
2 radiative flux at piControl levels. Specifically, we define the TOA radiative flux
3 Imbalance (TOA-Imb) as the annual/global-mean TOA net radiative flux (incoming SW
4 minus outgoing LW+SW) minus the average TOA net radiative flux of the piControl
5 period. The piControl TOA net radiative flux is positive ($+0.27 \text{ W/m}^2$) as anthropogenic
6 GHGs were emitted prior to 1860 (the piControl reference period). By sufficient aerosol
7 injection, we aim to maintain TOA-Imb=0. This scenario represents our interpretation of
8 ‘equal amount of geoengineering’ for each aerosol. The advantage of returning net
9 radiation to piControl levels (rather than completely equilibrating TOA fluxes) is that
10 piControl had already been simulated comprehensively for CMIP5 (240 model-years),
11 hence permitting robust statistics to be calculated. The TOA radiative imbalance is a
12 metric that satellites are able to measure (e.g. CERES [L’Ecuyer et al, 2015] and
13 EarthCare [Illingworth et al, 2015]), albeit with $\pm 3 \text{ W/m}^2$ accuracy at present [Priestley
14 et al, 2011; von Schuckmann et al., 2016]. Therefore our target could be applicable to an
15 actual SAI scenario. In contrast, adjusted Radiative Forcing (RF) (the net radiation
16 perturbation at the tropopause from some external forcing, after stratospheric
17 adjustment), cannot be directly measured by satellites and therefore it would be difficult
18 to obtain a specified radiative forcing in an actual SAI scenario. Of course, other metrics
19 could be chosen [e.g. MacMartin et al., 2013], with each metric having its own
20 signal/noise characteristic.

21 To determine the injection rates required to maintain TOA-Imb balance, we first conducted
22 15-year atmosphere-only simulations of 1 Tg aerosol (or SO_2 for sulfate) injection per year to
23 calculate the specific radiative effect for each aerosol. We then used the radiative effect to
24 calculate the injection rate necessary to offset the RCP8.5 anthropogenic radiative forcing
25 (ARF) for the 2020-2100 period (with ARF values from Meinshausen et al (2011)). We used
26 the ARF to estimate the injection rates required to produce TOA-Imb=0 as this seemed a
27 sensible method for approximating the necessary aerosol injection. As the geoengineering
28 simulations progressed, we altered the injection rate when necessary to ensure that TOA-Imb
29 balance was maintained (Fig. S2 in the Supplement). This feedback-orientated method is
30 similar to the methods suggested by MacMartin et al. (2014) and Kravitz et al (2014). A
31 detailed description of our methods is provided in the supplementary material (Section S2).

1 Our analysis focuses initially on the temporal evolution of the TOA-Imb and global mean
2 temperature changes to show that our simulations provide plausible counterbalances to global
3 mean temperature changes under RCP8.5. However, our main focus is on the differences
4 between a recent historical period (1980-2005) (hereafter denoted HIST) and the
5 geoengineering experiments during the period 2090-2100, with an emphasis on different
6 geographical patterns. As we were not explicitly attempting to reach a specific global
7 mean temperature, the choice of reference period was left until after the geoengineering
8 simulations had been completed. We then selected a recent historical period from which
9 the 2090s global-mean temperature anomaly for geoSulf was negligible (Fig. 3b). The
10 HIST period selected is close to the historical control period used in the IPCC AR5 report
11 (1986-2005) [e.g. Fig. 12.10 from Collins et al., 2013] which facilitates comparison of our
12 RCP8.5 climate changes with the CMIP5 multi-model means.

13

14 **4 Results**

15 *4.1 Effectiveness at maintaining global mean TOA-Imb and near surface temperature*

16 Figure 3 shows the global/annual-mean TOA-Imb and near-surface air temperature anomaly
17 for the geoengineering and RCP8.5 simulations, with respect to the HIST period. For all of
18 the geoengineering simulations we were able to maintain $\text{TOA-Imb} \approx 0$ for the entirety of the
19 80-year period (Fig. 3a). For geoSulf, geoTiO₂ and geoBC, the TOA-Imb was maintained
20 within ± 0.21 , ± 0.18 and $\pm 0.20 \text{ Wm}^{-2}$, respectively (1 standard deviation throughout the
21 2020-2100 period).

22 The near-surface global temperature response differs between the aerosols with a greater
23 cooling trend for sulfate than for titania or BC (Fig. 3b). To determine the cause of the
24 anomalous warming in geoBC, we assess the net radiative flux at the top of the
25 atmosphere for 2020-2100. Fig. S3 in the Supplement shows the global-mean net-
26 downward radiative flux anomaly for the geoengineering experiments, evaluated at the
27 TOA and the tropopause; and the global-mean net-downward heat flux anomaly at the
28 surface. The radiative flux changes at the TOA and tropopause, and the heat flux
29 anomaly at the surface, are comparable for the geoSulf and geoTiO₂ experiments for the
30 duration of 2020-2100. In contrast, geoBC exhibits an increasingly positive net radiative
31 flux anomaly at the tropopause ($+0.2 \text{ W/m}^2$ averaged over 2020-2100) despite the
32 negligible TOA radiative flux anomaly. After stratospheric temperature adjustment,

1 radiative perturbations at the TOA and tropopause are equal for a given climate forcing,
2 which implies that the consistently non-adjusted stratosphere (due primarily to increasing
3 aerosol injection rates) is responsible for the differences in TOA and tropopause radiative
4 perturbations in geoBC. Hansen et al (1997) also found that instantaneous and adjusted
5 radiative forcing differ most when there is a large heating affecting the tropopause. This
6 implies that if we had injected aerosol sufficiently to produce an equal radiative effect at
7 the tropopause, the temperature trends for the geoengineering experiments in Fig. 3
8 would have been more comparable. If we were to choose stabilisation of temperature as our
9 basic metric, then one could approximate the results by simply scaling the results by the ratio
10 of the temperature perturbation relative to 1980-2005 to that for geoSulf. The scaling would
11 be 1 (by design) for geoSulf, 1.1 for geoTiO₂ and 1.28 for geoBC. If the metric chosen were
12 instead to keep the global mean precipitation the same, then the scaling would be 1 (by
13 design) for geoSulf, 0.91 geoTiO₂ and 0.68 for geoBC. However, we shall see that the
14 changes in many of the variables we consider are dominated by large scale changes in the
15 spatial patterns of response rather than the 10-30% changes in magnitude of the response that
16 applying such a scaling would induce. We therefore choose to present un-scaled results here
17 but caveat that such a scaling could be applied should we wish to apply a different metric.
18 From Fig. 3b, geoSulf exhibits a near-surface air cooling trend with respect to 2020 despite a
19 net gain of atmospheric energy, which is likely due to an uneven vertical distribution of this
20 energy gain.

21 Fig. 3c shows the global mean precipitation anomaly with respect to the HIST period.
22 The precipitation reduction is greater for BC than for sulfate and titania, despite the
23 positive temperature trend in geoBC (Fig. 3b). The hydrological sensitivity to
24 geoengineering, defined as the global mean precipitation change per unit temperature
25 change, is 2%/°C for sulfate, 2.5%/°C for titania, and 4.6%/°C for BC. The hydrological
26 sensitivity for RCP8.5 is 1.32 %/°C, which is close to the CMIP5 ensemble-mean [Fig. 12.7
27 from Collins et al., 2013]. For comparison, Bala et al (2008) found a hydrological
28 sensitivity of 2.4%/°C for solar irradiance reduction and 1.4%/°C for CO₂ increase.

29

30 **4.2 Aerosol distribution**

31 The time-averaged injection rates for the 2090s period are 14 Tg[SO₂]/yr, 5.8 Tg/yr and 0.81
32 Tg/yr for geoSulf, geoTiO₂ and geoBC, respectively. This SO₂ injection rate is approximately

1 equivalent to 1 Mt Pinatubo eruption per year [Dhomse et al., 2014]. These injection rates
2 equate to global aerosol mass-burden anomalies of 49.5, 20.2, and 5.1 Tg for geoSulf,
3 geoTiO₂ and geoBC, respectively. The geoBC mass burden is comparable to the equilibrium
4 burdens of the high-altitude (HA) and small-radius (SmR) experiments from Kravitz et al
5 (2012), although they injected BC at a constant rate of 1 Tg/yr, around 20% higher than in our
6 study. Their SmR experiment involved a 10-year injection of BC particles with a uniform
7 radius of 0.03 μm, into a region between 100-150 mb altitude and over the latitude range
8 10°S-10°N, against baseline perpetual year 2000 conditions. Figure 4 shows the 2090s annual,
9 June-July-August (JJA) and December-January-February (DJF) aerosol mass concentration
10 anomalies (annual mean aerosol optical depths are shown in Fig. S4 in the Supplement). Peak
11 sulfate concentrations are found at the injection region at the equator (Figs. 4a,d,g) and over
12 the winter pole. Titania and BC reach greater altitudes than sulfate (>50 km), which is due to
13 their smaller size-distributions and self-lofting from SW-absorption [Kravitz et al., 2012].
14 While sulfate aerosol concentrations are highest at the equator, the highest concentrations of
15 BC are found in the polar stratosphere. This is because the larger particle size of the sulfate
16 aerosol is subject to a larger sedimentation velocity (Fig. S1 in the Supplement) and thus a
17 greater fraction of aerosol is removed close to the source region. The results from titania
18 suggest a spatial distribution intermediate between sulfate and BC owing to the intermediate
19 size distribution.

20 Figure 5 shows the total annual, JJA and DJF aerosol deposition anomalies averaged over the
21 2090s (the seasonal cycle of the deposition anomalies are shown in Fig. S5 in the
22 Supplement). Sulfate is predominantly deposited in the Northern Hemisphere (NH)
23 extratropics in the boreal spring and summer (Fig. 5d) which is likely attributable to
24 tropopause fold events in the lower branch of the Brewer-Dobson circulation (BDC) [Kravitz
25 et al., 2012]. In contrast, Titania and BC are primarily deposited at high latitudes in the polar
26 winter, which is attributable to the diabatic descent of air in the deep branch of the BDC [e.g.
27 Tegtmeier et al., 2008]. Kravitz et al (2012) also found in their SmR experiment that BC
28 deposition was limited to the polar regions, but their maximum deposition was during polar
29 summer rather than polar winter. The global/annual-mean deposition rates of sulfate and BC
30 from geoengineering are 37 and 1.5 mg/m²/yr, respectively. These amounts may be compared
31 with 231 and 12.7 mg/m²/yr from non-geoengineering sources, amounting to increases of 16
32 % and 12 % respectively. The global/annual-mean deposition rate for titania is 11 mg/m²/yr.

1

2 ***4.3 Temperature and precipitation***

3 Figure 6 shows the annual mean near-surface air temperature (Figs. 6a-d) and precipitation
4 anomalies (Figs. 6e-h) with respect to HIST. RCP8.5 (Fig. 6a) shows the typical global
5 warming signal of amplified warming at high-latitudes due to temperature feedbacks
6 [Pithan and Mauritsen, 2014] and the surface-albedo feedback [e.g. Kharin et al., 2013].
7 This results in an annual mean warming of +11.3 °C averaged over the Arctic region (> 60
8 °N) and an average NH land warming of +7.3 °C. This figure provides an alarming picture of
9 the change in global mean temperature by the end of this century should global society follow
10 the RCP8.5 (essentially a business as usual) pathway. All 3 SAI experiments produce a
11 surface-cooling with respect to RCP8.5, with geoSulf exhibiting the greatest global-mean
12 cooling effect of -4.85 °C, considering TOA-Imb is balanced for each geoengineering
13 experiment. The latitudinal distribution of cooling varies markedly between the SAI
14 experiments, with relative tropical cooling for geoSulf and geoTiO₂ (Figs. 6b,d) and polar
15 cooling for geoBC (Fig. 6c). Defining the ‘SAI cooling effect’ as the temperature difference
16 between SAI and RCP8.5, the ratio of cooling effect at high latitudes (> 60°) between geoBC
17 and geoSulf is 1.19 and between geoBC and geoTiO₂ is 1.23. In the tropics and mid-latitudes
18 (< 60°) the equivalent ratios are 0.64 and 0.71 respectively. The high-latitude cooling in the
19 case of geoBC is attributable to the zonal distribution of BC (Figs. 4c,f,i) which is more
20 evenly spread over the stratosphere than for geoSulf and geoTiO₂. The result is a greater
21 surface SW forcing at high-latitudes in the summer hemisphere for geoBC. For instance, in
22 the Arctic (>60°N) in JJA, the surface SW forcing is -25.65 Wm⁻² in geoBC and -3.3 and -
23 6.55 Wm⁻² in geoSulf and geoTiO₂ respectively. Although the global-mean precipitation rate
24 increases for the RCP8.5 scenario (Fig. 6e), certain regions such as the Amazon basin exhibit
25 a drying trend. This is in line with the CMIP5 multi-model projections documented in the
26 Intergovernmental Panel on Climate Change 5th assessment report (IPCC AR5) [e.g. Fig.
27 12.10 from Collins et al., 2013]. All of the SAI experiments show a global-mean precipitation
28 reduction with respect to both HIST and RCP8.5 (Figs. 6f-h), which is due to the deceleration
29 of the hydrological cycle and is a robust model response to SAI [e.g. Yu et al., 2015; Tilmes
30 et al., 2013; Bala et al., 2008]. The magnitude of the precipitation changes are greater for
31 geoBC than for geoSulf or geoTiO₂; for instance, the global mean precipitation anomaly is -
32 0.26 mm/day for geoBC compared to -0.12 mm/day for geoSulf and -0.14 mm/day for

1 geoTiO₂. In order to maintain TOA-Imb=0, BC must produce a greater SW perturbation
2 at the tropopause and at the TOA than sulfate or titania, which is compensated by the
3 increased LW perturbation resulting from stratospheric warming. The troposphere is
4 relatively transparent to SW radiation but absorbs efficiently in the LW spectrum,
5 therefore the annual-mean surface radiative forcing in the geoBC experiment is greater
6 (-18.6 W m^{-2}) than for geoSulf or geoTiO₂ (-7.4 and -9.6 W m^{-2} respectively – see Fig.
7 S6 in the Supplement). Bala et al (2008) and Muller and O’Gorman (2011) have shown that
8 the magnitude of the global-mean precipitation response to an imposed forcing is dependent
9 on the energy flux entering/leaving the atmosphere (the radiative forcing of the atmosphere).
10 The radiative forcing of the atmosphere is the difference between net radiative fluxes at the
11 TOA and at the surface. As the net radiative flux anomaly at the TOA is, by design, equal for
12 the different geoengineering scenarios here and the net radiative flux anomaly at the surface is
13 greater for geoBC (Fig. S6 in the Supplement), the precipitation reduction is therefore
14 amplified in the geoBC scenario. It is important to note that if the RCP8.5 warming relative to
15 HIST was completely offset in the geoBC and geoTiO₂ experiments, the hydrological
16 response would be greater than in Fig. 6. Using the hydrological sensitivities calculated in
17 section 4.1, the precipitation changes relative to HIST would be -0.34 mm/day for geoBC and
18 -0.16 mm/day for geoTiO₂. From Fig. S6 in the Supplement, the reduction in surface SW
19 flux in the RCP8.5 scenario is due to increases in water vapor [Haywood et al., 2011].
20 Haywood et al (2011) report a clear-sky reduction of -5.7 W/m^2 while our study is
21 consistent at a value of -5.4 W/m^2 (not plotted). However, in all geoengineering cases,
22 this reduction is comprehensively overwhelmed by aerosol direct effects.

23 Figure 7 shows the JJA temperature (Figs. 7a-d) and precipitation (Figs. 7e-h) anomalies. In
24 the geoSulf and geoTiO₂ scenarios, the temperature is effectively maintained at HIST levels
25 (Figs. 7b,d). However, a slight bias towards high-latitude NH warming in geoSulf and
26 geoTiO₂ results in a northward displacement of the Inter-Tropical Convergence Zone (ITCZ),
27 which is exemplified by the Sahelian precipitation increase in Figs. 7f,h. This phenomenon
28 was noted by Haywood et al (2013) and has been observed after large hemispherically
29 asymmetric volcanic eruptions [Oman et al., 2006]. Although the general pattern of
30 precipitation change is similar for the 3 SAI scenarios, geoBC again displays a greater drying
31 signal, with 80% of the total land area experiencing a JJA precipitation reduction in geoBC
32 compared to 70% for geoTiO₂, 57% for geoSulf and 52% for RCP8.5.

1 Figure 8 shows the DJF temperature (Figs. 8a-d) and precipitation (Figs. 8e-h) anomalies. The
2 temperature reduction over Greenland in geoBC (Fig. 8c) is due to the significant decrease in
3 downwelling SW radiation at the surface during the Arctic sea-ice formation season
4 (September-October-November), which leads to a positive sea-ice albedo feedback and
5 further localised cooling. This inference is corroborated by Fig. 9, which shows the Arctic
6 DJF sea-ice extent in terms of the average DJF sea-ice boundary (the Antarctic DJF sea-ice
7 extent is shown in Fig. S7 in the Supplement). The sea-ice boundary in geoBC (Fig. 9c)
8 extends to well below Greenland, and the total sea-ice extent anomaly is +1.72 million km²
9 which vastly exceeds the HIST standard deviation of +/- 0.52 million km². In comparison, the
10 sea-ice extent anomaly of -11 million km² for RCP8.5 (Fig. 9a) marks a reduction by 43% of
11 the total HIST sea-ice extent. Returning to Fig. 8, the poleward shift in the NH extratropical
12 rain-belt over the Atlantic in RCP8.5 (Fig. 8e) is a robust result of GHG-induced global
13 warming and is related to storm track displacement [Lombardo et al., 2015]. This same
14 response is evident in the geoengineering simulations (Figs. 8f-h), although to a much lesser
15 extent in geoSulf and geoTiO₂.

16

17 ***4.4 Stratospheric changes***

18 Figure 10 shows the zonal-mean temperature change as a function of latitude and altitude for
19 the JJA and DJF seasons. The stratospheric cooling in conjunction with tropospheric warming
20 in RCP8.5 (Figs. 10a,e) is a robust result of increasing GHG-concentrations [e.g. Schmidt et
21 al., 2013]. Aerosols directly affect temperature by absorbing radiation, and indirectly by
22 scattering radiation and by ambient dynamical and chemical changes [Carslaw and Kärcher,
23 2006]. Sulfate predominantly absorbs in the LW and near-infra-red spectrum (Fig. 1a).
24 The stratospheric radiative heating in geoSulf is most pronounced in the tropical region,
25 where sulfate absorbs outgoing LW radiation from the warm troposphere below, and then
26 emits comparatively less radiation from the ambient cold stratosphere [Ferraro et al.,
27 2011]. In contrast, titania and BC absorb in both the SW and LW spectrum (Figs. 1b,c), and
28 therefore preferentially warm the summer-hemisphere and tropical stratosphere, where solar
29 radiation is most prevalent. geoBC produces the most significant warming effect, with an
30 average stratospheric (15-50 km altitude) temperature increase of +33 °C and a maximum
31 temperature increase of +68 °C, which occurs in JJA (Figs. 10c,g). The maximum BC-
32 induced heating relative to the baseline RCP8.5 scenario is +76 °C (Fig. S8 in the

1 Supplement), which is comparable to the ~ 80 °C temperature change Kravitz et al (2012)
2 found in their SmR scenario. For comparison, the maximum sulfate-induced and titania-
3 induced heating relative to RCP8.5 are far more modest at +7 °C and +22 °C, respectively.

4 A warming of the lower tropical stratosphere could have multiple climatic repercussions such
5 as a weakening of the tropospheric tropical circulation [Ferraro et al., 2014], strengthening of
6 the polar vortex [Driscoll et al., 2012] and modification of the QBO [Aquila et al., 2014].
7 Additionally, an increase in the Tropical Tropopause Layer (TTL) temperature would increase
8 the specific humidity of air entering the stratosphere [Dessler et al., 2013]. Changes to the
9 stratospheric water vapor content could have significant chemical and radiative impacts,
10 contributing to ozone depletion via the HO_x cycle and stratospheric warming via LW-
11 absorption [Kravitz et al., 2012]. To assess the effects of geoengineering on stratospheric
12 water vapor, we calculate the time-averaged H₂O mixing ratio averaged between 20°S-
13 20°N and 16-20 km altitude. In the HIST era, the H₂O MMR is 4.2 ppmv, in close
14 agreement with HALOE observations [Gettelman et al., 2010]. In the 2090s, the average
15 H₂O MMR is 6.3 ppmv for RCP8.5, 4.8 ppmv for geoSulf, 7.1 ppmv for geoTiO₂, and
16 32.7 ppmv for geoBC. The stratospheric water vapor feedback is therefore greater for
17 geoBC and geoTiO₂ than for geoSulf.

18 A strengthening of the polar vortex could be instigated by an increased temperature gradient
19 between the tropical/mid-latitude and polar stratospheres, a phenomenon which was observed
20 after the Pinatubo eruption [Stenchikov et al., 2002]. We concentrate on the Arctic wintertime
21 (DJF) response to SAI, and adopt a similar metric to that used by Ferraro et al (2011) to
22 determine the stratospheric temperature gradient. Explicitly, we determine the difference in
23 temperature between 20°N-20°S (Tropics) and 50°N-90°N (North Pole) at 17-22 km altitude
24 in the DJF season. Using this metric, the change in temperature gradients for geoBC, geoSulf
25 and geoTiO₂ are +10.4 °C, +7 °C, and +10.1 °C, respectively, indicating a steeper temperature
26 gradient between the tropics and poles. Additionally, Fig. 11 shows the 50hPa DJF
27 geopotential height anomalies over the Arctic for RCP8.5 and the 3 SAI experiments. The
28 negative geopotential height anomaly centered over the North Pole in all the SAI experiments
29 is indicative of a strengthened polar night jet and a positive Arctic Oscillation phase
30 [Stenchikov et al., 2002]. The DJF zonal-mean zonal-wind anomaly (Fig. S9 in the
31 Supplement) substantiates our inference of a strengthened polar-night jet under SAI, with

1 increased zonal windspeeds at 65°N / 40km altitude of 62 m/s, 17 m/s, and 37 m/s for geoBC,
2 geoSulf, and geoTiO₂ respectively.

3 The Quasi-Biennial Oscillation (QBO) is a periodic change in the equatorial zonal wind
4 pattern in the stratosphere, which fluctuates between easterly and westerly-shear phases
5 [Baldwin et al., 2001]. Aquila et al (2014) showed that radiative heating in the aerosol layer
6 could prolong the westerly-phase of the QBO (where the phase is defined at 40 hPa) by
7 enhancing the residual-mean upwelling motion and strengthening the westerly winds.
8 HadGEM2-CCS includes a non-orographic gravity wave scheme that permits the model to
9 internally generate a QBO and is therefore capable of assessing QBO changes [The
10 HadGEM2 Development Team, 2011]. The average QBO period for the HIST-era ensemble
11 is 27 months (Fig. S10 in the Supplement) which agrees closely with observations [e.g.
12 Baldwin et al., 2001]. Figure 12 shows the 2090s QBO timeseries for one ensemble member
13 of the RCP8.5 and SAI experiments (Figs. S11a,b in the Supplement show the QBO
14 timeseries for the other 2 ensemble members). The average QBO periods for this timespan,
15 which are determined using all 3-ensemble members, are 20 months for RCP8.5, 31 months
16 for geoSulf and 36 months for geoTiO₂. For geoBC, no QBO-like oscillation can be
17 detected in the 10-year time span, suggesting a persistent westerly-phase such as observed
18 by Aquila et al (2014) in their 5 Tg[SO₂]/yr injection scenario (G₅^{22-25km}) scenario. In their
19 HadGEM2-CC simulations, Kawatani and Hamilton (2013) also observed a decline in the
20 QBO period for the RCP8.5 scenario, although they were unable to provide a reason for this.
21 A robust inference from this work is that the magnitude of SAI's impact on stratospheric
22 zonal winds correlates with the magnitude of the stratospheric warming.

23

24 **5 Discussion**

25 In this work, we have assessed the climatic impacts of sulfate, black carbon and titania-
26 injection against a baseline RCP8.5 scenario, by comparing the 2090s climate with a
27 simulated historical period. We have shown that, although the distribution of climate changes
28 are similar for the 3 SAI scenarios, the magnitude of the changes differ, for instance BC
29 produces a substantially greater stratospheric warming signal with concomitantly greater
30 changes to stratospheric dynamics. The severity of the stratospheric temperature changes
31 effectively excludes BC from being a viable option for geoengineering. Additionally, we
32 have shown that producing an equivalent top of the atmosphere radiative perturbation with a

1 SW-absorbing aerosol such as BC (or to a lesser extent titania) compared to a SW-scattering
2 aerosol such as sulfate, induces a comparatively greater SW forcing at the surface. Bala et al
3 (2008) showed that reduced latent heat fluxes compensate for the SW reduction at the surface,
4 instigating a deceleration of the hydrological cycle that is proportional to the magnitude of the
5 SW reduction. This explains the comparatively greater precipitation reduction exhibited by
6 geoBC in figures 6-8. Our results complement Niemeier et al (2013), who showed that a LW-
7 absorbing sulfate layer would produce a greater hydrological perturbation per TOA SW
8 forcing than a simple solar irradiance reduction scenario. The geoBC scenario displays a
9 greater cooling at high-latitudes than the geoSulf and geoTiO₂ scenarios (Figs. 6-8), which
10 comparatively exhibit a net tropical cooling. This raises the question of whether a
11 combination of aerosols could potentially be injected to produce a latitudinally-uniform
12 cooling if necessary. Although SAI with sulfate and titania effectively maintains the regional
13 distribution of temperature at HIST levels, with a slight residual warming at high latitudes, the
14 hydrological cycle decelerates substantially in all SAI scenarios which is exemplified by a
15 global-mean reduction in precipitation. However, annual-minimum sea-ice extent in both
16 hemispheres and global-mean thermosteric sea-level (Fig. S12 in the Supplement) are almost
17 entirely maintained at HIST levels for all SAI scenarios. The results of our Antarctic sea-ice
18 extent anomalies are comparable to McCusker et al (2015). In particular, both their Fig. 2 and
19 our Fig. S7 in the Supplement show limited spatial retraction of sea-ice in the sulfate scenario.
20 We have used the same criterion as McCusker for determining which gridcells contain sea-ice
21 (sea-ice fraction of >15%), which aids in the comparison. Additionally, both our results and
22 McCusker's show that SAI can reduce Antarctic temperatures substantially (their Fig. 2, our
23 Fig. 6) compared to the RCP8.5 climate.

24 We find that sulfate induces less stratospheric warming than titania. In contrast, Ferraro et al
25 (2011) found that the peak stratospheric warming for titania was approximately a third of that
26 from sulfate. Although the different climatologies, model configurations, and aerosol spatial
27 distributions will contribute to the difference in stratospheric temperature adjustment between
28 our and Ferraro's work, the primary reason for the disparity is likely to be the aerosol size
29 distributions. Our titania is smaller (median radius = 0.045 μm compared to 0.1 μm for
30 Ferraro et al (2011)) and therefore scatters and absorbs SW more efficiently, producing a
31 greater localised 'solar' warming. Their sulfate distribution contains a larger spread ($\sigma = 2.0$
32 for Ferraro et al (2011) compared to $\sigma = 1.25$ here), resulting in more coarse-mode particles
33 and greater LW absorption. This disparity highlights the sensitivity of climatic effects to the

1 specified aerosol size distribution. On a separate note, Ferraro et al (2011) neglected to alter
2 the titania density component in the calculation of their aerosol mass and specific optical
3 properties [A. Ferraro, personal communication]. The density that they used for titania of
4 1600 kg/m^3 is appropriate for sulfate, but should have been altered to $\sim 4000 \text{ kg/m}^3$ for titania.
5 Therefore, their titania aerosol burden should be multiplied by 2.5 to give 7.5 Tg, and their
6 optical coefficients divided by 2.5, to obtain appropriate values.

7 It is important to note that the climate impacts described in section 4 are dependent on the
8 optical properties of the aerosol, which are further dependent on the aerosol particle's size,
9 shape, and composition [e.g. Kravitz et al., 2012]. In this investigation, the dry-mode size
10 distribution of the aerosol species is held constant and hygroscopic growth is not represented
11 in the BC and titania schemes, nor are the effects of internal mixing represented.
12 Observations have shown that fresh BC aerosol is predominantly hydrophobic, but the
13 uptake of soluble particulates (e.g. secondary organics) results in increased
14 hygroscopicity [Liu et al., 2013]. Mineral dust, which contains 1-10% titania by mass
15 [Ndour et al., 2008], exhibits low hygroscopicity for radii $< 0.1 \mu\text{m}$ and similar growth to
16 equivalently-sized sulfate aerosol thereafter [Koehler et al., 2009]. Although the
17 historical stratospheric water vapor content is low ($\sim 4.2 \text{ ppmv}$ in the tropical lower
18 stratosphere during the HIST period), aerosol-induced stratospheric warming in the TTL
19 would increase the specific humidity of air entering the stratosphere, therefore impacting
20 hygroscopic growth. The injection of aerosol into pre-existing aerosol layers would lead to
21 larger particles through coagulation and condensation, which would further alter the aerosol's
22 optical and physical properties. The actual size of the aerosol in an SAI scheme would
23 therefore depend on the injection strategy (e.g. location/ season) and the size and composition
24 of the injected species [e.g. Carslaw and Kärcher, 2006; Heckendorn et al., 2009]. Recent
25 research from Heckendorn et al (2009), Pierce et al (2010), English et al (2012), and
26 Weisenstein et al (2015) have highlighted the importance of representing aerosol growth in
27 SAI simulations. Incorporating aerosol microphysics would result in a better representation of
28 the aerosol's optical properties; this is particularly important for solid aerosols that form
29 chain-like fractals. However, it is also important that the model's climatology is able to
30 respond to radiative changes induced by the aerosol. A more detailed assessment would
31 couple a 3D GCM with a detailed aerosol microphysics module, but such experiments over
32 the centennial timescales of this work are currently too computationally expensive. A detailed

1 assessment of the aerosol microphysics for sulfate, BC, and titania injection is therefore not
2 within the scope of this paper, but presents an important subject for future work.

3 The climatic impacts described in section 4 are specific to geoengineering against a
4 baseline RCP8.5 scenario. If instead we had used a middle-of-the-road GHG-
5 concentrations scenario such as RCP4.5 [Taylor et al., 2012], as used in the first tier of
6 GeoMIP scenarios [Kravitz et al., 2011], then less aerosol-injection would be needed to
7 obtain TOA-Imb=0 and therefore the aerosol deposition rates and atmospheric mass
8 concentrations would be less than those reported in section 4. One would expect that the
9 magnitude of stratospheric temperature changes (Fig. 8) and therefore zonal-mean zonal
10 wind changes (Fig. 12) would be much less for each of the aerosols, possibly
11 confounding the conclusions giving here relating to their comparative efficacy. An
12 estimate for the amount of SAI required for RCP4.5 can be garnered from integrating the
13 temperature anomalies for RCP8.5 and RCP4.5 for the period 2020-2100. The ratio of the
14 integrated temperature anomalies for RCP4.5 to RCP8.5 is 0.43, hence we can assume
15 that the injection rates required for RCP4.5 are ~0.43 of those for RCP8.5, producing a
16 climate perturbation ~0.43 times as great. A further set of simulations, which instead
17 utilise RCP4.5 as the baseline scenario, would be required to test this hypothesis.

18 We have used prescribed ozone fields in these simulations because representing stratospheric
19 chemistry is prohibitively computationally expensive for the multiple centennial simulations
20 performed here [The HadGEM2 development team, 2011]. Kravitz et al (2012) showed in
21 their SmR scenario that BC injection could potentially result in global ozone depletion of
22 >50%, therefore the chemistry changes in SAI could potentially exceed the importance of the
23 physical changes in terms of climatic impacts (e.g. UV radiation at the surface). Tilmes et al
24 (2012) showed that SW-scattering by geoengineered sulfate could potentially compensate for
25 ozone-loss by back-scattering UV radiation in the tropics, but that this effect was
26 insufficiently compensatory at high latitudes. Their result was scenario-dependent; ozone loss
27 due to heterogeneous chemistry is enhanced for smaller particles and in the presence of higher
28 free-radical concentrations. Therefore, additional research is needed in order to understand the
29 effects on atmospheric chemistry of injecting alternative aerosols. This work has already
30 been started for titania by Tang et al (2014).

31 Another important aspect of SAI which is comparatively under-researched is the potential for
32 impacts on human health. Aerosol concentrations in the air near the surface are of interest

1 because of potential human respiratory impacts [Robock, 2008]. For instance, the USA's
2 National Institute for Occupational Safety and Health (NIOSH) recommends maximum
3 exposure limits of 0.3 mg m^{-3} for ultrafine titania particles (radius $<0.05 \text{ }\mu\text{m}$) and 2.4 mg
4 m^{-3} for fine particles (radius $< 1.5 \text{ }\mu\text{m}$) [Dankovic et al., 2011]. After undergoing
5 coagulation and ageing in the atmosphere, it is likely that the second exposure limit is
6 more applicable to this work. In our simulations, the maximum 2090's near-surface air
7 concentration of titania (e.g. Fig. 4) for land regions between 60°S - 60°N is 254 ng/m^3 , which
8 is of the order of 10^2 less than the NIOSH 'fine-particle' exposure limit. The equivalent
9 maximum concentration anomalies of BC in geoBC and SO_4 in geoSulf are 10 ng/m^3 and
10 1851 ng/m^3 respectively. More work is needed to assess the potential impacts of SAI on air
11 quality and human health.

12 Another thus far unmentioned aspect of this research is the potential for surface albedo
13 modification by aerosol deposition. In particular, BC deposition on snow reduces the snow
14 albedo through enhanced snow-melt and the coarsening of snow grains, which results in
15 amplified high-latitude warming [Marks and King, 2013]. HadGEM2-CCS does not include
16 the BC-on-snow feedback; therefore we estimate it by comparing the deposition rates for
17 2090s geoBC with the historical period. Jiao et al (2014) report that the simulated annual
18 mean Arctic ($>60^{\circ}\text{N}$) BC deposition for the 2006-2009 period ranges from $13\text{-}35 \times 10^7 \text{ kg/yr}$
19 for the AEROCOM Phase II models. The annual mean Arctic BC deposition for the 2006-
20 2009 period from our HadGEM2-CCS simulations is $23 \times 10^7 \text{ kg/yr}$, which is within the
21 AEROCOM range. The annual mean Arctic BC deposition anomaly for the 2090s period in
22 geoBC is $19.6 \times 10^7 \text{ kg/yr}$. Therefore, the effects of dirty snow in such an SAI scenario would
23 likely be significant, which would have impacts on the distribution of temperature,
24 particularly at high latitudes, potentially confounding some of our conclusions. Although we
25 have emphasized this issue with respect to BC, it is important to note that any particle that
26 absorbs SW radiation will instil this forcing. Therefore, titania, which has a non-unitary single
27 scattering albedo at short wavelengths, will also cause snow-grain coarsening and snow-melt
28 by absorbing solar radiation and warming the top layer of the snow pack.

29 This research has highlighted potential climate impacts of injecting various stratospheric
30 aerosols in order to ameliorate global warming. However, further research is needed to further
31 assess the climatic impacts of stratospheric aerosol injection such as the impacts on ozone.
32 Whilst research indicates that SAI is capable of averting certain climate changes such as

1 surface-warming, SAI provides no amelioration for other climate impacts, such as ocean
2 acidification. It is therefore important to note that the safest possible solution to avoiding the
3 sort of climate change instantiated by (e.g.) Fig. 6a of this report is to effectively mitigate
4 greenhouse-gas emissions.

5

6 **Author contribution**

7 ACJ designed the experiments, performed the simulations, analysed the data, and wrote the
8 manuscript with guidance and advice from JMH and AJ.

9

10 **Data sets**

11 Data used to generate figures, graphs, plots and tables are freely available via contacting the
12 lead author: aj247@exeter.ac.uk.

13

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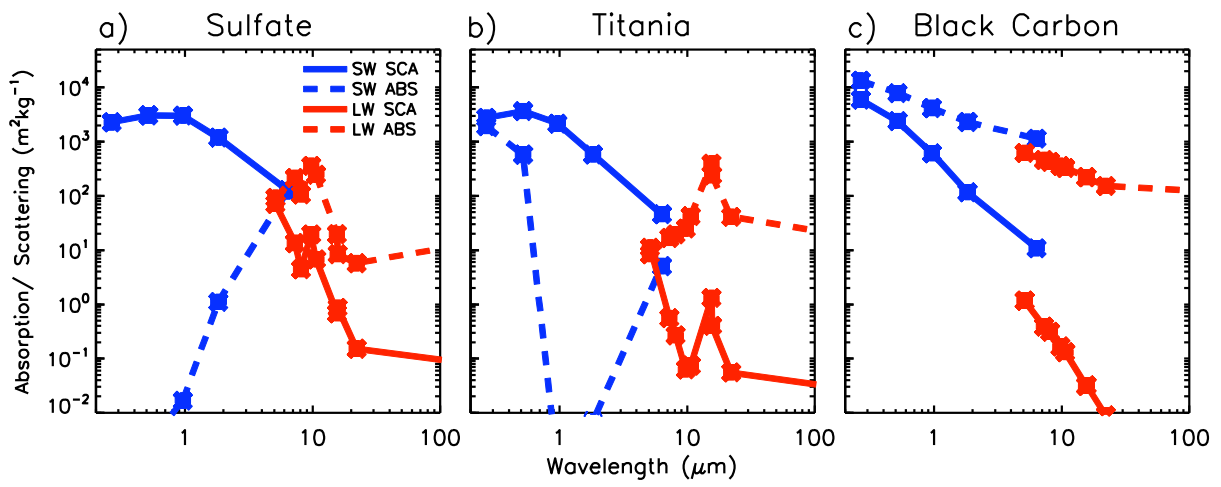
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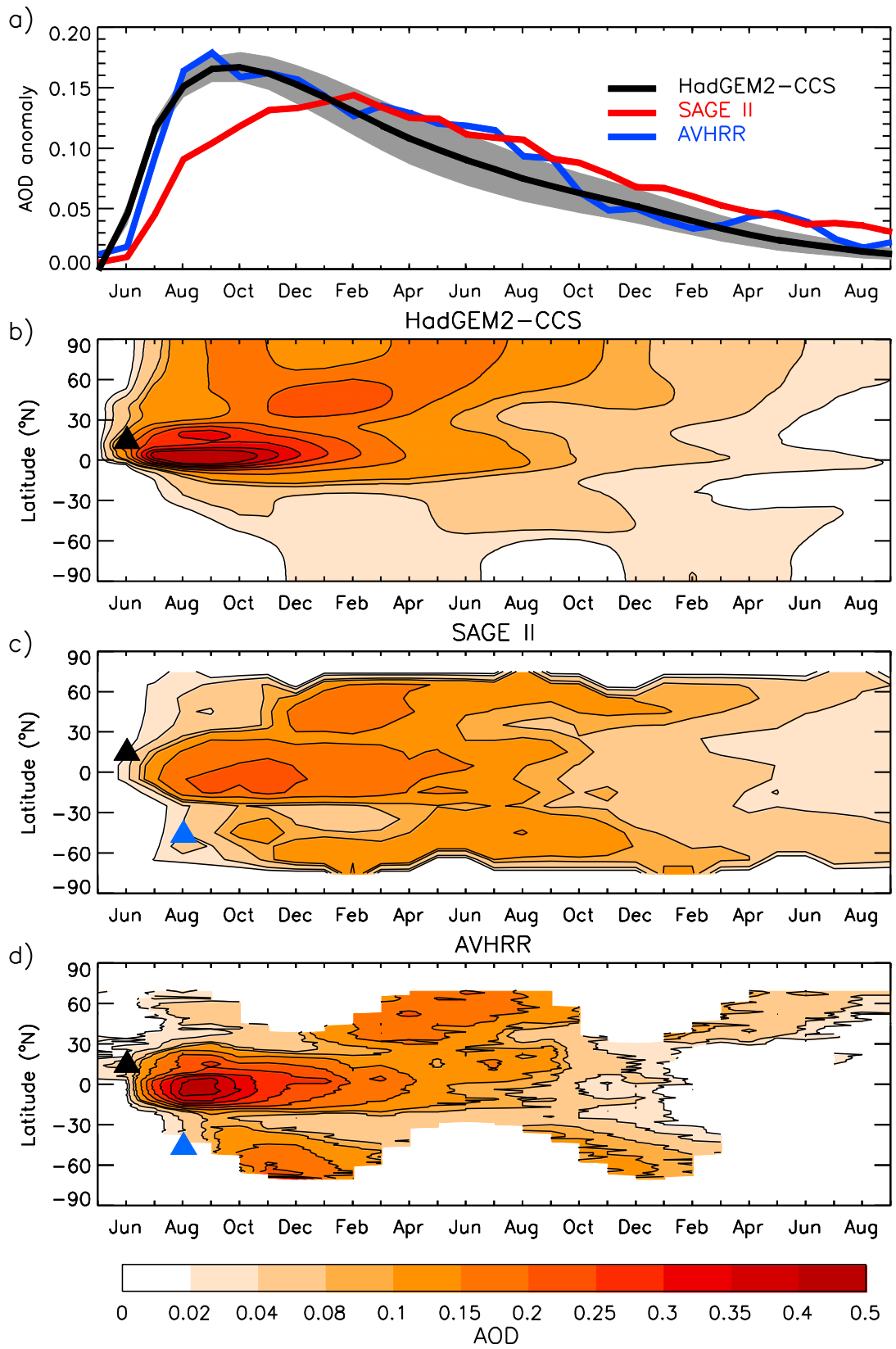
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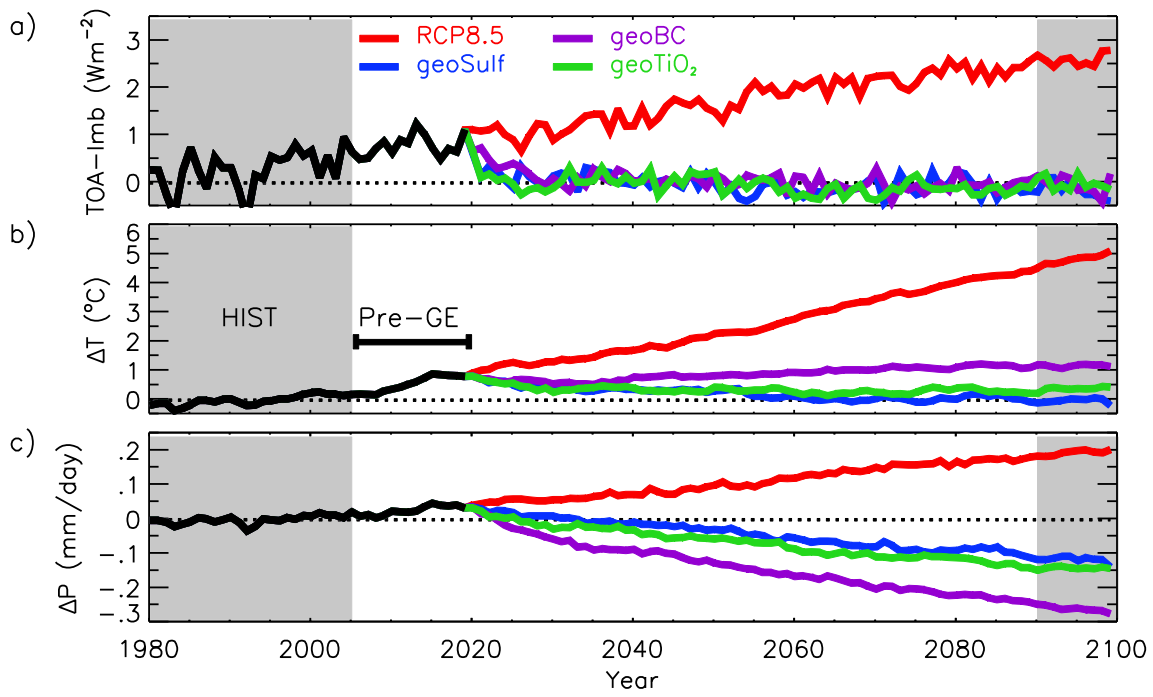
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Figure 1. *Optical properties as a function of wavelength for a) accumulation-mode sulfate, b) titania, c) black carbon. Points are plotted at the middle of each spectral waveband, as detailed in Bellouin et al (2007)*



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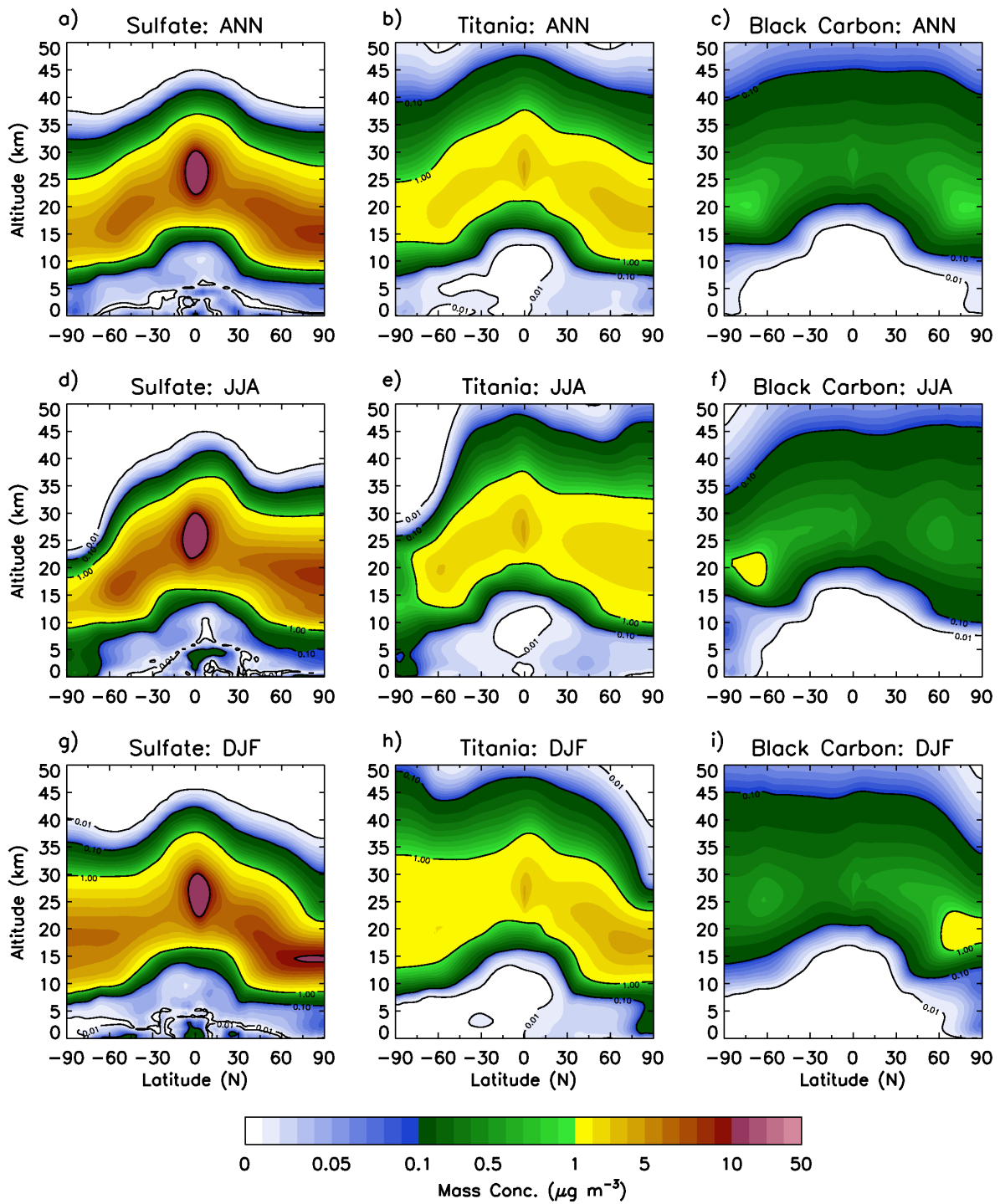
2 **Figure 2.** a) 75°S-75°N-mean 550nm sulfate AOD anomaly for the Pinatubo simulations and
 3 observations, b-d) timeseries of zonal-mean 550nm sulfate AOD anomaly



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2 **Figure 3.** Timeseries of annual/global-mean a) top-of-the-atmosphere radiative flux anomaly
 3 with respect to the pre-industrial control simulation b) near-surface air temperature anomaly
 4 with respect to the HIST period c) global mean precipitation anomaly with respect to HIST

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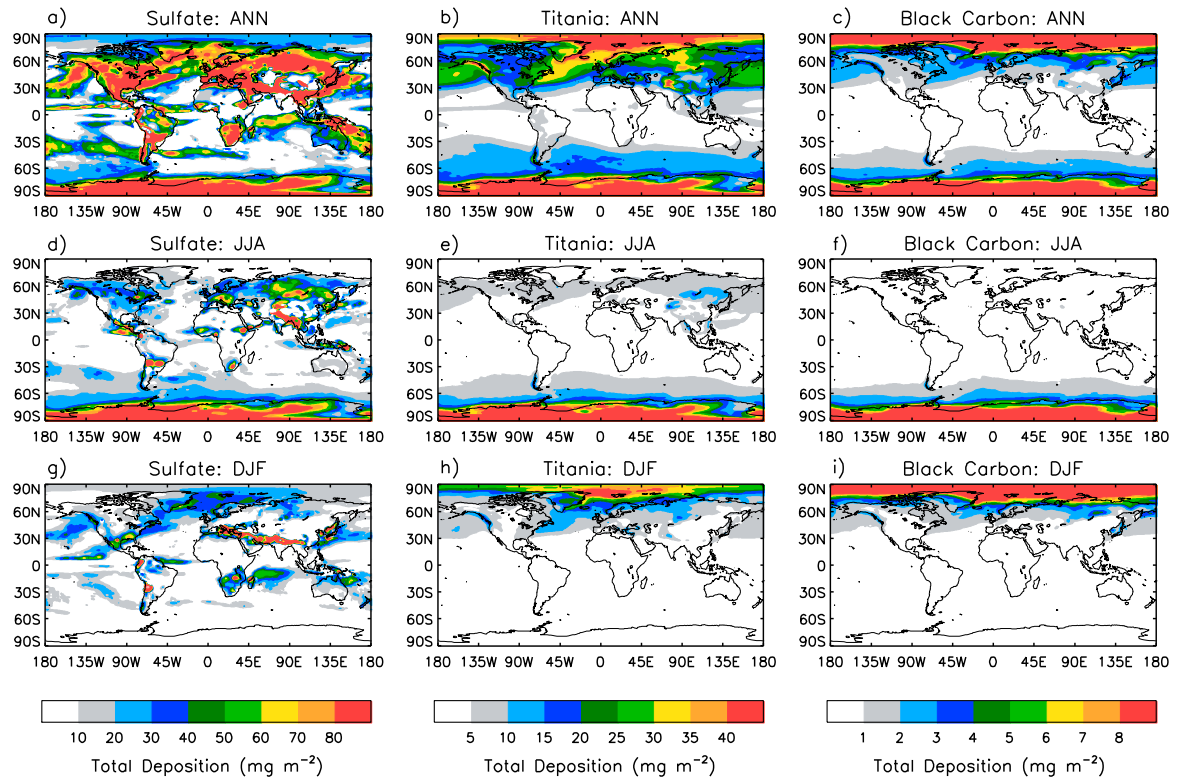


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2 **Figure 4.** Annual and seasonal zonal-mean mass concentration anomalies for sulfate (*geoSulf*

3 - left), titania (*geoTiO₂* - centre) and black carbon (*geoBC* - right)

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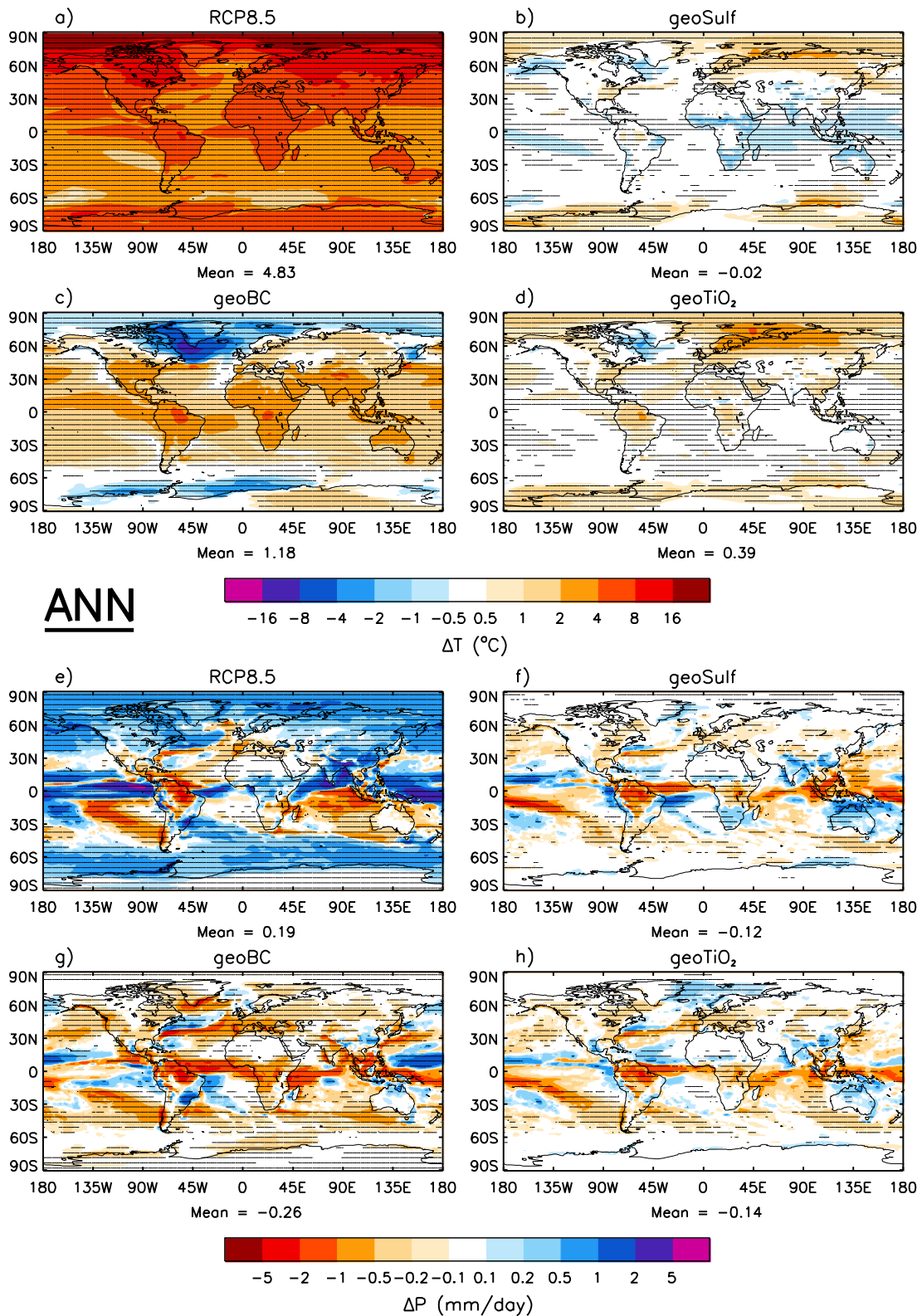


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2 **Figure 5.** Annual and seasonal total deposition anomalies (in units of $\text{mg m}^{-2} \text{yr}^{-1}$ and $0.25x$
 3 $\text{mg m}^{-2} \text{yr}^{-1}$ respectively)

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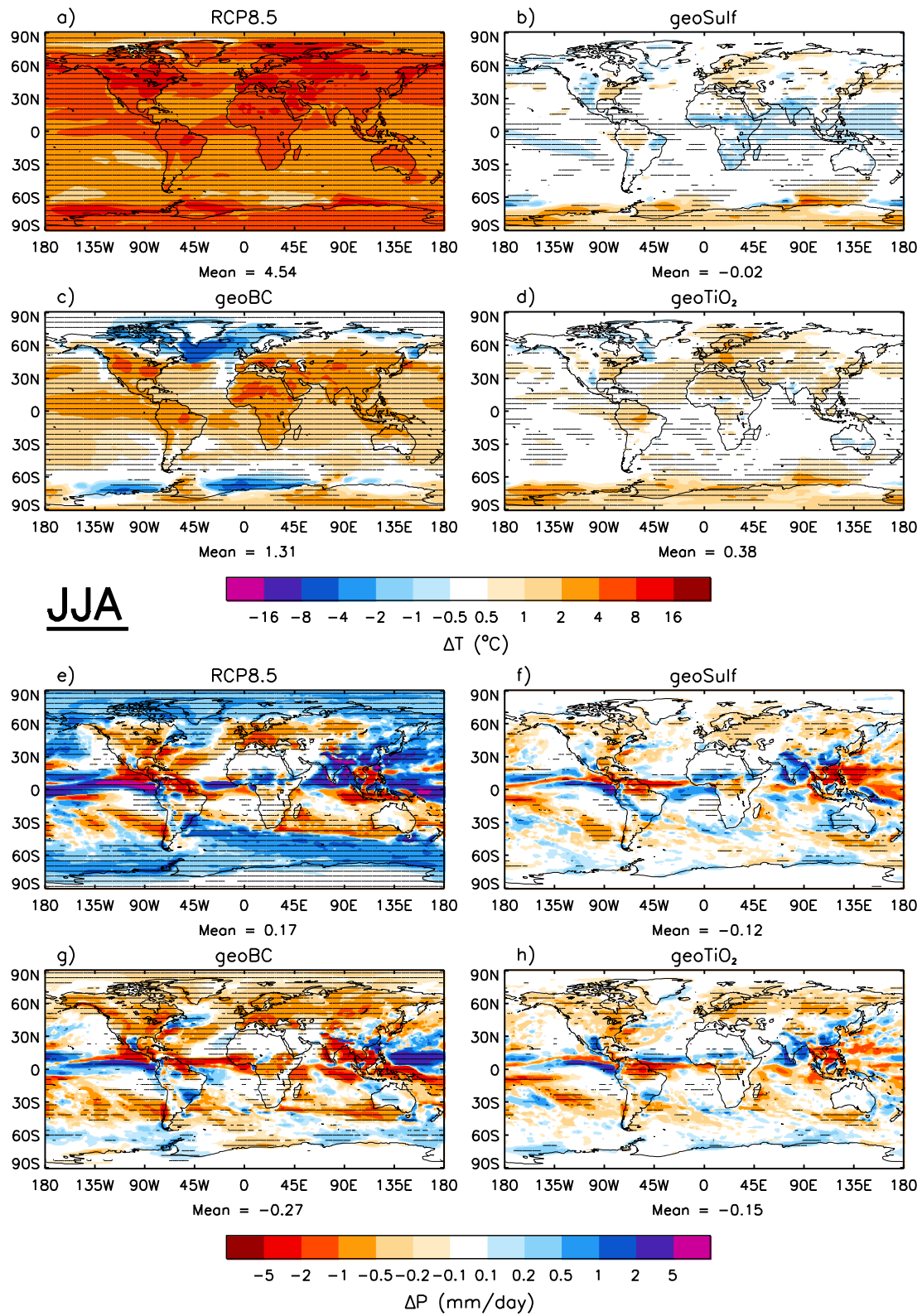
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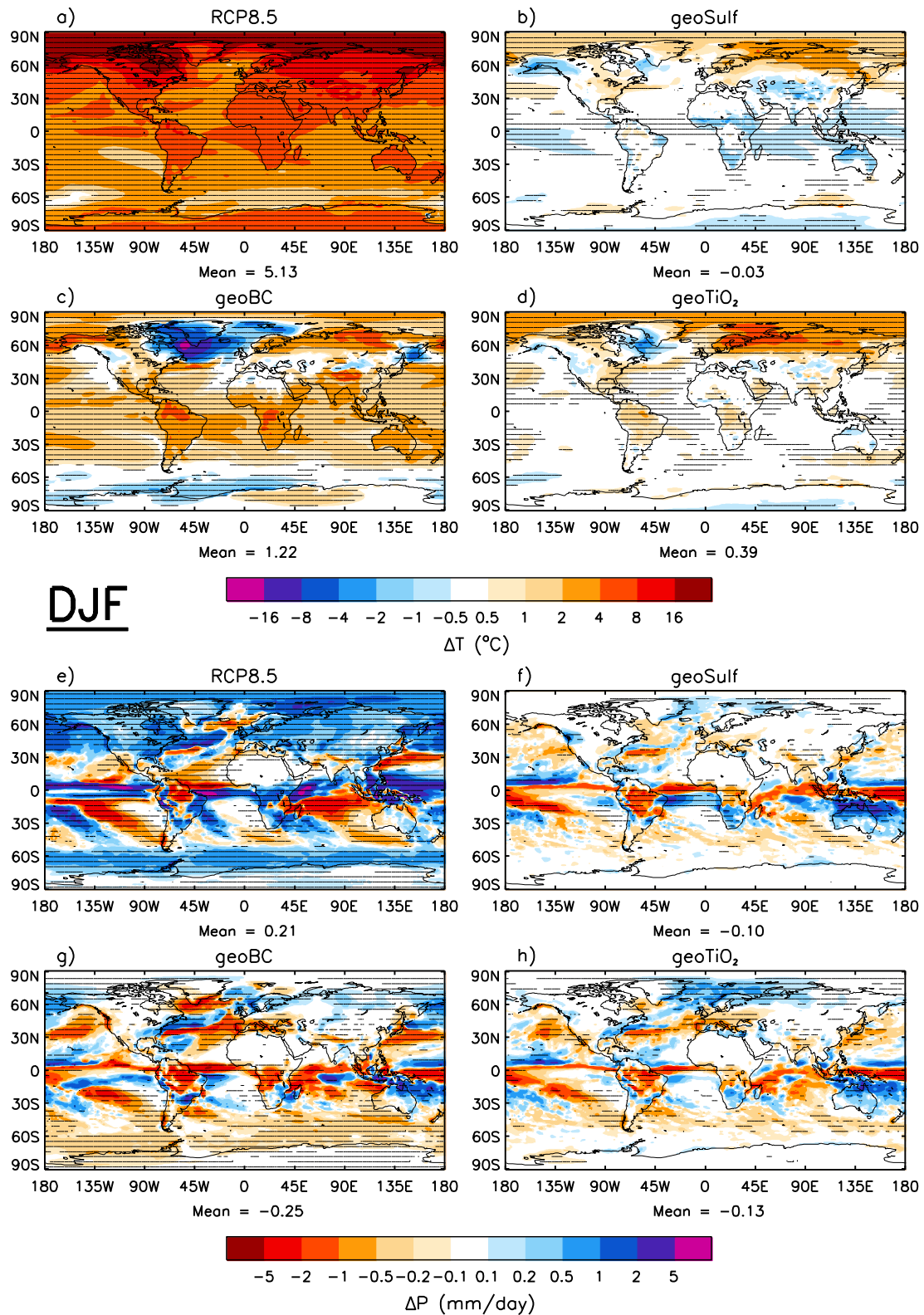
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Figure 6. Annual-mean near-surface air temperature (top) and precipitation rate (bottom) anomalies with respect to HIST. Stippling indicates where changes are significant at the 5% level using a two-tailed Student's *t*-test



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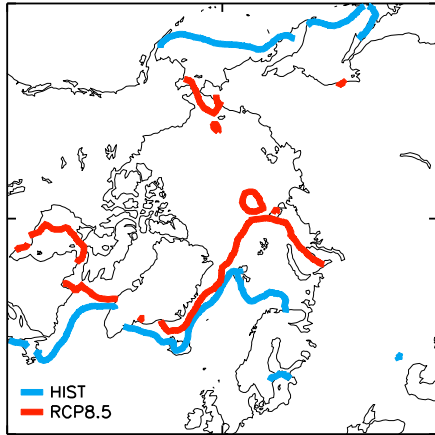
Figure 7. JJA near-surface air temperature (top) and precipitation rate (bottom) anomalies with respect to HIST



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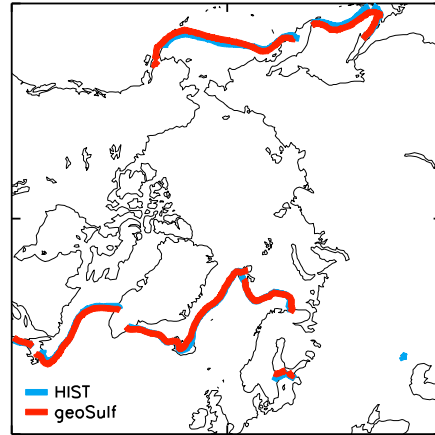
Figure 8. DJF near-surface air temperature (top) and precipitation rate (bottom) anomalies with respect to HIST

a) RCP8.5: Sea-Ice DJF



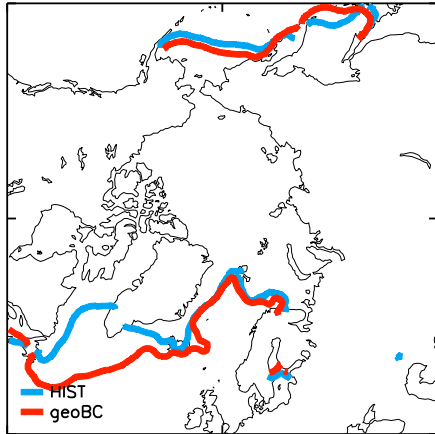
$$\Delta = -11.00 \text{ million km}^2$$

b) geoSulf: Sea-Ice DJF



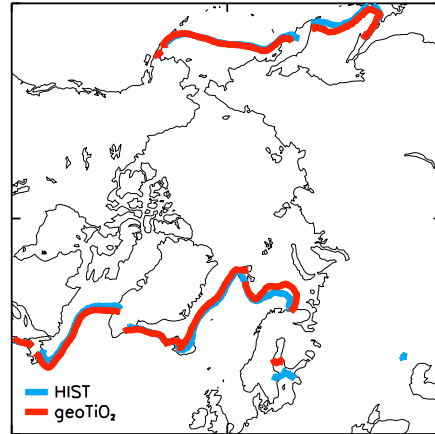
$$\Delta = -0.15 \text{ million km}^2$$

c) geoBC: Sea-Ice DJF



$$\Delta = +1.72 \text{ million km}^2$$

d) geoTiO₂: Sea-Ice DJF



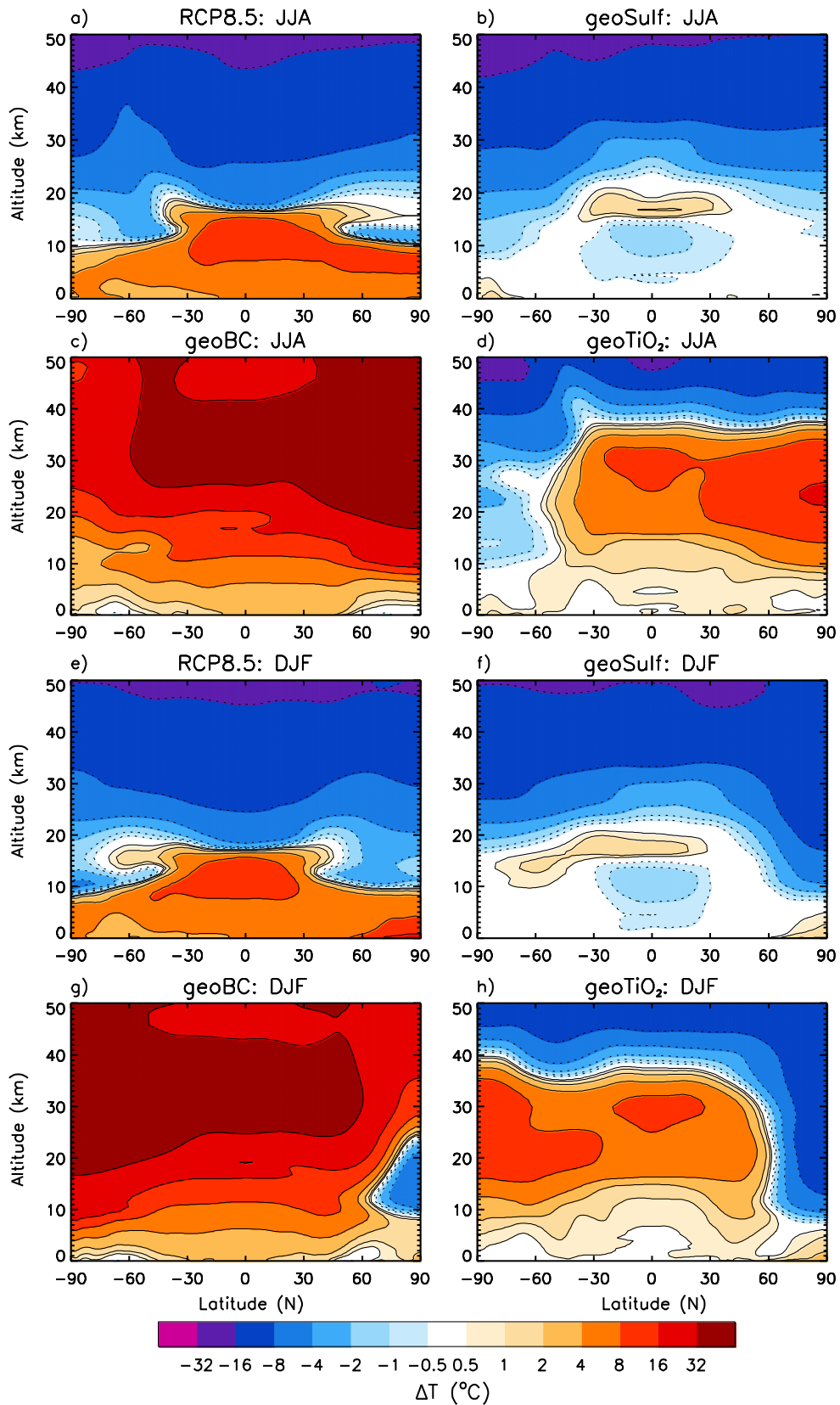
$$\Delta = -0.39 \text{ million km}^2$$

1

2

Figure 9. DJF northern-hemisphere sea-ice edge plotted with the HIST extent

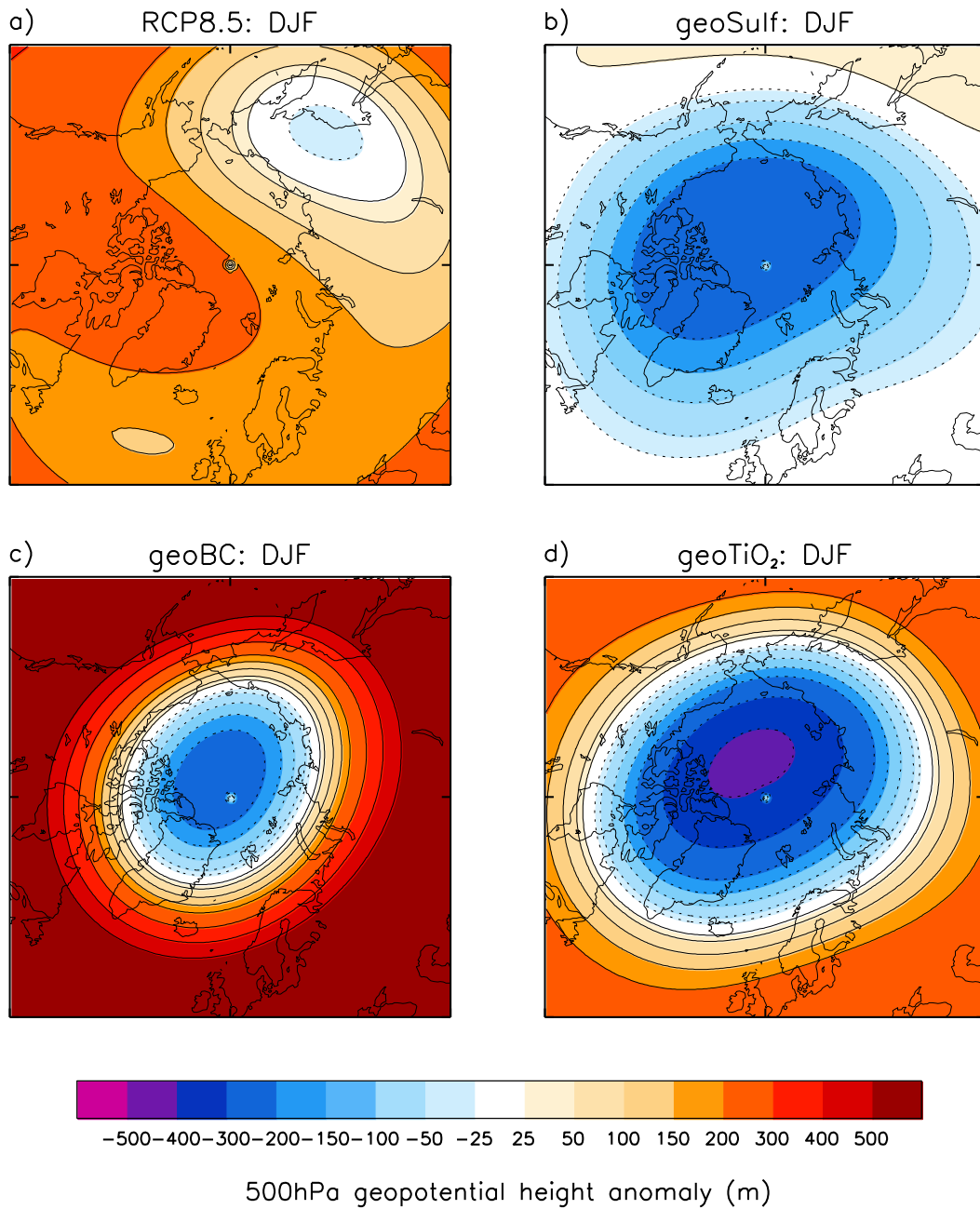
3



1

2 **Figure 10.** *JJA (top) and DJF (bottom) zonal-mean temperature anomaly with altitude, with*

3 *respect to HIST*



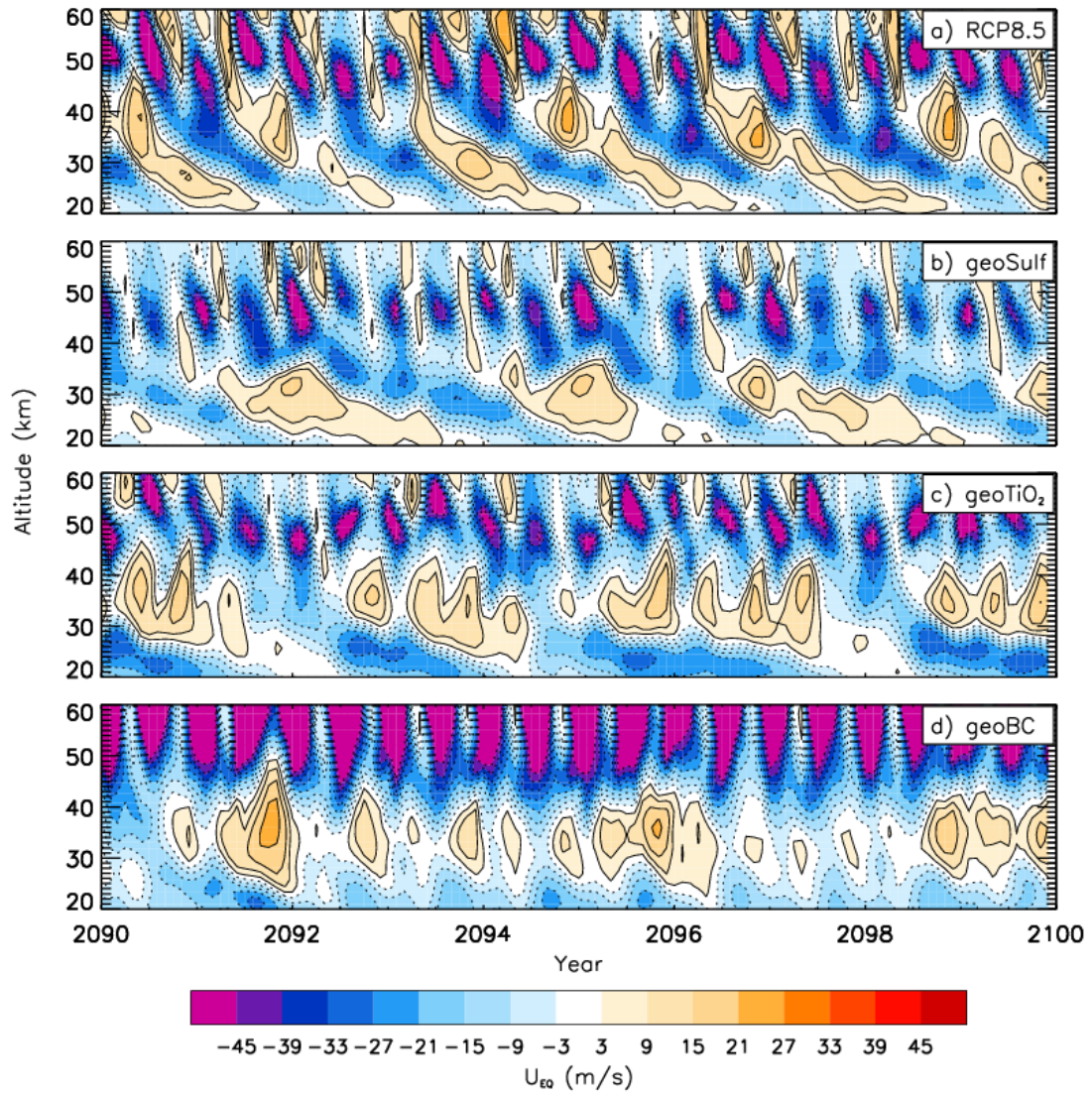
1

2

Figure 11. DJF 50hPa geopotential height anomaly

3

4



1

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Figure 12. Timeseries of equatorial (5°S-5°N) zonal-mean zonal wind profile

3