Climate of the Past Discussions

Does Antarctic glaciation cool the world?

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Abstract

In this study we compare the simulated climatic impact of adding the Antarctic Ice Sheet to the "Greenhouse World" of the Eocene and removing the Antarctic Ice Sheet from the Modern world. The Modern surface temperature anomaly (ΔT) induced by Antarctic Glaciation ranges from -1.22 to -0.18 K when CO_2 is dropped from 2240 to 560 ppm, whereas the Eocene ΔT is nearly constant at -0.3 K. We calculate the climate sensitivity parameter $S_{[Antarctica]}$ which is defined as the change in surface temperature (ΔT) divided by the change in radiative forcing ($\Delta Q_{Antarctica}$) imposed by prescribing the glacial properties of Antarctica. While the ΔT associated with the imposed Antarctic properties is relatively consistent across the Eocene cases, the radiative forcing is not. This leads to a wide range of $S_{[Antarctica]}$, with Eocene values systematically smaller than Modern.

This differing temperature response in Eocene and Modern is partially due to the smaller surface area of the imposed forcing over Antarctica in the Eocene and partially due to the presence of strong positive sea-ice feedbacks in the Modern. The system's response is further mediated by differing shortwave cloud feedbacks which are large and of opposite sign operating in Modern and Eocene configurations. A negative cloud feedback warms much of the Earth's surface as a large ice sheet is introduced in Antarctica in the Eocene, whereas in the Modern this cloud feedback is positive and acts to enhance cooling introduced by adding an ice sheet. Because of the importance of cloud feedbacks in determining the final temperature sensitivity of the Antarctic Ice Sheet our results are likely to be model dependent. Nevertheless, these model results show that the radiative forcing and feedbacks induced by the Antarctic Ice Sheet did not significantly decrease global mean surface temperature across the Eocene-Oligocene transition (EOT) and that other factors like declining atmospheric CO_2 are more important for cooling across the EOT. The results indicate that climate transitions associated with glaciation depend on the climate background state. This means that using paleoclimate proxy data by itself, from the EOT to estimate Earth System Sensitivity, into

the future, is made difficult without relying on climate models and consequently these modelling estimates will have large uncertainty, largely due to low clouds.

1 Antarctic Ice Sheet temperature sensitivity

During the Eocene-Oligocene Transition (EOT) global climate deteriorated as the warm
5 and relatively ice free conditions of the Eocene gave way to a colder, glaciated state in the early Oligocene (Lear et al., 2000; Zachos et al., 2001; DeConto and Pollard, 2003; Macksensen and Ehrmann, 1992; Scher et al., 2011; Hambrey and Barrett, 1993). In contrast, the Modern Earth System is currently in a glaciated state, but is showing signs of potentially losing glacier ice in the Arctic and Antarctic (Joughin and Alley, 2011;
10 Jacob et al., 2012; Velicogna, 2009; Chen et al., 2009; Pritchard et al., 2009; Liston and Hiemstra, 2011). Evidence now exists that the cooling (Liu et al., 2009; Eldrett et al., 2009; Zanazzi et al., 2007; Ivany et al., 2000) and glaciation (Lear et al., 2000; Edgar et al., 2007; Coxall et al., 2005; Miller et al., 2009; DeConto and Pollard, 2003; Zachos et al., 2001) that occurred across the EOT was caused by a drop in CO_2 mixing
15 ratios from ∼1000 ppm to ∼600 ppm (Pagani et al., 2011; Pearson et al., 2009) – the likely range of values over the next century. One major, unanswered question in future climate change prediction is the degree to which melting of ice sheets in the future will substantially and irreversibly alter climate (Solomon et al., 2009). Past climate changes, such as the EOT may provide unique information to answer that question.
20 Indeed, with both greenhouse gas forcing and temperature change values in hand from EOT proxy records there is a significant temptation to estimate an Earth System Sensitivity (ESS) parameter (Lunt et al., 2010), i.e. a climate sensitivity parameter that includes the direct, fast feedback responses to radiative perturbation combined with the slower feedbacks, such as ice sheet growth, greenhouse gas and vegetation feedbacks
25 (Palaeosens members, 2012; Royer et al., 2012). One approach is to use this Eocene ESS to draw a straightforward analogy to the future (Hansen et al., 1984, 2010; Hansen and Sato, 2012; Kiehl, 2011; Hay, 2011) thus avoiding the messy details of accurately

predicting the individual processes and feedbacks that plague most modelling efforts (Roe and Baker, 2007). This type of ESS estimate could hold great promise for predicting the future long-term climate evolution if the approach is valid. Reasons for skepticism exist, not the least of which being the importance of hysteresis in ice sheet evolution (Pollard and DeConto, 2005).

We are motivated in this paper by other concerns. On long timescales does a glaciated Antarctica cool global mean temperature? Does the Antarctic Ice Sheet induce additional positive or negative climate feedbacks and if so what are the strength of these feedbacks? Are the results of this change state dependent?

Recent estimates indicate that the CO_2 levels over the EOT fell from 1000 to 600 ppm (Pearson et al., 2009; Pagani et al., 2011), directly causing a 2.1–2.5 $W m^{-2}$ radiative forcing (Myrhe et al., 1998). To reconcile the temperature shift at the EOT of $\sim 3.5 \pm 1.5$ K (Liu et al., 2009) – assuming that this shift was entirely due to the fast feedbacks – this would require a Charney-type temperature sensitivity of ~ 1.5 K $(W m^{-2})^{-1}$. This is roughly double the typical estimated Modern value of 0.8 K $(W m^{-2})^{-1}$ (Bitz et al., 2012; Kay et al., 2012a; Gettelman et al., 2012). Thus, while the shift in CO_2 values over the EOT is more or less well established as the prime candidate for driving the cooling, this implies either a surprisingly large value of fast sensitivity or substantial slow, Earth System positive feedbacks that enhance the sensitivity. It is currently unknown, and indeed impossible to know directly from proxy data, what fraction of the cooling at the EOT was a direct climate response involving the fast, Charney type feedbacks like shifts in clouds and sea ice (Hansen et al., 1981, 1997; DeConto et al., 2007) and what fraction of the cooling involved the slower feedbacks like changes in the Antarctic Ice Sheet (Lunt et al., 2010) given that proxy records for the radiative forcing due to ice sheets do not exist. Attempts to estimate the radiative forcing due to clouds and ice sheets indirectly by inferring ice sheet volume changes are a reasonable first attempt, but as we show below, there is no a priori reason to assume a close relationship between the direct surface radiative forcing over the ice sheet and the global mean top of atmosphere (TOA) forcing over the ice sheet after the systems fast feedbacks have

operated. For example, adding or removing an ice sheet in a region with a thick, low cloud deck will have a completely different forcing on climate than the same changes made in a cloud-free region. Furthermore, local cloud shortwave feedbacks will determine the change in radiative forcing of the ice sheet alterations and those local cloud

5 shortwave feedbacks are unconstrained. Consequently, the TOA radiative perturbation associated with adding or removing the Antarctic Ice Sheet is likely to both state dependent and model dependent. This is a less straightforward problem than determining the forcing due to a doubling of CO_2.

We are focused in this paper on understanding what slow, Earth system feedbacks

10 were operating across the EOT and their interactions with fast feedbacks, to help evaluate whether these feedbacks operate in the same way (in the model) in the Modern (Haywood et al., 2011). Specifically, we use the National Center for Atmospheric Research (NCAR) Community Earth System Model (CESM1.0) in slab ocean mode to investigate the impact of replacing the Antarctic Ice Sheet with vegetation for the future

15 case, and replacing vegetation with an ice sheet for the EOT cases. We ask the following questions: what is the climatic impact and its sensitivity to adding or removing a large Antarctic Ice Sheet? Does this response depend on the climate state, is the response in the Eocene different than in the Modern? What feedbacks are important for modulating this response?

20 The remainder of paper will be focused on explaining and quantifying temperature change, radiative forcing perturbation, and the resulting climate sensitivity parameter induced by removing and adding the Antarctic Ice Sheet and comparing this response in Eocene and Modern contexts. This paper is broken into four sections. Section 2 describes the CESM1.0 modelling framework and how we constructed our Eocene and

25 Modern glaciated and unglaciated simulations. Then we present the climate sensitivity to Antarctic glaciation in Modern and Eocene slab ocean simulations (Sect. 3.1) and describe the atmospheric response to Antarctic glaciation in the Eocene and Modern slab ocean simulations (Sect. 3.2). Sections 4 and 5 include the discussion and conclusion, respectively.

2 Methods

2.1 CESM1.0 modelling framework

We perform a series of slab ocean global climate model simulations using the NCAR CESM1.0 as described in Neale et al. (2010), Gent et al. (2012), and Bitz et al. (2011).

5 The CESM1.0 configuration includes the Community Atmosphere Model (CAM4), the Community Land Model (CLM4) (Lawrence et al., 2012), and the Community Sea-Ice Model (CICE4) (Hunke and Lipscomb, 2008; Brady et al., 2012) coupled to a slab ocean.

CAM4 employs the revised Zhang and McFarlane parameterized deep convection

10 scheme and finite dynamical core (Lin, 2004; Gent et al., 2012; Mishra et al., 2011; Zhang and McFarlane, 1995). We use the $2° \times 2.5°$ finite volume core because it is able to adequately resolve some of the finer scales important for atmospheric hydrology and energy conservation and this configuration has a reduction in numerical dispersion in comparison to CAM3 spectral core (Neale et al., 2010), which we have used for past

15 paleoclimate applications (Huber and Caballero, 2011). CAM4 has an improved calculation of freeze drying which reduces biases in the low cloud properties and the radiative budget in the high latitudes compared to CAM3 (Neale et al., 2012; Vavrus and Waliser, 2008). These improvements lead to improved high latitude temperature seasonality in Modern simulations between CAM3 and CAM4 (Bitz et al., 2011). When

20 CAM4 cloud distributions are compared against the international satellite cloud climatology project (ISCCP) and CALIPSO data the model is able to spatially match cloud observations in the tropics and extra-tropics (Kay et al., 2012b), but CAM4 underrepresents the total cloud fraction in these regions (Kay et al., 2012b), especially in the Arctic (Boer et al., 2012).

25 Until recently, deep time paleoclimate simulations have used prescribed aerosol datasets based on pre-industrial values or have set the global aerosol concentrations to 0. Here, we create prescribed aerosol forcing files specifically for the late Eocene. Building these files requires a two step process. First, we run CAM4.0 in bulk aerosol

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mode (BAM) (Tie et al., 2005) with late Eocene boundary conditions. The CAM4 BAM configuration allows for the aerosol variables like sea-salt, dust, SO_4, SO_2, to be solved prognostically (Seland et al., 2008; Kirkeva et al., 2008) within a late Eocene climate simulation. The equilibrated CAM4 BAM model output is then used to create prescribed

5   aerosol forcing files for input CAM4 Eocene simulations. The prescribed aerosol files should improve the realism and self consistency of the cloud forcing response in the Eocene simulations because aerosol concentrations and spatial coverage are derived from Eocene boundary conditions. Another, improvement of CESM1.0 over prior modelling efforts is that the ice model (CICE4) includes a new scattering parametrization

10 scheme (Brieglob and Light, 2007) which should increase the realism of snow albedo and shortwave forcing effects (Gent et al., 2011).

The CESM1.0 slab configuration has fully interactive sea ice unlike the previous version of CCSM3.0 which had purely thermodynamic sea ice (Kay et al., 2011, 2012a). The slab configuration incorporates heat convergence, mixed layer depths, and salinity

15 from existing NCAR Community Climate Model version 3 (CCSM3.0) fully coupled simulations. A series of previous CCSM3.0 fully coupled Eocene simulations were used to create the slab ocean data sets. These CCSM3 simulations were integrated over 3000 model years and run at (560, 1120, 2240 ppm CO_2). Details can be found in Liu et al. (2009), Ali and Huber (2010), Huber and Caballero (2011), and Huber and Goldner

20 (2011). The final 40 yr of ocean heat convergence, salinity, temperature, and ocean currents from the fully coupled Eocene simulations are used as climatologies to create the CESM1.0 slab ocean forcing file. In a series of simulations from Eocene through Miocene and using a variety of models we have shown that ocean heat transport is relatively stable (Huber and Sloan, 2001; Huber et al., 2004; Sijp et al., 2011; Herold et al.,

25 2012) and not the first order control on Antarctic surface conditions (Huber and Nof, 2006). Additionally, preliminary CESM1.0, fully coupled simulations show that there are no appreciable ocean circulation differences between the models so we are confident that utilizing CCSM3 ocean fields is not a concern. The slab ocean configuration allows us to run many sensitivity studies to equilibrium. In this paper will present only a small

subset of sensitivity studies conducted, and the results focus on the main features revealed from all simulations.

To set the stage for describing the methodology for the glacier simulations we present Fig. 1, which is a schematic of the simulations that were completed to explore the climate impacts to changes in the Antarctic Ice Sheet. Because shifts in the Antarctic Ice Sheet include topographical changes and albedo changes we show the possible simulations using a three dimensional schematic (Fig. 1).

2.2 Antarctic sensitivity study methods

To investigate the Modern Antarctic glacier sensitivity we take the default Modern Antarctica topography dataset (Fig. 2a) and decrease its height uniformly by 80 %. This lowers the Antarctic topography to 500–1000 m (Fig. 2b) and is a gross estimation for what the unglaciated Modern world would look like without the Modern Antarctic height and albedo and after allowing for glaciostatic rebound.

The unglaciated low topography used in the Eocene simulation is plotted in Fig. 2c (Sewall et al., 2000). To create the glaciated Eocene simulations we introduce a large Modern ice sheet over the Antarctic continent, increasing the mean height to 3000–4000 m (Fig. 2d). More advanced approaches, for example using the new ANTSCAPE Antarctic paleotopography (Wilson et al., 2011) or using ice sheet tographies from De-Conto et al. (2007) would enhance the realism of our study, but at the expense of adding complexity, we prefer simplicity for this set of experiments. Preliminary results with more realistic Antarctic topographies indicate that our main results are not changed by this simplification.

We separately and in conjunction study the sensitivity as we remove the glacier albedo and replace the land surface type with broadleaf boreal forest. The surface albedo anomaly between the glaciated and unglaciated simulation after fast feedbacks (snow) have operated is shown in Fig. 3. In Fig. 3, there is roughly a 60 % drop in albedo locally in Antarctica when the topography is lowered and replaced with broadleaf boreal forest during the austral summer (December, January, February). We run the glaciated

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versus unglaciated simulations at equivalent CO_2 levels, but test Antarctic sensitivity to CO_2 by varying CO_2 levels over a wide range (560, 1120, and 2240 ppm).

In summary, the height of Antarctica is identical in the glaciated Eocene and Modern simulations, but the area of Antarctica in the Eocene is roughly 30 % smaller than Modern because the landmask is based off of Sewall et al. (2000), and we did not alter the landmask in Eocene simulations. Numerous research groups have attempted to estimate Antarctic ice volume growth at the EOT and some studies have suggested that the volume of the Antarctic Ice Sheet was smaller or near Modern size (Miller et al., 1987, 2009; Edgar et al., 2007; Bohaty et al., 2012). While other studies suggest that the ice volume in Antarctica had to be as large or larger than Modern because Mg/Ca paleotemperature records do not show deep ocean cooling, implying that the shift in $\delta^{18}O$ is too large to be accounted by Antarctica Ice Sheet volume alone (Coxall et al., 2005; Lear et al., 2000; Katz et al., 2008; Pusz et al., 2011). Recently Liu et al. (2009) and Deconto et al. (2008) estimated EOT ice volume to be between 40 % to 120 % of Modern by accounting for the +1.5 % shift in benthic $\delta^{18}O$ and a deep ocean cooling of 3 to 5 K. Thus the size of our Antarctic Ice Sheet is well within the range of possible given constraints imposed by proxy data. Estimates for Antarctic glacial extent during the EOT is still uncertain so our approach is simply one of many possible approaches and this work should be considered as an exploratory sensitivity study.

2.3 Radiation diagnostics and climate sensitivity parameter calculations

We calculate the globally averaged temperature change (ΔT) by comparing two cases and varying one or more parameters. In the results below we will refer to the term ΔT Antarctica which is ΔT over the Antarctic region of 60° S to 90° S. For cases in which the Antarctic Ice Sheet has been changed, we denote ΔT with the subscript (α) for changing albedo and (oro) for height of Antarctica, so we distinguish between simulations in which only the surface properties are changed $\Delta T_{(\alpha)}$ and those in which both the height and surface properties are changed $\Delta T_{(\alpha+\text{oro})}$. We treat each separately because this will help elucidate the importance of changing the height of the Antarctic

Ice Sheet versus changing the albedo of the ice sheet. Additionally, in some simulations we change (α, oro, and CO_2) and these cases are defined as $\Delta T_{(\alpha + \text{oro} + CO_2)}$.

Quantifying the forcing of the Antarctic Ice Sheet is not straightforward and here we investigate a couple different approaches for the calculation. First, prior work estimated a forcing as being directly related to only the change in surface albedo and the inferred change in the surface energy budget (Hansen et al., 1997; Rohling et al., 2012; Myhre and Myhre, 2003; Myhre et al., 1998), while ignoring or applying a correction to account for all the other possible radiative and dynamical feedbacks. But, it is arguable that the forcing that is relevant for climate is the TOA radiation budget change over the region in which the ice sheet is being altered. This requires including some local, fast feedback terms within the definition of the forcing, but we argue that this is both customary in more commonly studied situations such as changing CO_2 and necessary to capture the true magnitude of the forcing. In the first instance, it is common to calculate the forcing due to CO_2 changes after allowing for a brief stratospheric adjustment. In the second instance, we note that while the directly imposed albedo forcing is dictated by specified albedo differences between glacier and forest in CLM4, the radiative forcing experienced will be mediated immediately by fast feedbacks locally such as snowfall.

To address these issues we will calculate the change in the energy budget at the surface and TOA to explore if the surface induced change is representative of what occurs at the TOA. Since the forcing of the Antarctic Ice Sheet is occurring over a specified region, unlike CO_2, which is a globally distributed forcing, we will explicitly calculate the forcing of the Antarctic Ice Sheet by quantifying the change in surface and TOA short-wave fluxes over just the Antarctica landmask region and the equations describing this calculation will be described in detail below. We take the approach here that inclusion of feedbacks that are local to Antarctica and act immediately (i.e. within weeks) such as snow albedo changes over Antarctica should be folded into forcing for the longer time scales of interest to us in this study. This approach has been used in other studies (Cess et al., 1989; Lunt et al., 2010). This loosening of the definition of forcing can be extended to include local, fast cloud feedbacks over Antarctica by calculating

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the change in TOA energy budgets which makes the forcing more analogous to that of CO_2. Additionally, to understand the global feedback response to Antarctic glaciation we will also do a global calculation exploring the change in the shortwave energy budget at the surface and the TOA.

5   The global calculation is defined as ΔFSNT and we will calculate this value by taking the global mean change in net shortwave radiation at the TOA between glaciated cases ($FSNT_{glaciated}$) and unglaciated cases ($FSNT_{unglaciated}$). The following equations, in which FSNT represents global mean net TOA shortwave fluxes and the subscripts refer to the cases in question. The calculations are performed both for simulations where

10   only (α) was changed (Eq. 1), in which ($\alpha + oro$) was changed (Eq. 2), and simulations in which ($\alpha + oro + CO_2$) are changed (Eq. 3). Similar calculations were performed using the net surface shortwave flux change (FSNS) (equations not shown) instead of FSNT.

$$\Delta FSNT_{(\alpha)} = FSNT_{glaciated_{(\alpha)}} - FSNT_{unglaciated_{(\alpha)}} \tag{1}$$

15

$$\Delta FSNT_{(\alpha+oro)} = FSNT_{glaciated_{(\alpha+oro)}} - FSNT_{unglaciated_{(\alpha+oro)}} \tag{2}$$

$$\Delta FSNT_{(\alpha+oro+CO_2)} = FSNT_{glaciated_{(\alpha+oro+CO_2)}} - FSNT_{unglaciated_{(\alpha+oro+CO_2)}} \tag{3}$$

To explicitly evaluate the forcing of the Antarctic Ice Sheet we calculate the globally

20   weighted forcing over just the Antarctica region (60° S to 90° S) to get a value of the forcing without the inclusion of the global feedbacks. We take a weighted sum of ΔFSNT over the model grid cells that include the land grid cells within the Antarctic landmask area ($\Delta FSNT_{landmask}$) and scale this value by the ratio (SL), which is the cells associated with the Antarctic landmask (m^2) divided by the area of globe (m^2) (Eq. 4). The

25   globally weighted forcing over the Antarctica region is $\Delta Q_{Antarctica}$ (Eq. 5).

$$SL = \left(\frac{\text{forcing area}}{\text{area globe}} \right) \qquad (4)$$

$$\Delta Q_{\text{Antarctica}} = \Delta\text{FSNT}_{\text{landmask}} \cdot SL \qquad (5)$$

The calculations for $\Delta Q_{\text{Antarctica}}$ are performed using FSNT, FSNS, and clearsky net
shortwave flux at the surface (FSNSC) and these derived Antarctic Ice Sheet forcings
are summarized in Table 1.

Table 1 also includes the Eocene and Modern glaciated versus unglaciated simu-
lations exploring ESS (Eq. 7), and S the climate sensitivity parameter. S measured
in $\text{K}(\text{Wm}^{-2})^{-1}$ is defined as the change in surface temperature (ΔT) divided by the
change in radiative forcing of the Antarctic Ice Sheet (Eq. 8). As an important reference
point, ΔQ_{CO_2}, the change in radiative forcing due to a doubling of atmospheric carbon
dioxide from 280 to 560 ppm in CAM4.0 simulations (Bitz et al., 2011; Gettelman et al.,
2012) is approximately $3.5\,\text{Wm}^{-2}$ (Eq. 6) which is close to the standard value used in
previous work (Myhre et al., 1998). We note that this value for ΔQ_{CO_2} is an approxima-
tion, and is model dependent (Bitz et al., 2011) and not constant at higher CO_2 levels
(Senior and Mitchell, 2000; Boer and Yu, 2003).

$$\Delta Q_{\text{CO}_2} = 3.5\,\text{Wm}^{-2} \qquad (6)$$

$$\text{ESS} = \frac{\Delta T_{(\alpha+\text{oro}+\text{CO}_2)}}{\Delta Q_{\text{CO}_2}} \qquad (7)$$

$$S_{[\text{Antarctica}]} = \frac{\Delta T_{(\alpha+\text{oro})}}{\Delta Q_{\text{Antarctica}}} \qquad (8)$$

$$S_{[\text{Antarctica, CO}_2]} = \frac{\Delta T_{(\alpha+\text{oro}+\text{CO}_2)}}{\Delta Q_{\text{CO}_2} + \Delta Q_{\text{Antarctica}}} \qquad (9)$$
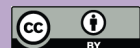
Here we will calculate $\Delta Q_{\text{Antarctica}}$ using FSNT, FSNS, and also FSNSC allowing for a comparison between different Antarctic forcing values. First we calculate $S_{[\text{Antarctica}]}$ (Eq. 8), by prescribing the glacial properties of Antarctica at constant atmospheric CO_2. Second, we calculate $S_{[\text{Antarctica, } CO_2]}$ (Eq. 9) by reducing the atmospheric CO_2 from
5 1120 ppm to 560 ppm and removing the Antarctic Ice Sheet. In what follows, we will refer the reader to Table 1, which describes the different experiments for the Eocene and Modern presented in the results section including all values for ESS and S.

3   Results

3.1   Sensitivity to Antarctica Ice Sheet in Modern and Eocene

10 In general, the Modern glaciated simulations have larger global mean ΔT than the Eocene glaciated simulations (Table 1). The Modern Antarctic glacier experiment in which only albedo is changed (Table 1, α cases), has a $\Delta T_{(\alpha)} = -1.14$ to -0.86 K, while the corresponding Eocene experiment (Table 1, α cases) have $\Delta T_{(\alpha)} = -0.36$ K to -0.27. When considering the sensitivity to both components of ice sheet growth, which
15 we have done at a range of CO_2 values, we find that the Eocene has a $\Delta T_{(\alpha+\text{oro})} = -0.16$ to -0.30 while the Modern has a $\Delta T_{(\alpha+\text{oro})} = -0.23$ to -1.25 K (Table 1, α + oro cases).

   The results mentioned above can be summarized clearly in Fig. 4, where we plot the mean annual temperature (MAT) of the Eocene and Modern unglaciated
20 and glaciated simulations across a range of atmospheric CO_2 levels. Comparing the Eocene unglaciated and glaciated simulations at constant atmospheric CO_2 levels results in little global temperature change (Fig. 4). Comparing the Modern glaciated and unglaciated simulations at 560 and 1120 ppm CO_2, results in a smaller MAT change compared to the MAT change that occurs at 2240 ppm CO_2. This is because at lower
25 atmospheric CO_2 (560 and 1120) when we remove the Antarctic Ice Sheet in the Modern, our imposed albedo change is offset by increased snowfall over Central Antarctica.

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The increased snowfall occurs because of elevated moisture transport into Antarctica because the Katabatic winds are reduced as the elevation over Antarctic is decreased resulting in onshore flow (figure not shown). While at 2240 ppm CO_2 the Antarctic temperatures are above freezing and the snow disappears leading to a much larger temperature sensitivity to the Antarctic Ice Sheet in the Modern (Fig. 4). The increases in snowfall over Antarctic does not occur in the Eocene low CO_2 cases because the Eocene cases are warmer than the equivalent Modern cases leading to above freezing temperatures in austral summer over Antarctica at all CO_2 levels.

To understand the differences in ΔT between the Eocene and Modern we must understand the relationship between the change in the energy budget between the TOA and the surface. Initial inspection of the ΔFSNT and ΔFSNS allows us to test the hypothesis that ΔFSNS at the surface induced by Antarctic Ice Sheet change is roughly the same as ΔFSNT at the TOA (Hansen et al., 1981, 1997). Here we confirm this relationship, as ΔFSNS in the Modern have very similar values at the surface compared with ΔFSNT (Table 1, Fig. 5a). In fact, across the breadth of simulations conducted for Eocene and Modern a clear linear relationship between surface and TOA shortwave radiation obtains, although there is some scatter on the order of $0.1\,\mathrm{K\,(W\,m^{-2})^{-1}}$ (Fig. 5a). Thus from the point of view of this model, global mean ΔFSNS changes are a good proxy for ΔFSNT.

The global average ΔFSNT and ΔFSNS induced by Antarctic glaciation in the Modern is generally larger than in the Eocene (Table 1, Fig. 5a). This reduced response is somewhat expected given that the Eocene land/sea distribution for Antarctica reduces the glaciated zone by $\sim 30\,\%$ in the Eocene simulations compared to the Modern (Fig. 2a, c), this does not explain the large difference in response to glaciation that occur with a given configuration (i.e. Modern or Eocene) as CO_2 is varied.

When ΔFSNT is plotted and compared with global mean ΔT, a strong linear relationship emerges in the Modern cases and ΔT increases strongly at higher CO_2 (Fig. 5b). In the Modern cases, the inspection of snow fields (figure not shown) shows that the amount of snow persisting in summer diminishes which results in increasing ΔT with

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increasing CO_2. Whereas in the Eocene, the global mean ΔT values are ~60 % less than the comparable Modern values (Fig. 5b, Table 1) and they do not increase strongly as CO_2 changes. This simple analysis indicates that there is a fundamentally different sensitivity of TOA energy budget and ΔT between the Eocene and Modern cases.

5   An analysis of Antarctica itself is necessary to separate forcing from response to establish sensitivity. When we compare the weighted temperature change ΔT Antarctica and the globally weighted forcing of the Antarctica Ice Sheet $\Delta Q_{\text{Antarctica}}$ (Fig. 5c). We find that substantial cooling occurs over Antarctica due to glaciation in both configurations, although far less local cooling occurs in the 2240 Eocene case than the comparable Modern case (Fig. 5c). But this comparison yields very different results of $\Delta Q_{\text{Antarctica}}$ compared to ΔT (Fig. 5d). Interestingly the $\Delta Q_{\text{Antarctica}}$ does not translate into a significant change in ΔT in the Eocene (Fig. 5d). Something is clearly offsetting the cooling caused by Antarctic perturbations that causes substantial cooling in the Modern (Fig. 5d). Below we show that less sea ice and negative cloud feedback processes dampens the cooling in the Eocene compared to the Modern. Similar comparisons were completed between ΔT Antarctica and $\Delta Q_{\text{Antarctica}}$ using FSNS and the general patterns of our results are robust (figure not shown), except in some Modern cases $\Delta Q_{\text{Antarctica}}$ ends up being smaller at the surface compared to the TOA (Table 1), which will become important when calculating S.

20   Differing feedbacks have important implications for S in Modern and Eocene configurations. Our calculations allow us to explore S to Antarctic glacier forcing under a variety of contexts and in a number of formulations for comparison with other work. Calculations using Eq. (9) reveal that the Modern and Eocene glacier simulations produce a wide range of values for $S_{[\text{Antarctica}]}$ in response to Antarctic glaciation holding a constant atmospheric CO_2 because the $\Delta Q_{\text{Antarctica}}$ calculated using FSNT, FSNS, and FSNSC have values that end up being slightly different. For example using Eocene cases, $\Delta Q_{\text{Antarctica}}$ FSNT includes cloud feedbacks over Antarctica and this value ends up being smaller ($\sim -0.49\,\text{W m}^{-2}$) compared to $\Delta Q_{\text{Antarctica}}$ FSNSC ($\sim -0.90\,\text{W m}^{-2}$) which includes no cloud response. This difference ends up affecting the value for S,

which on average is $\sim$0.66 K (W m^{-2})$^{-1}$ for $S_{\text{[Antarctica]}}$ FSNT and $\sim$0.46 K (W m^{-2})$^{-1}$ for $S_{\text{[Antarctica]}}$ FSNSC. Using the Modern cases, comparing $\Delta Q_{\text{Antarctica}}$ at the TOA versus the surface in some comparisons causes a change in sign between the forcing because of differences in clouds between the surface and TOA, which causes the value for $S_{\text{[Antarctica]}}$ FSNT to be higher than $S_{\text{[Antarctica]}}$ FSNS.

3.2 Antarctic glacier induced feedback response in the Modern and Eocene

To investigate the differences in the cloud response between Modern and Eocene we examine the global change in the cloud and temperature fields. Cloud feedbacks respond differently to surface perturbations in Modern and paleoclimate simulations (Thompson and Barron, 1981; Barron, 1983; Heinemann et al., 2009). Initial boundary conditions, land sea distribution, aerosols, and clouds end up being very important when calculating $\Delta Q_{\text{Antarctica}}$ due to imposed albedo forcings (Donohoe and Battisti, 2011).

As expected, the largest temperature anomaly between the glaciated and unglaciated Modern and Eocene cases occurs over the Antarctic continent (Figs. 6a, 7a). Yet, in the Eocene glaciated simulations the Southern Hemisphere is warmer than the unglaciated simulations (Fig. 6a). SWCF is commonly defined as the anomaly between clear-sky and cloudy-sky net downward ($\Downarrow$ downward minus $\Uparrow$ upward) shortwave (SW) radiation (Cess et al., 1995) calculated here at the TOA. The majority of the Southern Hemisphere warms because there is an decrease in shortwave cloud forcing (SWCF) in these regions (Fig. 6b) which increases the amount of solar radiation entering the system and acts to prevent Southern hemispheric sea ice from expanding around Antarctica (Fig. 6a). We diagnose the changes in low cloud cover (Fig. 8b) and the atmospheric greenhouse effect (Figs. 6c, 7c) which show the mechanisms that dampens the global temperature change in response to Antarctic glaciation in the Eocene.

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3.2.1 Antarctic glacier induced cloud feedback and sea ice response in the Modern and Eocene

A positive ΔSWCF is dampening the cooling in the glaciated Eocene simulations (Table 1), while the SWCF anomaly for the Modern cases is negative indicating a positive SWCF feedback while in the Eocene the SWCF anomaly is positive yielding a negative SWCF feedback. As described in Kay et al. (2011), CAM4 has improved parameterizations for stratus clouds which interact with variations in surface albedo such as sea-ice and the SWCF is not only dependent on cloud fraction but on the underlying surface albedo. This will be important in understanding changes in SWCF as the sea ice shifts between the glaciated and unglaciated simulations.

The SWCF anomalies in the Eocene simulations indicate less reflection by clouds in the glaciated cases, whereas in all the Modern experiments the clouds are reflecting more incoming radiation in the glaciated cases (Fig. 8a). The SWCF anomalies act to warm the glaciated Eocene simulation and cool nearly all the Modern glaciated simulations. One important Modern case exists when ($CO_2 = 2240$ ppm) the ΔSWCF reverses sign and becomes Eocene-like, but the cooling is nevertheless very strong and still linearly related to TOA forcing. In this high CO_2 Modern case the sea ice response in the Southern Hemisphere is large (Fig. 9c, d), more than offsetting the change in the SWCF forcing.

Changes in SWCF involves shifts in low clouds (Fig. 8b). The glaciated Eocene simulations have less low clouds than the respective unglaciated simulations from 60° S to 90° S (Fig. 8b). In the Modern cases there are increases in low cloud cover especially in the tropical regions with glaciation. We averaged over this latitude range because the decrease in Antarctic topography results in a significant decrease in low clouds over Antarctic. To verify that the global low cloud response is not just because low clouds decrease over Antarctica we globally average the low cloud response over all regions except Antarctica and show that the cloud response globally results in less low clouds

for the Eocene and more low clouds for the Modern (Fig. 8b). The total cloud forcing behaves essentially identically to the SWCF cloud forcing (Fig. 8c).

We can summarize the differences in clouds and sea ice by calculating the SWCF feedback parameter (λ) (Eq. 11) and sea ice feedback parameter in $W\,m^{-2}\,K^{-1}$ (Eq. 12). To calculate the sea ice feedback we must first calculate the globally weighted change in shortwave forcing due to the sea ice feedback in the Southern Hemisphere. This value is calculated the same way as Eq. (5) in Sect. 2.3, except the weighted sum of the ΔFSNSC values are done over the area where only sea ice anomalies occur (ΔFSNSC$_{Sea\ Ice\ Landmask}$) and the SL ratio is modified to only include the areas of sea ice (Eq. 10). We calculated the sea ice forcing in Northern Hemisphere, but found this value to be negligible in the global mean in all cases so it will not be included in the results.

$$\Delta FSNSC_{SI} = \Delta FSNSC_{Sea\ Ice\ Landmask} \cdot SL \tag{10}$$

$$\lambda_{swcf} = \left(\frac{\Delta SWCF}{\Delta T}\right) \tag{11}$$

$$\lambda_{sea\ ice} = \left(\frac{\Delta FSNSC_{SI}}{\Delta T}\right) \tag{12}$$

The Eocene glacier simulations have a negative SWCF feedback whereas in the Modern glacier simulations there is generally a positive SWCF feedback response (Fig. 9a). This is consistent with the SWCF forcing anomalies presented in Table 1 and the change in low cloud cover (Fig. 8b) which illustrate that in response to glaciation the Eocene has a reduction in low cloud cover and a negative SWCF feedback. The sea ice feedback is positive in all cases, but the magnitude of this feedback is much reduced (Fig. 9b) compared to the SWCF feedback. Thus the low cloud feedback in the Eocene simulations dominates over the sea ice feedback and acts to offset the cooling impact of adding the Antarctic Ice Sheet, whereas in the Modern the SWCF feedback and sea ice feedback are positive acting to enhance the cooling. This change in the

Modern acts to reflect more radiation in the Southern Hemisphere allowing for more sea ice area (Fig. 9c) and an increased radiative feedback response to the sea ice growth (Fig. 9b).

3.2.2 Antarctic glacier induced greenhouse effect in the Modern and Eocene

This analysis has focused on shortwave forcings, but long wave responses may also play a role in determining the temperature response to glaciation (Abbot et al., 2009). To explore the atmospheric greenhouse effect without the inclusion of clouds we use the diagnostic framework of Ramanathan and Inamdar (2006). F_c is the clearsky outgoing longwave radiation (Wm^{-2}), T_s is surface temperature, is the coefficient in the Stefan-Boltzmann equation, and G_a, is the greenhouse effect without the inclusion of clouds (Eq. 13). Rearranging to include the longwave cloud forcing we can re-write Eq. (13),

$$F_c = \sigma T_s^4 - G_a \tag{13}$$

$$F = \sigma T_s^4 - G \tag{14}$$

$$g_a = \frac{G_a}{\sigma T_s^4} \tag{15}$$

where $G = G_a + \text{LWCF}$ and F now equals the outgoing longwave radiation for cloudy skies giving us an expression for the greenhouse effect with the inclusion of clouds (Eq. 14). For our purposes, we want to solve for G_a, which is the greenhouse effect without the inclusion of clouds. We can then normalize G_a by σT_s^{-4} to get a value, g_a, which removes the variations in T from the greenhouse effect (Eq. 15) (Ramanathan and Inamdar, 2006). In Fig. 6c, we show this normalized percentage for g_a as an anomaly for the Eocene and the areas of warming in the Southern Hemisphere Fig. 6a are associated with an increase in the greenhouse forcing. The globally weighted average for the g_a anomaly is negligible around a tenth of a percent, but the regional changes in greenhouse effect explain some of the warming occurring in the Southern Hemisphere (Fig. 6c).

In the Modern glacier simulations a clearer pattern emerges over the tropical terrestrial surfaces which cool significantly and the decreases in temperature align with a reduction in the greenhouse forcing (Fig. 7c). This is not the case in the Eocene glaciated simulation as there is little change in the greenhouse effect over the terrestrial land surfaces (Fig. 6c). In addition, in the Modern glaciated simulations there are decreases in the greenhouse effect in the Southern Hemisphere (Fig. 7c), especially around South America and Africa where in the Eocene glaciated cases there is an increase in the greenhouse effect (Fig. 6c).

4 Discussion

4.1 Antarctic glacier response in Eocene and Modern

This study finds that the forcing due to the Antarctic Ice Sheet is constant at $\sim$0.6 W m^{-2} in the Eocene, regardless of CO_2 level, whereas the forcing increases from nearly 0 to 1.4 W m^{-2} as CO_2 is increased in the Modern cases. Global cooling in the Eocene due to the introduction of substantial Antarctic Glaciation is much less ($\sim$0.30 K) than the Modern world ($\sim$0.72 K) for a wide range of CO_2 values. Additionally, cooling in the Eocene is substantially less than in the Modern with comparable forcing values. In the Eocene, regional impacts due to glaciation in the Southern Hemisphere are large but, globally the changes are negligible. In the Modern, positive feedbacks overwhelm negative feedbacks and cooling is more widespread. Comparison between our glaciated versus unglaciated simulations indicate that a larger ΔT ($\sim$0.72 K) in the Modern simulations (as opposed to $\sim$0.30 K in the Eocene cases) is related to an enhanced sea ice growth compared to the Eocene (Fig. 8c). In the Modern cases, a strong sea ice response dominates over the weaker cloud feedbacks, although it should be noted that SWCF is not independent of the underlying surface albedo. But, in the Eocene, without substantial (positive) sea ice feedbacks the underlying negative low cloud feedback dominates and leads to little global mean temperature change. Cooling is substantial in

some parts of the South Pacific Ocean and in some continental interiors in the Northern Hemisphere, but this cooling is nearly offset by substantial (∼3.0 K) warming over the subtropical ocean, the South Atlantic and Northern Eurasia.

To our knowledge no recent climate modelling study has focused explicitly on calculating the climate sensitivity to the removal and addition of the Antarctic Ice Sheet in Eocene and Modern contexts, so comparison with prior work is difficult. Nevertheless we can compare the results generally with other studies and provide constraints on the physical processes explored in this study. Most prior work has concentrated on the potential impact on global climate to changes in Northern Hemisphere cryosphere in the Modern and near-future (Flanner et al., 2011; Hudson, 2011; Graversen and Wang, 2009; Lunt et al., 2004; Stone et al., 2010), on the Last Glacial Maximum (LGM) (Chiang et al., 2003; Brady et al., 2012; Pausata et al., 2008; Brandefelt and Otto-Bliesner, 2009; Hewitt and Mitchell, 1997), or looked at the problem from the glacial to interglacial paleoclimate perspective (Yin et al., 2009; Manabe and Broccoli, 1985; Koehler et al., 2010) or a Pliocene perspective (Lunt et al., 2008; Koenig et al., 2011). The work of Deconto and Pollard is most comparable in terms of the time intervals covered. We discuss each type of previous study in turn, below.

4.2 Antarctic glacier response in Eocene and Modern and comparison with Modern and near-future results

A study by Flanner et al. (2011), showed that the mean radiative impact of the Northern Hemisphere cryosphere on the Earth System is 1.1–1.3 W m^{-2} and this value nearly doubles when clouds are removed from the calculation. Hudson (2011) found that if all the sea-ice was removed in the Arctic than the forcing would equal ∼0.7 W m^{-2}, but this value could be dependent on whether clouds increase during the summer months in Arctic. Similarly, Graversen and Wang (2009) found that when fixing surface albedo and elevating atmospheric carbon dioxide in a climate model that water vapor and clouds accounted for almost half the total change in the Arctic region. These results highlight the importance of clouds which ultimately could be affecting the value of sensitivity

in models and in observational datasets (Bony and Dufresne, 2005; Medeiros et al., 2008; Trenberth and Fasullo, 2009).

A study by Hansen and Nazarenko (2004) found that although forcings may have similar magnitudes this may not translate into identical changes in global mean temperature. This "efficacy" term is defined as the global temperature change per unit forcing for a chosen climate variable compared against the standard CO_2 forcing (Hansen et al., 2005). A major conclusion of these studies is that efficacy values for different forcings is not expected to be constant between different climate states. Here the climate forcing of the Antarctic Ice Sheet in our simulations is not constant and the climate change in the Eocene due to the Antarctic Ice Sheet is much smaller than one would expect from a similar Wm^{-2} forcing of CO_2.

Studies exploring future climate change of Antarctica found that significant portions of the Antarctic Ice Sheet can be melted by increasing atmospheric CO_2. Using a model of intermediate complexity, Huybrechts et al. (2011) slowly increased CO_2 to 4 × pre-industrial levels over 3000 models years and found that polar temperatures were 10 K warmer than today. These large temperature increases in the polar regions decreased the size of the Antarctic Ice Sheet, but it must be noted that this model was not fully equilibrated. Using the same model of intermediate complexity, Goelzer et al. (2011) found that the initial size of the Greenland and Antarctic Ice Sheets are important in determining climate sensitivity especially if sea ice growth is dynamic.

4.3 Antarctic glacier response in Eocene and Modern and comparison with paleoclimate studies

For LGM modelling a typical result is that the total temperature change introduced by non-GHG forcing is ∼2 Wm^{-2} (Braconnot et al., 2012) and the resulting temperature change (∼2.5 K) is consistent with that inferred from the sensitivity from a doubling of CO_2 (Brady et al., 2012). Braconnot et al. (2012) also found that changing ice-sheet elevation was important in explaining some of the regional cooling over the ice sheet regions. Brandefelt and Otto-Bliesner (2008) found that a combined glacial and trace

gas forcing in the LGM simulations caused a MAT decrease of ∼1–2 K and as much as a 5–15 K regional cooling. Hewitt and Mitchell (1997), found a climate sensitivity of ∼1 W m^{-2} K^{-1} to LGM conditions, but found that the majority of the cooling was caused by changes in cloud feedbacks in addition to surface albedo changes. The cloud feedback response described in the previous study is very similar to the cloud response in the Modern glaciated versus unglaciated simulations in this study.

On glacial-interglacial time scales, there are studies that have used the framework developed by Hansen et al. (2007), to develop a time series of radiative forcing over the last 500 000 yr (Rohling et al., 2012; Kohler et al., 2010). These studies calculate the changes in radiative forcing based on surface radiative forcing, ignoring cloud and water vapor feedbacks, which we would argue are an important component in understanding the integrated radiative forcing and sensitivity to changes in ice sheets over these glacial and interglacial time periods. We believe that surface albedo perturbations, such as might be reconstructed from inferred land ice and sea ice distributions, are only one small part of the story because of changes that could occur in clouds and water vapor.

This poses serious problems for estimating sensitivity of the climate system to large perturbations from paleoclimate data across the EOT. If one could reconstruct the TOA albedo change for these kinds of perturbations then it would be relatively straightforward to infer the forcing of adding or removing an ice sheet. But, clouds–for which we have no proxies–can play a large, and difficult-to-predict role in enhancing or offsetting the better constrained surface albedo perturbation. If the sign of the feedbacks was clearly understood (sea ice being a good example) and their magnitude was constrained then approximate and reasonable bounds could be placed on estimates of sensitivity. But, for clouds, especially low clouds, that remain the most important source of uncertainty in the Modern physical climate system (Bony and Dufresne, 2005; Trenberth and Fasullo, 2009) there is no a priori reason to expect that the feedbacks should be either positive or negative or constant as a function of climate state. Indeed, in the

examples considered here the sign of the low cloud SWCF feedback changes between the Eocene and Modern.

Lunt et al. (2012) conducted a recent study in which they altered Greenland and Antarctic topography and albedo in Pliocene contexts and found that the regional im-
pacts of these alterations was significant, but the global response to these variables was weak (~10 % of the total, or ~0.30 K). Surface albedo forcing was deemed more important than cloud forcing in the Lunt et al. (2012) study, but the diagnostic frame-work used does not account for correlations between surface and cloud albedos. Other Pliocene modelling studies have focused on understanding the role of the Greenland
Ice Sheet in affecting climate sensitivity (Lunt et al., 2008; Koenig et al., 2011). Their results found that the Greenland Ice Sheet has strong regional control on temperature sensitivity, but the global impact to changing the Greenland Ice Sheet is negligible.

4.4 Antarctic glacier response in Eocene and Modern and comparison with EOT studies

The coupled atmosphere-ice sheet modelling of Deconto and Pollard (2003, 2007) is the closest modelling approach to that tried here although those studies were focused on a very different problem and did not present results showing how global mean tem-perature was affected by the Antarctic Ice Sheet. Interestingly, those studies show a ΔT of 0.80 K from such a perturbation (DeConto, personal communication), which is
significantly larger than the Eocene results presented here, although within the range of Modern values we have calculated. This result also involved changes in Earth's or-bital parameters, which also influences global mean temperature. This makes it difficult to directly compare with our results, but based on our own preliminary work where we changed obliquity and glaciation like the Deconto and Pollard simulations, we estimate
that half of the 0.80 K cooling could be due to orbital changes and not to the ice sheet itself.

The importance of forcing factors and feedbacks for which no proxies exist also complicates attempts at evaluating model predictions with proxies across the EOT

although some general statements can be made. The temperature change associated with adding the Antarctic Ice Sheet and dropping atmospheric CO_2 by 560 ppm produces a good match for the cooling detected in the proxy record, especially the Southern Hemisphere ODP sites 277, 511, and 689 (Liu et al., 2009; Macksensen and Ehrmann, 1992) and the cooling in the Northern Hemisphere sites 913, 336, 643, and 985 (Liu et al., 2009; Eldrett et al., 2009). This combined forcing is also able to match the terrestrial record temperature drop of 3–8 K over North America (Zanazzi et al., 2007). Whereas adding the Antarctic Ice Sheet at constant atmospheric CO_2 produces warming in the Southern Hemisphere in contrast to the proxy record described above. This highlights the importance in CO_2 forcing for causing cooling at the EOT (Pagani et al., 2011; DeConto and Pollard, 2003). These results also suggest that CESM1.0 has strong negative feedbacks (or too weak, or neglected positive feedbacks) given that including both a drop of CO_2 from 1120 to 560 and the growth of a large Antarctic Ice Sheet cools the Eocene simulations by 3.7 K. This temperature drop yields a model derived ESS of $1.05\,\mathrm{K\,(W\,m^{-2})^{-1}}$, as compared with the value of $\sim 1.5\,\mathrm{K\,(W\,m^{-2})^{-1}}$ calculated from EOT proxies (see Sect. 1.1).

So, in short the estimated forcing and global mean temperature changes are well within those expected from prior work, but an exact comparison is currently impossible. Additional simulations invoking similar experimental methodologies and diagnostics are required to ascertain whether our results are robust or strongly model dependent. Given the importance of low clouds to our results it is likely that the results of this study will only be as robust as the spread of model differences in the representation of low clouds.

4.5 Limitations of this study

This study has many limitations that bear mentioning. More realistic representations of the Modern ice-free conditions and Eocene glaciated cases may lead to different results. Treatment of sea level variations (Gasson et al., 2012) and changes in orbital parameters (Lee and Poulsen, 2009; Sloan and Huber, 2001), which we have ignored,

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is an important weakness of this study. The results may also be sensitive to model resolution treatment of subgrid-scale processes such as orographic wave drag.

The uncertainty in low cloud feedbacks makes it difficult to constrain temperature sensitivity in Eocene and Modern to changes in the Antarctic Ice Sheet and because the negative low cloud feedback may be a model dependent, additional simulations using different models and cloud parameterizations need to be conducted. For example implementing the Antarctic Ice Sheet simulations into the recently released CAM5.0 which has a more complete cloud parameterizations and aerosol interaction in comparison to CAM4.0 should provide more insights into this problem, although not solve it.

In this study, we did not explore the oceans response to changes in Antarctic glaciation in a fully coupled atmosphere, land, ice, and ocean simulation. Undoubtedly, the ocean response to Antarctica glaciation is important and in the future we need to run fully coupled climate modelling simulations. We argue, that this coupled approach makes feedback analysis even more difficult and we believe the simpler approach here has some merits.

To understand the full evolution of the climate feedbacks within the Eocene simulations and the physical mechanisms inducing the negative cloud feedback, we need to look at how the atmosphere responds to the instantaneous glacier forcing. To do these calculations we would need to explore other approaches (Gregory et al., 2002; Shell et al., 2000) which might give different perspectives on sensitivity of Antarctic Ice Sheet forcing in Eocene and Modern.

5 Conclusions

For the first time we have calculated $S_{[Antarctica]}$ due to the removal of the Antarctic Ice Sheet using a global climate model in Modern and Eocene contexts. To date, no climate modelling study has separated the Antarctic Ice Sheet component in terms of $S_{[Antarctica]}$ for these time periods, and we hope the results can be used to compare

against proxy data derived climate sensitivity estimates. In the future it will be important for modelling groups to simulate Antarctic Ice Sheet sensitivity using different climate models, at varying resolutions, and using different cloud parameterizations to evaluate the robustness of the results presented in this study.

The results lead to 3 major conclusions about the climatic impacts of the Antarctic Ice Sheet in Modern and Eocene climate. The results we use to draw our conclusions are occurring within one model framework and the results should be taken within this context, but the response is robust within CESM1.0.

1. The equilibrium response in the radiation budget induced by Antarctic glaciation including the global feedbacks results in a linear relationship between ΔFSNS and ΔFSNT.

2. Adding the Antarctic Ice Sheet to the Eocene greenhouse climate has a strong negative low cloud feedback response resulting in minimal global cooling even though the $\Delta Q_{\text{Antarctica}}$ is substantial. The results suggests that Antarctic glaciation at the EOT transition may not have had a significant global temperature response because of negative feedbacks.

3. Removing the Antarctic Ice Sheet in the Modern simulations at 560 and 1120 ppm CO_2 has a reduced temperature sensitivity compared to the removing the glacier at 2240 ppm because our imposed albedo change (at the lower CO_2 levels) is offset by increased snowfall and year round freezing temperatures over Antarctica.

The importance of model sensitivity – especially to the low cloud parameterization – is one of the main lessons of this study. Acknowledging the fact that this is only one particular model and an idealized study, we can nevertheless conclude – for this one model – that growth of Antarctic land ice played little role directly or through fast feedbacks in cooling the world at the Eocene Oligocene Transition (< 0.30 K). Thus, in this model, the Antarctic Ice Sheet at the EOT plays a relatively minor role in global mean climate change. Whereas in the Modern the cloud and sea-ice feedbacks induced by Antarctic glaciation strongly enhance the global cooling response.

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The reality is that if ice sheets are strongly mediated by poorly constrained, fast cloud feedbacks then models are likely to give divergent results in regards to sensitivity to ice sheet forcing. Using proxy-data based ESS approaches may seem at first to be a means to avoid performing this separation but the results of this study indicate that the feedbacks involved may be strongly state dependent – i.e. the Eocene is not a good analogue for the Modern (Haywood et al., 2011; Francis and Williams, 2011) – in which case calculating ESS across the EOT may have little direct value for making inferences about the future. This also specifically suggests that there may not be much gained by using proxy data records from the EOT and projecting by analogy into the future because unravelling the different forcings and feedbacks in the past can not be done from proxy records. Since there are no cloud proxies, these neglected cloud feedbacks will be incorrectly attributed instead to the processes observed in the proxy record, thus leading to inflated or reduced estimates of paleoclimate feedbacks. Instead, progress will likely rely on using proxies from the EOT to discriminate between models that match proxies and those that do not and using those models to project into the future.

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Table 1. Eocene and Modern glaciated minus unglaciated simulations labeled by experiment type, time interval, orography change (labelled Y(yes) and N(no)), albedo change (labelled with Y and N), and atmospheric CO_2 in ppm. Globally weighted anomalies are given for ΔT in (K), ΔSWCF, ΔLWCF, and total cloud forcing SWCF + LWCF in $\mathrm{W\,m^{-2}}$. Second, the table lists ΔFSNT and the globally weighted Antarctic forcings ($\Delta Q_{\mathrm{Antarctica}}$) calculated using FSNT, FSNS, and FSNSC. Third, the table gives values for the surface temperature sensitivity induced by the changes in albedo, topography, and CO_2, like $\Delta T_{(\alpha)}$ the change in surface temperature due to the albedo forcing of the Antarctic Ice Sheet, and $\Delta T_{(\alpha+\mathrm{oro})}$ the temperature change due to the albedo of the ice sheet and the topography of the ice sheet. Last, the table calculates ESS and S in $\mathrm{K\,(W\,m^{-2})^{-1}}$ using the different ΔT and $\Delta Q_{\mathrm{Antarctica}}$ using FSNT, FSNS, and FSNSC values. Where $S_{[\mathrm{Antarctica,\ CO_2}]}$ is the sensitivity to changing CO_2, Antarctic albedo, Antarctic topography and $S_{[\mathrm{Antarctica}]}$ is the sensitivity of climate to changes in Antarctic Ice Sheet holding atmospheric CO_2 constant. Complete descriptions of the equations are written in detail in methods Sect. 2.3.

Experiment Comparison	Time Interval	Oro Change	Albedo Change	CO_2	ΔT	ΔSWCF	ΔLWCF	Total cloud forcing	ΔFSNT$_{(\alpha)}$	ΔFSNS$_{(\alpha)}$	ΔFSNT$_{(\alpha+\mathrm{oro})}$	ΔFSNS$_{(\alpha+\mathrm{oro})}$
α	MODERN	N	Y	1120	−1.14	−0.21	−0.23	−0.44	−1.84	−1.59	−	−
α	MODERN	N	Y	560	−0.86	−0.26	−0.21	−0.47	−1.70	−1.64	−	−
α + oro	MODERN	Y	Y	2240	−1.22	0.52	−0.37	0.39	−	−	−1.69	−1.43
α + oro	MODERN	Y	Y	1120	−0.68	−0.31	−0.13	−0.43	−	−	−1.01	−1.16
α + oro	MODERN	Y	Y	560	−0.18	−0.60	−0.13	−0.73	−	−	−0.88	−0.49
α + oro + CO_2	MODERN	Y	Y	1120–560	2.94	−0.36	−0.79	−1.15	−	−	−	−
CO_2	MODERN	N	N	2240–1120	3.81	0.57	−0.79	−0.22	−	−	−	−
CO_2	MODERN	N	N	1120–560	3.11	0.24	−0.66	−0.42	−	−	−	−
α	EOCENE	N	Y	1120	−0.36	0.18	−0.11	0.76	−0.64	−0.43	−	−
α	EOCENE	N	Y	560	−0.27	0.06	−0.10	0.43	−0.39	−0.25	−	−
α + oro	EOCENE	Y	Y	2240	−0.26	1.14	−0.10	1.04	−	−	−0.30	−0.26
α + oro	EOCENE	Y	Y	1120	−0.17	1.40	−0.14	1.26	−	−	−0.27	−0.29
α + oro	EOCENE	Y	Y	560	−0.29	0.91	−0.02	0.89	−	−	−0.55	−0.45
α + oro + CO_2	EOCENE	Y	Y	1120-560	3.74	−1.24	−0.73	−1.97	−	−	−	−
CO_2	EOCENE	N	N	2240-1120	2.91	1.06	−1.37	−0.31	−	−	−	−
CO_2	EOCENE	N	N	1120-560	3.46	−0.31	−0.78	−1.09	−	−	−	−

Table 1. Continued.

Experiment Comparison	$\Delta FSNT_{(\alpha+oro+CO_2)\_toa}$	$\Delta FSNS_{(\alpha+oro+CO_2)}$	$\Delta Q_{Antarctica\_toa}$	$\Delta Q_{Antarctica\_surface}$	$\Delta Q_{Antarctica\_clearsky}$	$\Delta T_{(\alpha)}$	$\Delta T_{(\alpha+oro)}$
α	–	–	−0.96	−0.92	−1.23	−1.14	–
α	–	–	−0.68	−0.63	−0.88	−0.86	–
α + oro	–	–	−1.37	−1.19	−1.88	–	−1.22
α + oro	–	–	−0.18	0.04	−0.56	–	−0.68
α + oro	–	–	−0.03	0.18	0.09	–	−0.18
α + oro + CO_2	2.07	−0.26	0.25	−0.02	0.38	–	–
CO_2	–	–				–	–
CO_2	–	–				–	–
α	–	–	−0.54	−0.53	−0.84	−0.36	–
α	–	–	−0.13	−0.12	−0.22	−0.27	–
α + oro	–	–	−0.68	−0.59	−1.29	–	−0.26
α + oro	–	–	−0.62	−0.51	−1.24	–	−0.17
α + oro	–	–	−0.49	−0.35	−0.93	–	−0.29
α + oro + CO_2	3.02	0.67	0.68	0.52	1.35	–	–
CO_2	–	–				–	–
CO_2	–	–				–	–

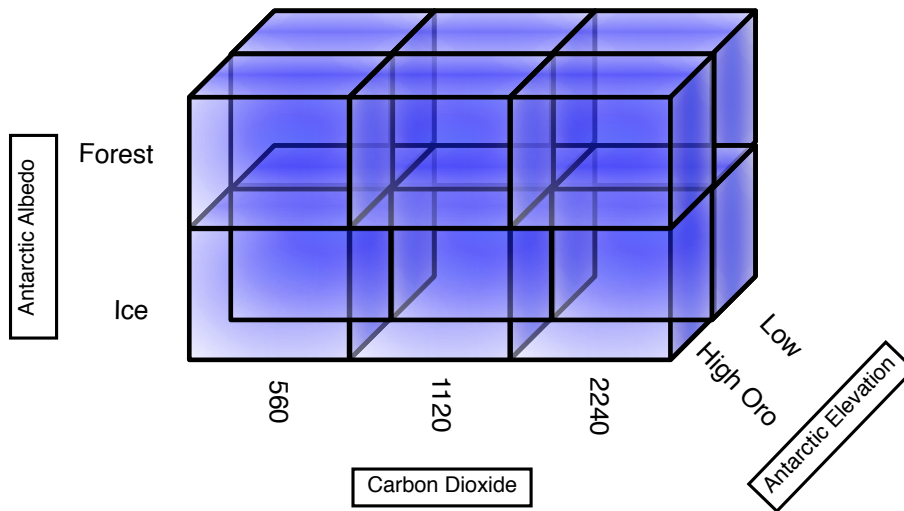| Experiment Comparison	ESS	$S_{[Antarctica\ CO_2]\_toa}$	$S_{[Antarctica,\ CO_2]\_surface}$	$S_{[Antarctica,\ CO_2]\_clearsky}$	$S_{[Antarctica]\_toa}$	$S_{[Antarctic]\_surface}$	$S_{[Antarctica]\_clearsky}$
α	–	–	–		1.19	1.24	0.93
α	–	–	–		1.26	1.37	0.98
α + oro	–	–	–		0.89	1.03	0.65
α + oro	–	–	–		3.78	−19.43	1.21
α + oro	–	–	–		7.20	−1.00	−2.00
α + oro + CO_2	0.84	0.78	0.84	0.76	–	–	
CO_2	–	–	–		–	–	
CO_2	–	–	–		–	–	
α	–	–	–		0.67	0.68	0.43
α	–	–	–		2.08	2.25	1.23
α + oro	–	–	–		0.38	0.44	0.20
α + oro	–	–	–		0.27	0.33	0.14
α + oro	–	–	–		0.59	0.83	0.31
α + oro + CO_2	1.06	0.89	0.93	0.77	–	–	
CO_2	–	–	–		–	–	
CO_2	–	–	–		–	–	

Fig. 1. Schematic describing the suite of Modern and Eocene Antarctic glacier simulations that were completed in this study. Along the x-axis we plot the levels of CO_2, along the y-axis we plot the changes in albedo over Antarctica, and along the z-axis we plot the topographical changes over Antarctica.

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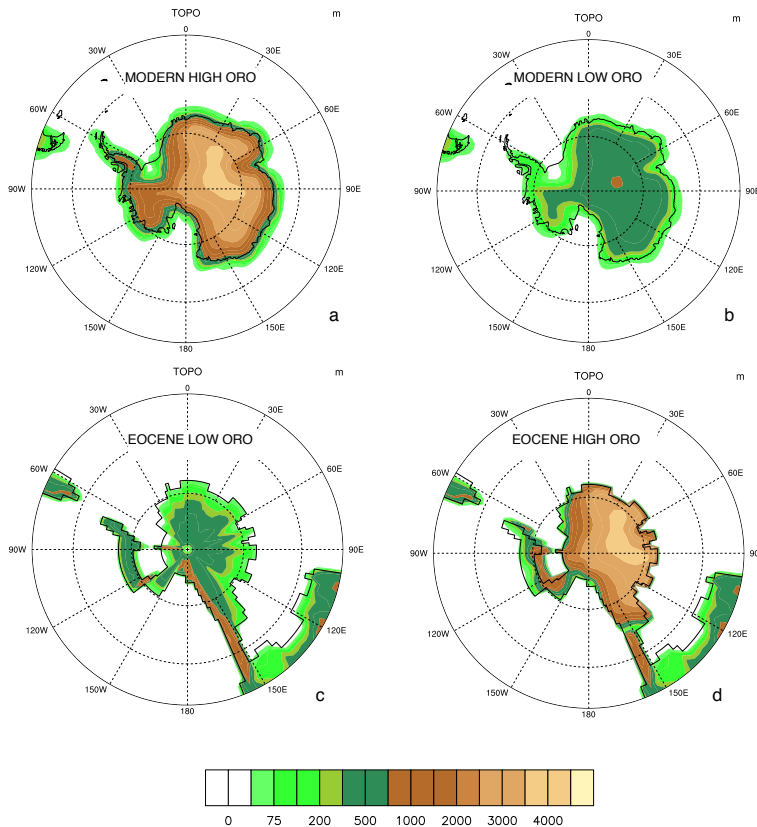Back | Close

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Fig. 2. (a) Antarctic topography in the Modern glaciated high topography simulations (m). **(b)** Antarctic topography used in the unglaciated Modern low topography simulations. **(c)** Antarctic topography in the unglaciated Eocene simulation. **(d)** Antarctic topography in the Eocene glaciated simulation based off of the Modern day Antarctic height.

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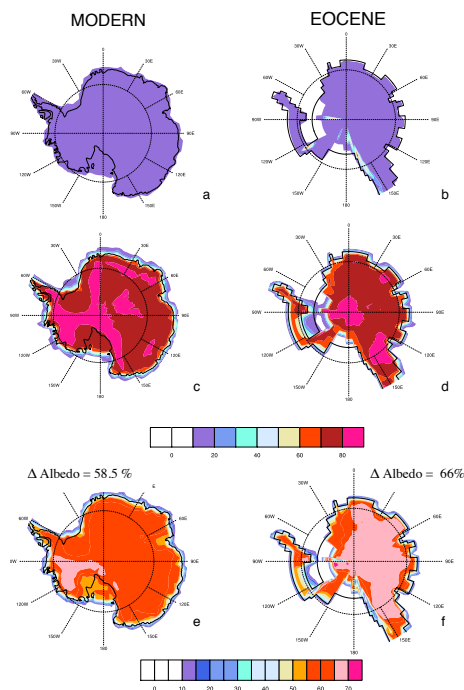
Fig. 3. Surface albedos (%) for Modern and Eocene glaciated and unglaciated simulations averaged over the austral summer at 2240 ppm CO_2. **(a)** Modern unglaciated simulation, **(b)** Eocene unglaciated simulation, **(c)** Modern glaciated simulation, **(d)** Eocene glaciated simulation, **(e)** Modern glaciated versus unglaciated, **(f)** Eocene glaciated versus unglaciated.
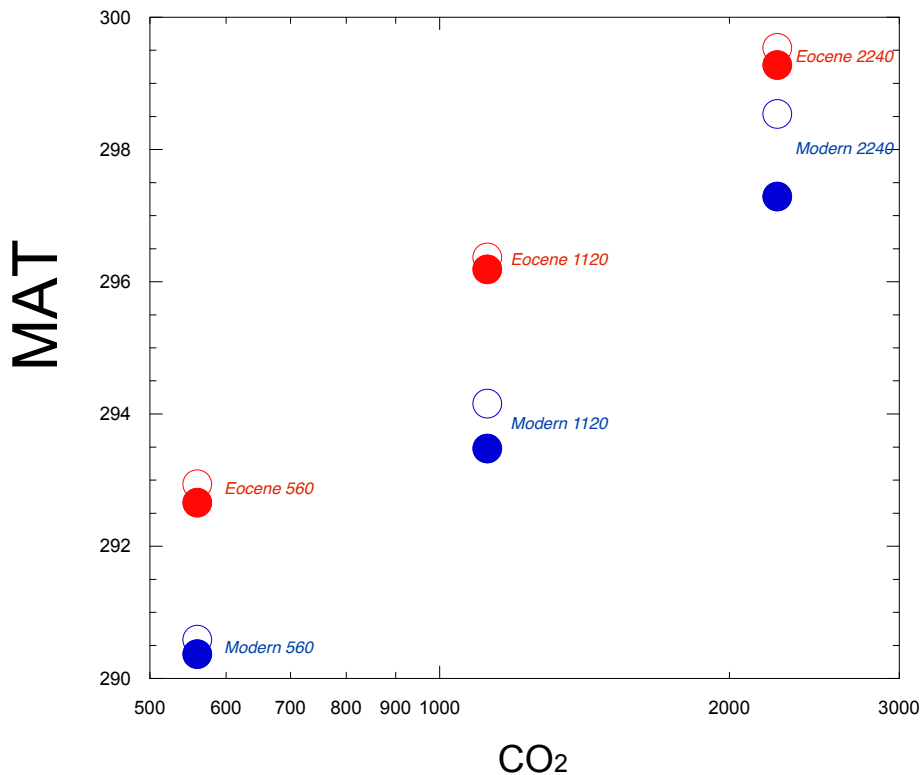
Fig. 4. The unshaded red colored circles represent the Eocene unglaciated simulations, while the filled red circles represent Eocene glaciated simulations. The unshaded blue circles represent the Modern unglaciated simulations, while the filled blue circles represent Modern glaciated simulations. The atmospheric CO_2 levels in ppm (560, 1120, 2240) is plotted on a logarithmic scale on x-axis and the MAT (K) for the glaciated and unglaciated simulations is plotted along the y-axis.

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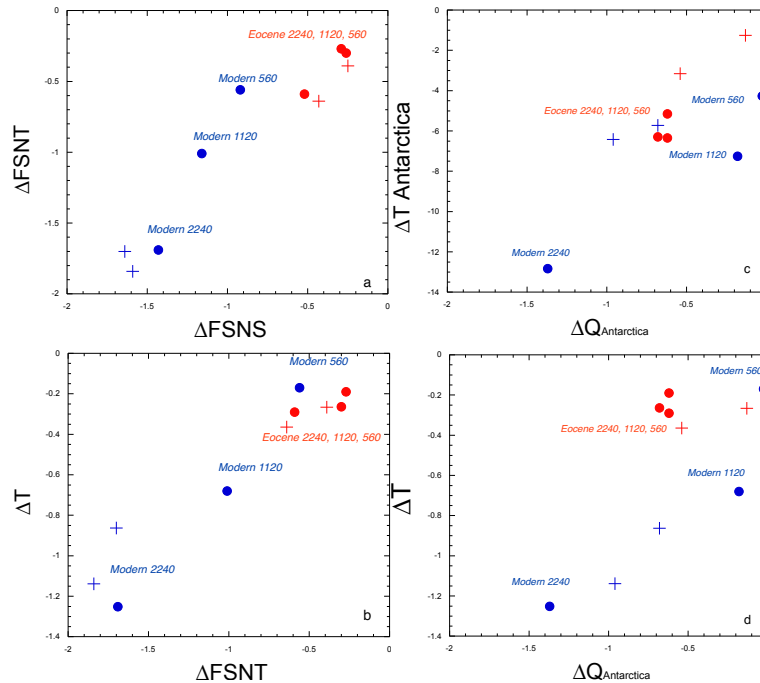
Fig. 5. The colored circles represent slab ocean glaciated versus unglaciated comparisons where albedo and topography are changed and the circles are labeled with their corresponding CO_2 level, time period, and color (Eocene-red (italic), Modern-blue (bold)) (Table 1). The colored red crosses are the Eocene (colored red, italic) and Modern (colored blue, bold) glaciated versus unglaciated comparisons where only albedo was changed (Table 1). **(a)** ΔFSNS ($\mathrm{W\,m^{-2}}$) along the x-axis compared against ΔFSNT ($\mathrm{W\,m^{-2}}$) along the y-axis. **(b)** ΔFSNT along the y-axis compared against ΔT along the y-axis. Same colored dots and crosses described in **(a, b)**. **(c)** $\Delta Q_{\mathrm{Antarctica}}$ ($\mathrm{W\,m^{-2}}$) at the TOA along the x-axis and ΔT Antarctica along the y-axis. **(d)** $\Delta Q_{\mathrm{Antarctica}}$ at the TOA ($\mathrm{W\,m^{-2}}$) along the x-axis and global ΔT along the y-axis. Definitions for $\Delta Q_{\mathrm{Antarctica}}$, ΔT, ΔT Antarctica can be found in methods Sect. 2.3.

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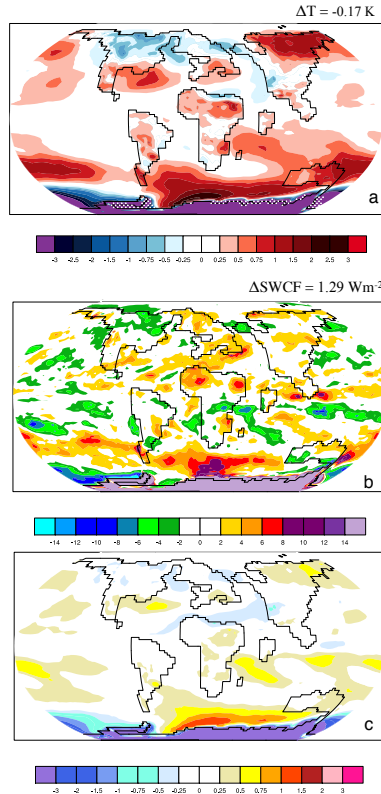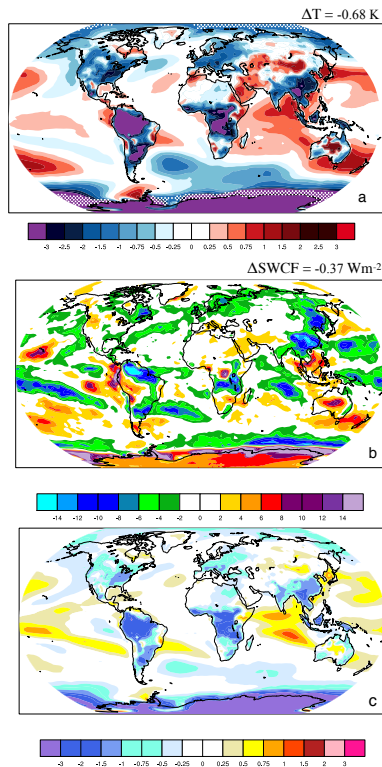
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Fig. 6. Glaciated minus unglaciated Eocene simulation at 1120 ppm CO_2 (($\alpha + $oro) experiment-highlighted with a dark grey shade) in Table 1. **(a)** Annually averaged anomalies for surface temperature (K) as the contour and the sea ice anomalies stippled in white, **(b)** shortwave cloud forcing (SWCF) in Wm^{-2}, and **(c)** normalized g_a (greenhouse effect without clouds) anomaly in %. The calculation for g_a is described in results Sect. 3.2.2, Eq. (15).

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Fig. 7. Glaciated minus unglaciated Modern simulation at 1120 ppm CO_2 ((α + oro) experiment-highlighted with a light grey shade) in Table 1. **(a)** Annually averaged anomalies for surface temperature anomalies (K) as the contour and the sea ice anomalies stippled in white, **(b)** short-wave cloud forcing (SWCF) in W m^{-2}, and **(c)** normalized g_a (greenhouse effect without clouds) in %. The calculation for g_a is described in results Sect. 3.2.2, Eq. (15).

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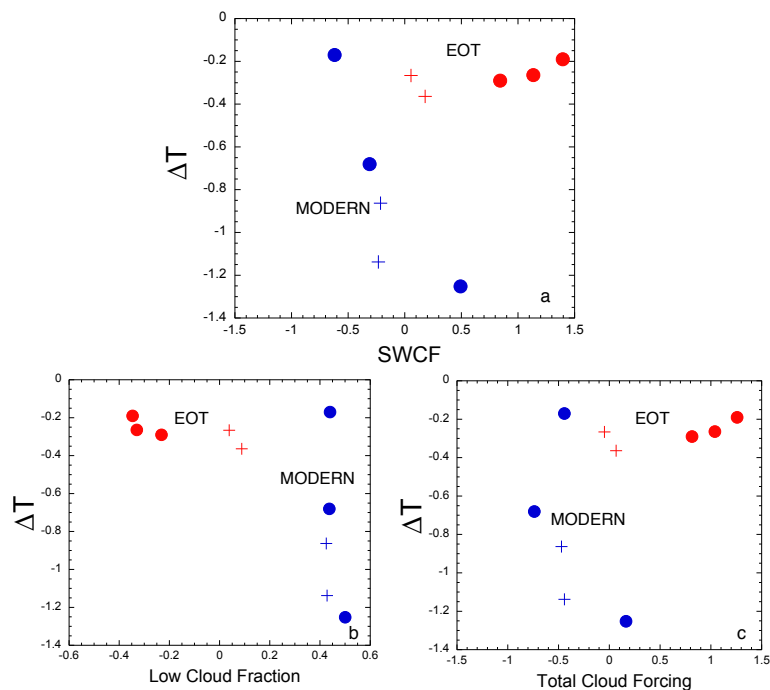Printer-friendly Version

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Fig. 8. Same colored dots and crosses described in Fig. 5. **(a)** The anomalous SWCF forcing anomaly (W m^{-2}) along the x-axis, compared against ΔT along the y-axis. **(b)** The anomalous low cloud fraction in (%) (averaged from 60° S to 90° S) along the x-axis, compared against ΔT along the y-axis. **(c)** The anomalous total cloud forcing anomaly (W m^{-2}) along the x-axis, compared against ΔT along the y-axis.
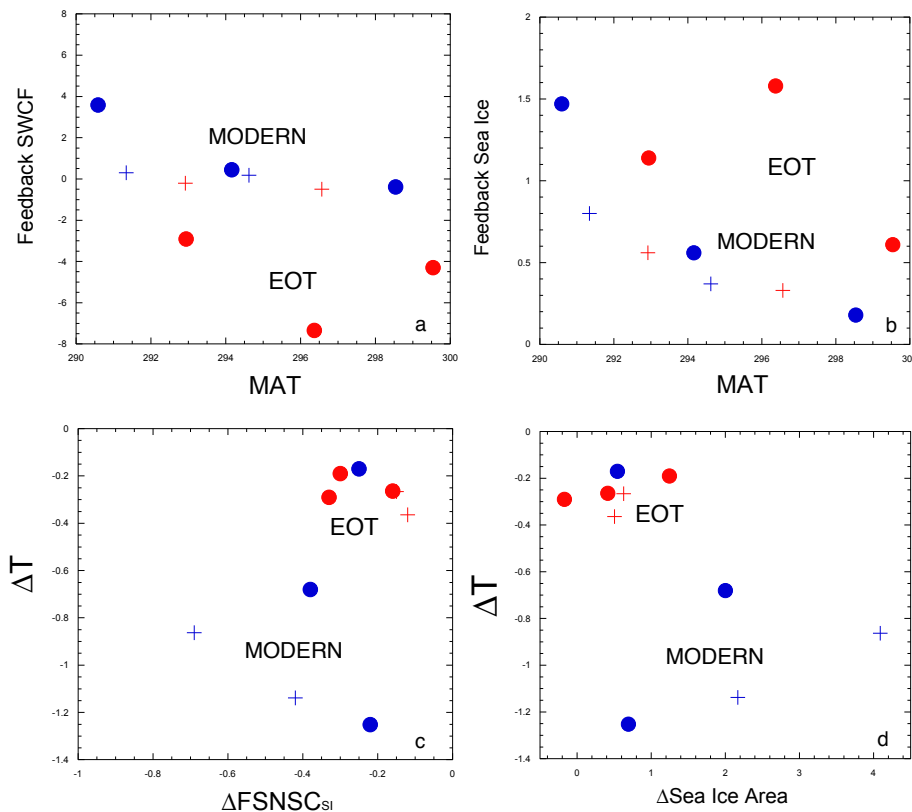
Fig. 9. Same colored dots and crosses described in Fig. 5. **(a)** The SWCF feedback ($\mathrm{W\,m^{-2}\,K^{-1}}$) using Eq. (12) described in Sect. 3.2.1 along the y-axis, plotted against the MAT of the unglaciated simulations. **(b)** The sea ice feedback ($\mathrm{W\,m^{-2}\,K^{-1}}$) using Eq. (11) described in Sect. 3.2.1 along the y-axis, plotted against the MAT of the unglaciated simulations, **(c)** $\Delta \mathrm{FSNSC_{SI}}$ along the y-axis compared against ΔT, **(d)** anomalous sea ice area ($\mathrm{m^2}$) compared against ΔT.

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