Estimating the Contribution of Sea Ice Response to Climate Sensitivity in a Climate Model

KEN CALDEIRA
Department of Global Ecology, Carnegie Institution for Science, Stanford, California

IVANA CVIJANOVIC
Department of Global Ecology, Carnegie Institution for Science, Stanford, California, and Centre for Ice and Climate, Niels Bohr Institute, University of Copenhagen, Copenhagen, Denmark

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ABSTRACT

The response of sea ice to climate change affects Earth’s radiative properties in ways that contribute to yet more climate change. Here, a configuration of the Community Earth System Model, version 1.0.4 (CESM 1.0.4), with a slab ocean model and a thermodynamic–dynamic sea ice model is used to investigate the overall contribution to climate sensitivity of feedbacks associated with the sea ice loss. In simulations in which sea ice is not present and ocean temperatures are allowed to fall below freezing, the climate feedback parameter averages \( \frac{1.31}{2} \text{ W m}^{-2} \text{ K}^{-2} \); the value obtained for control simulations with active sea ice is \( \frac{1.05}{2} \text{ W m}^{-2} \text{ K}^{-2} \), indicating that, in this configuration of CESM1.0.4, sea ice response accounts for \( \frac{20}{2} \% \) of climate sensitivity to an imposed change in radiative forcing. In this model, the effect of sea ice response on the longwave climate feedback parameter is nearly half as important as its effect on the shortwave climate feedback parameter. Further, it is shown that the strength of the overall sea ice feedback can be related to 1) the sensitivity of sea ice area to changes in temperature and 2) the sensitivity of sea ice radiative forcing to changes in sea ice area. An alternative method of disabling sea ice response leads to similar conclusions. It is estimated that the presence of sea ice in the preindustrial control simulation has a climate effect equivalent to \( \frac{3}{2} \text{ W m}^{-2} \) of radiative forcing.

1. Introduction

It is widely recognized that the sea ice loss amplifies the Arctic warming as a consequence of surface albedo feedback and “ice insulation feedback” (altered atmosphere–ocean heat and moisture exchange) (Holland et al. 2001; Holland and Bitz 2003; Hall 2004; Chapman and Walsh 2007; Serreze et al. 2009; Screen and Simmonds 2010; Overland et al. 2011). Other contributing factors to enhanced Arctic warming (also affected by sea ice decline) include changes in atmospheric and oceanic heat transport, water vapor content, and cloud cover (Alexeev et al. 2005; Cai 2005; Langen and Alexeev 2007; Graversen et al. 2008; Graversen and Wang 2009; Alexeev and Jackson 2013; Kapsch et al. 2013).

The effects of sea ice loss are felt globally, although attenuated far from the poles: in the absence of sea ice response, tropical regions also experience less global warming (Meehl and Washington 1990; Rind et al. 1995; Hall 2004; Caji and Caldeira 2014, manuscript submitted to Climate Dyn.). Furthermore, sea ice decline has been shown to affect midlatitude circulation patterns (Francis et al. 2009; Overland and Wang 2010; Deser et al. 2010; Screen and Simmonds 2013).

The overall influence of sea ice loss on global climate sensitivity and global temperature increase in CO2 doubling simulations has been previously addressed by Ingram et al. (1989) and Rind et al. (1995, 1997). Climate sensitivity to a change in radiative forcing is inversely proportional to the climate feedback parameter. Rind et al. (1995) found the climate feedback parameter, defined as the initial tropopause energy imbalance divided
by the equilibrium temperature response, to equal 0.95 W m\(^{-2}\) K\(^{-1}\) with sea ice response and 1.51 W m\(^{-2}\) K\(^{-1}\) without sea ice response. Thus, the sea ice response accounts for 37% of the temperature response to CO\(_2\) doubling in Goddard Institute for Space Studies model (Rind et al. 1995). Using the Met Office (UKMO) model, Ingram et al. (1989) estimated the same climate feedback parameters (with and without sea ice response, respectively) to equal 0.77 and 0.95 W m\(^{-2}\) K\(^{-1}\), which would imply that the global temperature response to a CO\(_2\) doubling would be \(\approx 19\%\) lower in the absence of sea ice feedbacks. Some of these differences may be associated with real model differences (Boe et al. 2009; Andrews et al. 2012). However, some of the differences in reported climate feedback parameter values may be attributed to different methodological approaches, as will be discussed below.

Surface albedo feedback has been estimated to contribute by \(\approx 15\%\) to the global mean temperature response to atmospheric CO\(_2\) forcing (Graversen and Wang 2009; Holland et al. 2001). Multimodel estimates of mean surface albedo feedback do not appear to differ substantially between transient (0.36 \pm 0.09 W m\(^{-2}\) K\(^{-1}\); Winton 2006) and equilibrium 2×CO\(_2\) simulations (0.36 \pm 0.19 W m\(^{-2}\) K\(^{-1}\); Colman 2003). Flanner et al. (2011) estimates the total (snow and sea ice) albedo feedback over the Northern Hemisphere to range between 0.33 and 1.07 W m\(^{-2}\) K\(^{-1}\) (using observational data) and 0.25 \pm 17 W m\(^{-2}\) K\(^{-1}\) using the Coupled Model Intercomparison Project phase 3 (CMIP3) multimodel mean. Crook and Forster (2014) also suggest that the current generation of climate models tends to underestimate the surface albedo feedback.

The overall sea ice effect estimated by Rind et al. (1995) or Ingram et al. (1989) includes sea ice feedbacks (due to ice albedo and ice thickness changes) as well as any other feedbacks (e.g., due to water vapor or cloud cover changes) associated with sea ice changes. Thus, these estimates are not directly comparable with assessments of the surface albedo feedback, especially given that the latter also incorporates land albedo changes (mainly due to snow cover changes). Ice albedo feedback alone has been suggested to play an important role in the recent sea ice decline (Perovich et al. 2007). Hudson (2011), using observations and a radiation transfer model, concludes that due to ice–albedo feedback only, complete Arctic sea ice removal would result in a forcing of \(-0.7\) W m\(^{-2}\).

In this study, we consider a set of slab ocean simulations with wide range of atmospheric CO\(_2\) concentrations that allows us 1) to investigate the overall effect of sea ice response on the climate feedback parameter (and its longwave and shortwave components) and 2) to make a first-order approximation of the radiative forcing that would be exerted on the climate system due to the complete sea ice loss from the simulated control state. The latter is accomplished by extrapolating the achieved radiative forcing values to the point in which the sea ice area equals zero.

Assessing the strength of sea ice feedbacks in our model involves comparison with simulations in which there is no sea ice response and thus no sea ice feedbacks or any associated feedbacks arising from the changes in sea ice cover (i.e., cloud or water vapor changes). Thus, we estimate the overall effect that the change in sea ice cover exerts on the climate system. Because the laws of physics manifested in the real world and represented in climate models do have sea ice feedbacks, assessing the strength of sea ice feedbacks involves comparison with model configurations in which the laws of physics are violated. We refer to our standard model configuration employing the thermodynamic–dynamic sea ice model as the “active ice” treatment. We use two methods of eliminating sea ice feedbacks, with the model physics altered in two different ways. Confidence in our results is increased by the fact that these two different approaches of eliminating sea ice feedbacks yield similar results. As described below, in our “zero ice” treatment we allow the ocean temperatures to supercool, while in our “prescribed ice” treatment we do not allow sea ice thickness or extent to respond to climate change. In all of the high-CO\(_2\) cases that we consider, the active-ice simulations have more ice than the zero-ice simulations and less ice than the prescribed-ice simulations. Thus, sea ice extents in the high-CO\(_2\) simulations with these two sea ice treatments bracket sea ice extents in the standard active-ice simulations. In the context of the aims of this paper, analysis of the zero-ice simulations requires fewer assumptions than does analysis of the prescribed-ice simulations. Therefore, we treat the zero-ice simulations in greater detail, and provide the prescribed-ice treatment as confirmatory evidence.

2. Methods

The model setup applied in this study is identical to the one used in Cvijanovic and Caldeira (2014, manuscript submitted to Climate Dyn.). Simulations are obtained using the National Center for Atmospheric Research’s Community Earth System Model (CESM) version 1.0.4 (Gent et al. 2011). This configuration incorporates the Community Atmosphere Model version 4 (CAM4; Neale et al. 2013), the Community Land Model version 4 (CLM4; Lawrence et al. 2011), and the Los Alamos Sea Ice Model version 4 (CICE4; Hunke and Lipscomb 2008) coupled to a slab (mixed layer) ocean. Prescribed ocean heat flux (\(q_\text{flux}\)) and other fields used by the slab ocean model component have been derived from the preindustrial CESM simulation

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using the full-depth ocean. Since the ocean heat flux is fixed in the slab ocean configuration, ocean heat transport cannot respond to the imposed forcing. Atmospheric and land model utilize 1.9° × 2.5° (latitude × longitude) finite volume grid. There are 26 atmospheric levels in the vertical. The ice and ocean models employ a 1° displaced pole grid (gx1v6).

The importance of sea ice response for climate sensitivity is assessed by performing three sets of simulations, with each set having a different ice treatment. In the first set of simulations, we use a standard model configuration including a dynamic–thermodynamic sea ice model (CICE4), which we refer to as the active-ice treatment. In a second set of simulations, namely the zero-ice treatment, the ocean temperatures are allowed to fall below the freezing point, allowing the ocean points to supercool and thus suppress any ice formation. Although in this setup sea surface temperature can fall below the freezing point, they are also influenced by the prescription of ocean heat fluxes. In the third set of simulations (prescribed ice), ice extent and thickness are prescribed and not affected by climate change. Twelve monthly ice extents used in the prescribed sea ice simulations have been derived from the 1×CO2 active-ice slab ocean simulation while the ice thickness is set uniformly to 1 m in the Northern Hemisphere and 2 m in the Southern Hemisphere. Under this sea ice treatment, any change in sea ice thickness due to melting or growing processes is being reset at the end of the time step. The energy removed or added to the system as a consequence of maintaining the sea ice thickness is assessed from surface energy balance considerations. Since changes in sea ice thickness are not permitted, we do not consider the effect of seasonal sea ice thickness variations that are also expected to affect the climate system (Rind et al. 1995, 1997; Holland et al. 2006).

A prescribed-ice setup, similar to the one used in our study, has been previously employed by Rind et al. (1995). In their setup, however, the energy needed to maintain the constant sea ice cover was returned back to the mixed layer. Zero- and prescribed-ice setups are described in greater detail in Cvijanovic and Caldeira (2014, manuscript submitted to Climate Dyn.).

Active, zero, and prescribed sea ice simulation anomalies are reported relative to the control (1×CO2) simulations obtained with the corresponding sea ice treatment. For example, in our analyses of the 4×CO2 simulations, we compare the difference between the 4×CO2 and 1×CO2 zero-ice simulations with the difference between the 4×CO2 and 1×CO2 active-ice simulations. These 1×CO2 control simulations vary among sea ice treatments: for example, global annual mean temperature in prescribed-ice 1×CO2 simulation is ~0.5 K larger than in the active-ice simulation (mainly as a consequence of different ice thicknesses). In comparison, zero-ice 1×CO2 global annual mean temperature is ~2 K warmer than in the active-ice simulation. The zero-ice control simulation (relative to the active-ice control simulation) has substantially warmer high latitudes during the winter [December–February (DJF)] season, with cooling that is more pronounced over the continents than over the ocean. Winter surface temperatures over the northern high latitudes (>60°N) are on average ~13 K greater in the zero-ice control simulation relative to the active-ice control simulation. In other seasons, average northern high altitude surface temperatures are still larger in the zero-ice than in the active-ice control simulation, but these differences are not as substantial [e.g., <1 K greater in June–August (JJA)]. With higher CO2 concentrations, difference between zero- and active-ice simulations becomes smaller.

When analyzing climatological mean model results, we discard the first 40 years of each simulation, which contains the adjustment to a new equilibrium state, and consider the last 60 years in our analysis. Solar insolation and orbital parameters are set to preindustrial values, and CO2 forcing is applied instantaneously at 2, 4, 6, or 8 times the value of the control simulation. In total, there are 15 simulations performed (1×CO2, 2×CO2, 4×CO2, 6×CO2, and 8×CO2 simulations for each of the active-, zero-, and prescribed-ice treatments). In addition to this, for each high-CO2 simulation (i.e., CO2 concentration greater than in the 1×CO2 simulations), we have performed two additional 10-yr simulations with slightly altered initial conditions. This is done in order to improve the statistical power of our linear regressions by obtaining three-member ensembles representing each treatment.

The slab ocean model configuration employed here does not allow for changes in ocean heat transport. Changes in ocean heat transport are likely to further shape the response of the climate system to sea ice loss (Polyakov 2005; Mahlstein and Knutti 2011; Stroeve et al. 2012). Use of the slab ocean versus the full-depth ocean model in estimating the climate sensitivity has been discussed by Danabasoglu and Gent (2009). Their study showed that the slab ocean model configuration provides a good estimate for the equilibrium climate sensitivity of the fully coupled Community Climate System Model, version 3 (CCSM3).

3. Results
a. Active-ice simulations

We use the term “achieved steady-state temperature change” to refer to the global annual mean warming
achieved in our simulations after removing the first 40 simulated years of transient warming. Basic global mean values for our climate model simulations are presented in Table 1. (All reported error ranges represent 1 standard error.) Global mean temperatures and sea ice areas are shown in Fig. 1.

Achieved steady-state temperature change in the 2×CO₂, 4×CO₂, 6×CO₂, and 8×CO₂ active-ice simulations relative to the 1×CO₂ equals 3.18, 6.59, 8.75, and 10.38 K (Table 1; Fig. 1). Resulting 2×CO₂ warming is in agreement with the CAM4 slab ocean experiments by Kay et al. (2012), who report a surface temperature response to CO₂ doubling of 3.2 K.

Gregory et al. (2004) made a major innovation when they recognized that valuable information about radiative forcing, climate sensitivity, and equilibrium temperature change could be gleaned by performing linear regressions on transient climate results for simulations involving step-function changes in radiative forcing. Here, we apply a slightly modified form of their analysis to our simulations. The regressions shown in Fig. 2 use the following equation:

$$N = RF_{CO₂} - \lambda \Delta T,$$

where N is the top-of-atmosphere energy imbalance (W m⁻²), RF⁻CO₂ (W m⁻²) is the imposed radiative forcing due to increased CO₂ concentration, \(\lambda\) (W m⁻²K⁻¹) is the climate feedback parameter, and \(\Delta T\) (K) is the change in temperature relative to the 1×CO₂ control simulation.

For each of our cases with increased atmospheric CO₂ concentration, we subtract the steady-state mean values from the corresponding 1×CO₂ simulation with the corresponding sea ice treatment. We then perform a geometric-mean linear regression, also known as reduced major-axis regression (Isobe et al. 1990) on the change in top-of-atmosphere (TOA) energy balance versus global mean temperature. These curve fits are shown in Fig. 2, with key results summarized in Fig. 3 and Table 2. Applying this method to the CO₂ doubling active-ice simulations, we find the climate feedback parameter \(\lambda_{active}\) to equal 1.10 ± 0.09 W m⁻² K⁻¹. This is in agreement with the previously reported value of 1.11 W m⁻² K⁻¹ by Kay et al. (2012), also derived from a CO₂ doubling in CAM4 slab ocean simulations. Taking into account all active-ice simulations, we estimate the mean climate feedback parameter for our high-CO₂ simulations to be 1.05 ± 0.05 W m⁻² K⁻¹.

We use the term “equilibrium temperature change” to refer to the temperature that achieves a net zero TOA energy balance in the active-ice regressions shown in Fig. 2, (i.e., the horizontal-axis intercept in a “Gregory regression”). In the active-ice cases, the equilibrium temperature change values are similar to the changes in achieved steady-state temperature change (Tables 1 and 2).

Gregory regressions may be performed independently on net upward longwave and shortwave fluxes at the top of the model. Using the same methods as applied in

<table>
<thead>
<tr>
<th>CO₂ level</th>
<th>Active ice</th>
<th>Zero ice</th>
<th>Prescribed ice</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Achieved steady-state temperature (K)</td>
<td>Sea ice area (10¹² m²)</td>
<td>Achieved steady-state temperature (K)</td>
</tr>
<tr>
<td>1</td>
<td>286.37 ± 0.01</td>
<td>28.21 ± 0.06</td>
<td>288.69 ± 0.01</td>
</tr>
<tr>
<td>2</td>
<td>289.55 ± 0.01</td>
<td>18.52 ± 0.05</td>
<td>290.96 ± 0.01</td>
</tr>
<tr>
<td>4</td>
<td>292.96 ± 0.01</td>
<td>8.81 ± 0.06</td>
<td>293.56 ± 0.01</td>
</tr>
<tr>
<td>6</td>
<td>295.12 ± 0.01</td>
<td>3.35 ± 0.08</td>
<td>295.35 ± 0.01</td>
</tr>
<tr>
<td>8</td>
<td>296.75 ± 0.01</td>
<td>1.05 ± 0.02</td>
<td>296.83 ± 0.01</td>
</tr>
</tbody>
</table>

FIG. 1. Achieved steady-state global mean temperature and sea ice area for three different sea ice treatments at five different atmospheric CO₂ concentrations. The active-ice case is similar to the prescribed-ice case at 1×CO₂ and similar to the zero-ice case at 8×CO₂. Note that the zero-ice case allows the ocean to cool below the freezing point without ice formation; the prescribed-ice cases remove energy as necessary to maintain sea ice thickness and extent. Between 1×CO₂ and 4×CO₂, the relationship between sea ice area and sea ice radiative forcing is quite linear. Line segments between data points are meant to guide the eye only.
estimating $\lambda_{\text{active}}$, we estimate contribution from the longwave component to be $\lambda_{L,\text{prescribed}} = 1.76 \pm 0.02 \text{W m}^{-2} \text{K}^{-1}$, and the contribution from the shortwave component to be $\lambda_{SW,\text{prescribed}} = -0.85 \pm 0.08 \text{W m}^{-2} \text{K}^{-1}$ (Table 3). Thus, as expected, longwave feedbacks tend to damp climate change whereas shortwave feedbacks tend to amplify climate change. These values are similar to the longwave and shortwave feedback parameters of 1.90 and $2 \pm 0.80 \text{W m}^{-2} \text{K}^{-1}$ reported by Kay et al. (2012) for CAM4 slab ocean simulations (numbers reported using the sign convention adopted by this study).

A straight line fit (using an ordinary least squares regression) through all active-ice simulations data points shown in Fig. 1 yields $\frac{\partial A_{\text{ice}}}{\partial T}$ of $2.67 \pm 0.14 \times 10^{12} \text{m}^{-2} \text{K}^{-1}$, where $A_{\text{ice}}$ refers to ice area and $T$ denotes achieved steady-state global mean temperature. It can be seen from Fig. 1 that nonlinear effects become more prominent in the $6\times\text{CO}_2$ and $8\times\text{CO}_2$ simulations; using only the $1\times\text{CO}_2$ through $4\times\text{CO}_2$ data points, $\frac{\partial A_{\text{ice}}}{\partial T}$ becomes $2.94 \pm 0.06 \times 10^{12} \text{W m}^{-2} \text{K}^{-1}$. Thus, our results suggest that in this model $\sim 3 \times 10^{12} \text{m}^2$ of sea ice are lost for each kelvin of warming. This is similar to the values estimated by Winton (2011) for Northern Hemispheric sea ice sensitivity to global temperature increase based on observations but is greater than the values estimated for the models he considered.

b. Zero-ice simulations

In the zero-ice simulations, achieved steady-state warming (relative to the $1\times\text{CO}_2$ zero-ice control simulation) equals 2.27, 4.87, 6.66, and 8.15 K for the $2\times\text{CO}_2$, $4\times\text{CO}_2$, $6\times\text{CO}_2$, and $8\times\text{CO}_2$ simulations, respectively. Thus, the zero-ice cases warm 28.7%, 26.1%, 23.9%, and 21.5% less than the corresponding active-ice cases for the $2\times\text{CO}_2$, $4\times\text{CO}_2$, $6\times\text{CO}_2$, and $8\times\text{CO}_2$ simulations. As in the active-ice simulations, the zero-ice equilibrium temperature change determined from the Gregory regressions is similar to the zero-ice achieved steady-state temperature change (Tables 1 and 2).

The mean climate feedback parameter in the zero-ice simulations $\lambda_{\text{zero}}$ is $1.31 \pm 0.03$. Comparison of $\lambda_{\text{zero}}$ and $\lambda_{\text{active}}$ indicates that sea ice response accounts for about 20% ± 4% of the climate sensitivity to an imposed change in radiative forcing. The contribution to the climate feedback parameter associated with sea ice response ($\lambda_{\text{ice}}$) can be estimated as the difference between the climate feedback parameters estimated from the zero-ice and active-ice simulations:

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**FIG. 2.** Gregory regressions of TOA energy imbalance vs global mean surface temperature, obtained using geometric mean regressions as described in the text. The vertical axis intercept at zero temperature change is the regressed radiative forcing. The horizontal axis intercept at zero TOA energy imbalance is the equilibrium temperature change. The slope of the regressed lines is an estimate of the climate feedback parameter. On average, the climate feedback parameter is $-20\%$ lower without sea ice feedbacks than with sea ice feedbacks.
component to be $\lambda_{SW,zero} = -0.64 \pm 0.06 \text{ W m}^{-2}\text{K}^{-1}$ (Table 3). If we subtract from these values with the corresponding values derived above for the active-ice simulations, we find that $\lambda_{SW,zero} - \lambda_{SW,active} = 0.21 \pm 0.03 \text{ W m}^{-2}\text{K}^{-1}$ and $\lambda_{LW,zero} - \lambda_{LW,active} = 0.09 \pm 0.02 \text{ W m}^{-2}\text{K}^{-1}$. Thus, we estimate that the change in longwave climate feedback parameter associated with sea ice response is nearly half as large as the corresponding change in shortwave climate feedback parameter. This suggests that longwave feedbacks must be considered when discussing effects of sea ice response on climate sensitivity.

c. Prescribed-ice simulations

Achieved warming in the prescribed-ice simulations (relative to $1\times CO_2$) equals 2.07, 4.29, 5.62, and 6.76 K for the $2\times CO_2$, $4\times CO_2$, $6\times CO_2$, and $8\times CO_2$ simulations respectively. This achieved warming is thus 35.3% ± 0.4% less than in the active-ice simulations for all of these $CO_2$ levels. Achieved warming in the prescribed-ice simulations is, however, not directly comparable with the achieved warming in the zero-ice simulations, because in the prescribed-ice treatment surface energy budget is altered in order to maintain the preindustrial sea ice extent. The global mean energy removal rate needed to maintain $1\times CO_2$ ice under higher-$CO_2$ conditions is shown in Table 1. In the prescribed-ice simulations, the equilibrium warming exceeds the achieved warming by an amount approximately equal to the energy removal rate divided by $\lambda_{prescribed}$. For example, the achieved warming in the $4\times CO_2$ prescribed-ice case is 291.1 K – 286.8 K = 4.3 K. The difference in energy removal rate is 1.13 W m$^{-2}$ and $\lambda_{prescribed}$ for this case is 1.32 W m$^{-2}$ K$^{-1}$, suggesting that this energy removal would cause a cooling of 0.86 K ($=1.13 \text{ W m}^{-2}/\lambda_{prescribed}$). Thus, a quadrupling of atmospheric $CO_2$ content in the absence of sea ice feedbacks and artificial energy loss would lead to an equilibrium warming of 4.3 K + 0.86 K = 5.16 K. This is in agreement with the assessed equilibrium temperature change from the Gregory regression of 5.17 ± 0.04 K (Table 2). Unlike the achieved warming, the equilibrium

![Graph showing climate feedback parameters](image)

**Table 2. Data for key values derived in Fig. 2 and shown in Fig. 3.**

<table>
<thead>
<tr>
<th>CO2 level</th>
<th>Active ice</th>
<th>Zero ice</th>
<th>Prescribed ice</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Climate feedback parameter (W m$^{-2}$K$^{-1}$)</td>
<td>Radiative forcing (W m$^{-2}$)</td>
<td>Equilibrium temperature change (K)</td>
</tr>
<tr>
<td>2</td>
<td>1.10 ± 0.09</td>
<td>3.42 ± 0.26</td>
<td>3.10 ± 0.02</td>
</tr>
<tr>
<td>4</td>
<td>1.02 ± 0.03</td>
<td>6.67 ± 0.20</td>
<td>6.51 ± 0.03</td>
</tr>
<tr>
<td>6</td>
<td>1.04 ± 0.02</td>
<td>9.02 ± 0.18</td>
<td>8.67 ± 0.03</td>
</tr>
<tr>
<td>8</td>
<td>1.05 ± 0.02</td>
<td>10.82 ± 0.20</td>
<td>10.30 ± 0.04</td>
</tr>
<tr>
<td>Mean</td>
<td>1.05 ± 0.05</td>
<td>8.92 ± 0.26</td>
<td>9.37 ± 0.02</td>
</tr>
</tbody>
</table>
warming in the prescribed-ice simulations is similar to the equilibrium warming in the zero-ice simulations. This suggests that climate sensitivity to the energy additions and subtractions needed to maintain sea ice is similar to the climate sensitivity to changes in atmospheric CO2 content.

The mean climate feedback parameter in the prescribed-ice simulations, $\lambda_{\text{prescribed}}$, equals 1.35 ± 0.09 W m$^{-2}$ K$^{-1}$, which does not differ significantly from the estimate for $\lambda_{\text{zero}}$ of 1.31 ± 0.03 W m$^{-2}$ K$^{-1}$. Comparison of $\lambda_{\text{prescribed}}$ and $\lambda_{\text{active}}$ indicates that sea ice response accounts for 22% ± 6% of the climate sensitivity to an imposed change in radiative forcing. We provide this analysis as evidence supporting conclusions obtained using the active-ice and zero-ice simulations.

d. “Effective sea ice radiative forcing” and an alternative approach to estimating the influence of sea ice response on the climate feedback parameter

1) USING RESULTS OF ACTIVE-ICE AND ZERO-ICE SIMULATIONS

We define “the effective radiative forcing from a change in sea ice area” using the active-ice and zero-ice simulations. For each CO2 level we take the difference in global mean temperature between the active-ice and zero-ice simulations (shown in Table 1) and multiply it by the mean climate feedback parameter for the zero-ice treatment ($\lambda_{\text{zero}}$ = 1.31 ± 0.03 W m$^{-2}$ K$^{-1}$)

$$RF_{\text{ice}} = \lambda_{\text{zero}}(T_{\text{active}} - T_{\text{zero}}),$$  \hspace{1cm} (3)

where $T_{\text{active}}$ and $T_{\text{zero}}$ are the achieved global mean temperature values from Table 1. The results of this calculation are plotted in Fig. 4. Standard errors for means calculated from the values that have their own uncertainties are determined considering both their estimated means and standard errors according to Eq. (5.38) in Headrick (2010) (assuming that each value represents the same number of samples).

In Eq. (3), we use $\lambda_{\text{zero}}$ and not $\lambda_{\text{active}}$, because in this analysis we consider sea ice as a forcing and not as a feedback (and $\lambda_{\text{active}}$ includes effects of sea ice response). The results of this calculation are plotted in Fig. 4. We fit straight lines to the calculated values shown in Fig. 4 and obtain regression slopes for the change in sea ice radiative forcing in the active-ice simulations relative to the zero-ice simulations of $\partial F_{\text{ice}}/\partial A_{\text{ice}} = 0.108 \pm 0.004 \times 10^{-12}$ W m$^{-2}$ for each square meter change in sea ice area. Above we showed that sea ice area decreases by $\partial A_{\text{ice}}/\partial T = 2.94 \pm 0.06 \times 10^{12}$ m$^2$ for each kelvin of global mean warming (for CO2 levels in the range from 1×CO2 to 4×CO2). The decrease in climate feedback parameter associated with sea ice response ($\lambda_{\text{ice}}$) can be considered as a product of the change in radiative forcing per unit change in sea ice area ($\partial F_{\text{ice}}/\partial A_{\text{ice}}$) and the change in sea ice area per unit change in temperature ($\partial A_{\text{ice}}/\partial T$):

$$\lambda_{\text{ice}} = (dF_{\text{ice}}/dA_{\text{ice}})(dA_{\text{ice}}/dT).$$ \hspace{1cm} (4)

![Fig. 4. Sea ice area relative to active-ice treatment (10^12 m^2).](image)

Table 3. Estimates of longwave and shortwave components of the climate feedback parameter, $\lambda$.

<table>
<thead>
<tr>
<th>CO2 level (times the preindustrial)</th>
<th>Active ice $\lambda_{\text{LW}}$ (W m$^{-2}$ K$^{-1}$)</th>
<th>Active ice $\lambda_{\text{SW}}$ (W m$^{-2}$ K$^{-1}$)</th>
<th>Zero ice $\lambda_{\text{LW}}$ (W m$^{-2}$ K$^{-1}$)</th>
<th>Zero ice $\lambda_{\text{SW}}$ (W m$^{-2}$ K$^{-1}$)</th>
<th>Zero ice minus active ice $\Delta\lambda_{\text{LW}}$ (W m$^{-2}$ K$^{-1}$)</th>
<th>Zero ice minus active ice $\Delta\lambda_{\text{SW}}$ (W m$^{-2}$ K$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>$1.77 \pm 0.04$</td>
<td>$-1.07 \pm 0.07$</td>
<td>$1.91 \pm 0.03$</td>
<td>$-0.79 \pm 0.05$</td>
<td>$0.13 \pm 0.05$</td>
<td>$0.28 \pm 0.08$</td>
</tr>
<tr>
<td>4</td>
<td>$1.81 \pm 0.01$</td>
<td>$-0.88 \pm 0.03$</td>
<td>$1.87 \pm 0.02$</td>
<td>$-0.64 \pm 0.03$</td>
<td>$0.07 \pm 0.02$</td>
<td>$0.24 \pm 0.04$</td>
</tr>
<tr>
<td>6</td>
<td>$1.73 \pm 0.01$</td>
<td>$-0.74 \pm 0.02$</td>
<td>$1.84 \pm 0.01$</td>
<td>$-0.56 \pm 0.02$</td>
<td>$0.10 \pm 0.02$</td>
<td>$0.18 \pm 0.03$</td>
</tr>
<tr>
<td>8</td>
<td>$1.72 \pm 0.01$</td>
<td>$-0.71 \pm 0.02$</td>
<td>$1.79 \pm 0.01$</td>
<td>$-0.55 \pm 0.02$</td>
<td>$0.07 \pm 0.02$</td>
<td>$0.16 \pm 0.02$</td>
</tr>
<tr>
<td>Mean</td>
<td>$1.76 \pm 0.02$</td>
<td>$-0.85 \pm 0.08$</td>
<td>$1.85 \pm 0.02$</td>
<td>$-0.64 \pm 0.06$</td>
<td>$0.09 \pm 0.02$</td>
<td>$0.21 \pm 0.03$</td>
</tr>
</tbody>
</table>
This approach parallels that of Winton (2006) wherein the surface albedo was partitioned into two components: change of TOA shortwave with surface albedo and change in surface albedo with near-surface temperature. Equation (4) suggests that sea ice feedbacks contribute approximately $\lambda_{\text{ice}} = 0.32 \pm 0.01 \text{ W m}^{-2} \text{ of sea ice radiative forcing for each kelvin of global mean warming. This compares with the value } \lambda_{\text{ice}} = 0.26 \pm 0.06 \text{ W m}^{-2} \text{ K}^{-1} \text{ estimated above using Eq. (2). The consistency of results obtained by these two different methods indicates the robustness of this analysis for the simulations considered here.}

2) USING RESULTS OF ACTIVE-ICE AND PRESCRIBED-ICE SIMULATIONS

For the prescribed-ice simulations, we modify Eq. (3) to account for the energy removed in order to maintain the sea ice and define the effective radiative forcing from a change in sea ice area $RF_{\text{ice}}$ for each CO2 level in the following way:

$$RF_{\text{ice}} = \lambda_{\text{prescribed}}(T_{\text{active}} - T_{\text{prescribed}}) - F_{\text{ice-maintenance-energy}}, \tag{5}$$

where $T_{\text{active}}$ and $T_{\text{prescribed}}$ are the achieved global mean temperature values from Table 1, and $F_{\text{ice-maintenance-energy}}$ is the energy needed to maintain sea ice under high CO2 conditions in the prescribed-ice simulations (Table 1). Equation (5) is founded on the concept that the temperature difference between the active-ice and prescribed-ice simulations is caused by the sum of the sea ice radiative forcing ($RF_{\text{ice}}$) and the energy removed to maintain sea ice in the prescribed-ice simulations. The results of this calculation are plotted in Fig. 4. We fit straight lines to the values shown in Fig. 4 and obtain regression slopes for the change in sea ice radiative forcing in the active-relative to the prescribed-ice simulations of $0.104 \pm 0.003 \times 10^{-12} \text{ W m}^{-2}$ for each square meter change in sea ice area. This is very similar to the $0.108 \pm 0.004 \times 10^{-12} \text{ W m}^{-2}$ estimated from the zero-ice simulations.

4. Discussion

As described in the introduction, many previous studies have focused on different aspects of the role of ice and snow in modulating the response of the climate system to increased greenhouse gas forcing (Ingram et al. 1989; Rind et al. 1995; Hall 2004; Graversen and Wang 2009; Hudson 2011). Our study estimates the total impact of sea ice response on climate sensitivity, thus including all the feedbacks arising from sea ice cover changes (such sea ice albedo and insulation feedbacks) as well as any other feedbacks arising from sea ice-induced changes in other climate system components. This effect thus includes more than just shortwave fluxes connected with albedo changes. Our simulations are designed to evaluate the overall effect of sea ice response on the climate feedback parameter and were not designed to analyze the strength of individual components (ice albedo feedback, ice thickness changes, or sea ice–induced cloud cover and water vapor changes).

Based on our active-ice and zero-ice simulations, we find that ~20% of the temperature response to a change in radiative forcing in the active-ice configuration of CESM can be associated with sea ice response. We find a substantial influence of sea ice response on the longwave climate feedback parameter, indicating that the overall impact of sea ice response on the climate feedback parameter should not be estimated based on shortwave feedbacks alone. Sea ice response can affect longwave climate feedbacks in several ways. For example, the presence or absence of sea ice affects the surface temperatures and vertical stability of the overlying atmosphere, thus affecting the longwave feedbacks (Boé et al. 2009; Pavelsky et al. 2011; Pithan and Mauritsen 2014). Nonlocal effects may also be important (cf. Rind et al. 1995).

In the prescribed-ice simulations (unlike in the zero- and active-ice simulations) the achieved warming and the “equilibrium temperature change” are not the same (a consequence of energy withdrawal needed to maintain the sea ice in prescribed-ice configuration). As a result, the achieved warming in the prescribed-ice simulations is actually 35.3% ± 0.4% less than in the active-ice simulations (at any CO2 level). This estimate is very close to the previously reported value of 37% by Rind et al. (1995) obtained with a method similar to our prescribed-ice setup. The achieved warming in our prescribed-ice simulations could be viewed in terms of an idealized scenario in which the underlying ocean takes up the energy surplus needed to maintain the sea ice cover. However, eventually this ocean heat uptake would need to be released back to the atmosphere, increasing the global mean temperature toward what was defined here as an equilibrium temperature.

Our study is based on the analysis of a slab-ocean model configuration with specified ocean heat transport. In the real world, and in dynamically coupled atmosphere–ocean models, changes in ocean heat transport could affect the climate feedback parameter. Ocean circulation and heat transport changes are shown to affect the sea ice distribution (Polyakov 2005; Mahlstein and Knutti 2011). Moreover, sea ice–ocean interactions can influence the deep water formation (Southern Ocean mainly; Goosse and Fichefet 1999) while sea ice...
parameterization (ice dynamics and ice thickness) can affect the sensitivity of oceanic response to climate change, alter the ocean heat transport, and impact the variability of the ocean thermohaline circulation (Holland et al. 2001; Bitz et al. 2001, 2006). Finally, a dynamically coupled atmosphere–ocean model can exhibit long-term responses that differ from those of an atmosphere–slab-ocean model (Gregory et al. 2004); therefore, caution should be exercised when interpreting results from slab-ocean model simulations. We thus encourage further studies that would incorporate the effects of ocean circulation changes.

Analysis of Coupled Model Intercomparison Project (CMIP) models shows that climate feedback parameter and its longwave and shortwave components can vary substantially among climate models (e.g., Winton 2006; Boé et al. 2009; Andrews et al. 2012); thus, results from any one model may not be a reliable indicator of real world feedback strengths. The multimodel study by Holland and Bitz (2003) found that the ratios of Arctic to global warming vary in the range from 1.5 to 4.5, indicating that the overall effect of sea ice loss on climate sensitivity may differ substantially among models. Moreover, modeled sea ice responses appear not to be sufficiently sensitive to temperature changes (Stroeve et al. 2012; Winton 2011). Thus, the approaches similar to the one described here (that allow isolation of effects of sea ice changes on climate sensitivity) could provide further insights into the causes of intermodel spread in overall climate response to global warming.

5. Conclusions

Using the Gregory analysis approach (Gregory et al. 2004), we have estimated the climate feedback parameters from the transient part of our model simulations with active, zero, and prescribed sea ice. We also show that we can estimate the climate feedback parameter for the standard active-ice configuration of a climate model by analyzing results from zero-ice (or prescribed ice) simulations and differences in ice areas and steady-state temperatures between the zero-ice (or prescribed ice) simulations and the active-ice simulations. This analysis shows that the contribution of sea ice response to the climate feedback parameter can be considered the sum of two factors, one dependent on sea ice and one independent of sea ice.

In our model, nearly one-third of the change in the climate feedback parameter caused by sea ice response is due to the longwave component of this feedback parameter. Furthermore the contribution of sea ice response to the climate feedback parameter can be decomposed into two terms, one related to changes in sea ice area with increasing temperature and another related to changes in sea ice radiative forcing per unit change in sea ice area.

Results obtained here indicate that in this configuration of CESM (CAM4 coupled to a slab ocean and the dynamic-thermodynamic sea ice model CICE4), approximately $3 \times 10^{12}$ m$^2$ of sea ice is lost for each kelvin of global mean warming and approximately 0.1 W m$^{-2}$ of “sea ice radiative forcing” is produced by each $10^{12}$ m$^2$ of sea ice loss, yielding a value of $-0.3$ W m$^{-2}$ K$^{-1}$ for the sea ice contribution to the overall climate feedback parameter. Because sea ice area in the $1 \times$ CO$_2$ control simulation is approximately $30 \times 10^{12}$ m$^2$, this suggests that complete loss of all sea ice from the $1 \times$ CO$_2$ state would produce a radiative forcing of about $3$ W m$^{-2}$, which is somewhat less than, but of the same order of magnitude as, the regressed radiative forcing from a doubling of atmospheric CO$_2$.

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