Isolating the Atmospheric Circulation Response to Arctic Sea Ice Loss in the Coupled Climate System

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ABSTRACT

In this study, coupled ocean–atmosphere–land–sea ice Earth system model (ESM) simulations driven separately by sea ice albedo reduction and by projected greenhouse-dominated radiative forcing are combined to cleanly isolate the sea ice loss response of the atmospheric circulation. A pattern scaling approach is proposed in which the local multidecadal mean atmospheric response is assumed to be separately proportional to the total sea ice loss and to the total low-latitude ocean surface warming. The proposed approach estimates the response to Arctic sea ice loss with low-latitude ocean temperatures fixed and vice versa. The sea ice response includes a high northern latitude easterly zonal wind response, an equatorward shift of the eddy-driven jet, a weakening of the stratospheric polar vortex, an anticyclonic sea level pressure anomaly over coastal Eurasia, a cyclonic sea level pressure anomaly over the North Pacific, and increased wintertime precipitation over the west coast of North America. Many of these responses are opposed by the response to low-latitude surface warming with sea ice fixed. However, both sea ice loss and low-latitude surface warming act in concert to reduce subseasonal temperature variability throughout the middle and high latitudes. The responses are similar in two related versions of the National Center for Atmospheric Research Earth system models, apart from the stratospheric polar vortex response. Evidence is presented that internal variability can easily contaminate the estimates if not enough independent climate states are used to construct them.

1. Introduction

The rapid retreat of Arctic sea ice (Stroeve et al. 2012) has motivated a number of studies examining how sea ice loss in isolation might impact the atmospheric general circulation. Model simulations can be used to address this fundamental research question in light of the short observational record, internal climate variability, and the difficulty of isolating sea ice variability from other processes (e.g., Magnusdottir et al. 2004; Deser et al. 2004; Chiang and Bitz 2005). In modeling studies, the most robust and intuitive response to sea ice loss is local warming and increased turbulent heat and moisture flux from the ocean into the atmosphere (e.g., Deser et al. 2010; Screen et al. 2013) driving strong Arctic amplification (Holland and Bitz 2003; Screen and Simmonds 2010; Screen et al. 2012). Many studies find a decrease in strength and an equatorward shift in the jet stream in response to reduced sea ice (e.g., Peings and Magnusdottir 2014; Deser et al. 2015, 2016). Other less robust circulation responses include a high pressure response over northern Eurasia that causes regional dynamical cooling in East Asia (e.g., Honda et al. 2009; Liu et al. 2012; Mori et al. 2014) and a weakening of the stratospheric polar vortex (e.g., Peings and Magnusdottir 2014; Kim et al. 2014). Beyond impacts on the mean circulation, a decrease in temperature variability on subseasonal time scales due to the reduced meridional temperature gradient and increased maritime influence is also found (Screen et al. 2015; Blackport and Kushner 2016), although some measures of subseasonal variability are found to increase in some studies (e.g., Peings and Magnusdottir 2014).

Some of the more recent cited studies have shown that sea ice loss in models with a full dynamical ocean component contributes to a moderate global warming signal (Deser et al. 2015, 2016; Blackport and Kushner 2016). Some aspects of the signal can be attributed to thermodynamic and dynamical changes in the ocean (Deser et al. 2016; Tomas et al. 2016) through experiments with prescribed sea surface temperatures and a slab ocean model that can be compared to simulations with a full ocean model. These studies show that the global warming signal is directly induced by polar warming, is
The purpose of the cited sea ice loss experiments is to isolate changes associated with sea ice loss in observed and projected climate change. Sea ice loss, under standard forcing scenarios [e.g., from phase 5 of the Coupled Model Intercomparison Project (CMIP5)], occurs concurrently with global warming, lower tropospheric polar amplification that reduces lower tropospheric temperature gradients, and upper tropospheric tropical amplification that increases upper tropospheric temperature gradients. These different temperature gradient changes have separable influences on storm track responses in the models contributing to CMIP5 (Harvey et al. 2014). Other work reveals a “tug of war” between the warming in the tropics and warming at the poles (Barnes and Polvani 2015): CMIP5 models that have more Arctic amplification tend to have a weaker and more equatorward jet response than models that have less Arctic amplification associated with sea ice loss. This is consistent with results from a simple general circulation model in which upper tropospheric and lower tropospheric temperature gradients can be separately controlled (Butler et al. 2010). These studies show that even though sea ice loss can impact the circulation at lower latitudes, this effect can be masked or counteracted by the effects of low-latitude warming (Barnes and Screen 2015).

In this study we ask whether it is possible to separate the direct influence of sea ice loss from low-latitude ocean warming using information from available coupled greenhouse-dominated radiative forcing simulations and sea ice forcing simulations. We probe this question with two sets of climate model experiments, each performed with two coupled Earth system models from the National Center for Atmospheric Research (NCAR; section 2). The first set of experiments is forced with historical and projected anthropogenic and natural forcing, we use the total sea ice area (SIA) and the spatial mean of low-latitude sea surface temperatures (SSTs). We hypothesize that we can use information from both types of experiments to separately attribute the pattern of atmospheric circulation response to sea ice loss without low-latitude surface warming and to low-latitude surface warming without sea ice loss (section 4). We demonstrate the consequences of this hypothesis (section 5) by decomposing the response under RCP8.5 forcing separately to the two processes. We show that potentially useful information can be produced when combining information from equilibrated time slice experiments (as in the sea ice loss simulations) and transient climate forcing experiments (RCP8.5), provided sufficient data are available to average out internal variability. Follow-on ideas for generalizing this approach and more carefully testing the hypotheses are presented in section 6. The appendix provides additional background on the response in the different Earth system model simulations.

2. Models and experiments

2.1. CESM1 experiments

We focus on two sets of experiments using the Community Earth System Model version 1 (CESM1), a coupled atmosphere–ocean–land–sea ice model that was developed at the National Center for Atmospheric Research. The atmospheric component is the Community Atmospheric Model version 5 (CAM5), which has 30 vertical levels up to about 3 hPa, while the ocean component is the Parallel Ocean Program (POP2), which has 60 vertical levels. The land and sea ice components are the Community Land Model version 4 (CLM4) and CICE4, respectively. All model components have approximately 1° horizontal resolution. More information about the different components of CESM1 can be found in Hurrell et al. (2013) and references therein.

To determine the atmospheric response to historical and projected anthropogenic and natural forcing, we use the CESM1 large ensemble (Kay et al. 2015). This is a 30-member initial condition ensemble that simulates the climate from 1920 to 2100. The 30 realizations are initialized by perturbing the temperature with numerical
round-off differences in air temperature on 1 January 1920 in a historical forcing simulation. These 30 realizations are then forced with historical greenhouse gas, aerosol, solar, and volcanic forcing, as well as prescriptions of land use change, until 2005, where they are then forced with the RCP 8.5 anthropogenic and natural forcings until 2100. Because each realization has the same forcing, any difference between the realizations is due to internal variability, which is averaged out in the ensemble mean provided a sufficiently large ensemble is used.

To isolate the climatic response to sea ice loss we follow the method used in Blackport and Kushner (2016). First, a 725-yr-long control simulation that has constant year 2000 radiative forcing is branched off one of the CESM1 large ensemble members at year 2000. After some adjustment as the planet continues to warm to its equilibrium state (see section 3), a sea ice albedo forcing simulation is branched off the control simulation at year 301 (nominal year 2300). As in Blackport and Kushner (2016), this is created by instantaneously altering three parameters in the sea ice code that change the albedo of the sea ice, allowing it to absorb more sunlight and rapidly melt. This sea ice albedo forcing simulation is run for a total of 525 years. The albedo change is applied to the whole sea ice field in both hemispheres, causing summertime sea ice loss in both the Arctic and Antarctic and warming in both hemispheres.

b. CCSM4 experiments

We will also use a similar set of experiments using the Community Climate System Model version 4 (CCSM4), an earlier version of the NCAR coupled climate model (Gent et al. 2011). The main difference from CESM1 is that CCSM4 uses the atmospheric component CAM4 instead of CAM5. The two atmosphere components have the same dynamical core, but very different physics parameterizations in the clouds, aerosols, radiation, and shallow convection. These differences lead to large differences in the Arctic in the two models. In particular, CESM1 has more Arctic amplification in response to a greenhouse gas increase as a result of differences in shortwave feedbacks due to optically thinner clouds in CAM5 than in CAM4 (Kay et al. 2012), and this drives significant differences in the sea ice loss in the two models (see section 3 and the appendix).

As for CESM1, we use sea ice albedo forcing and an RCP8.5 forcing experiment for CCSM4. The sea ice albedo forcing experiments are the same simulations used in Blackport and Kushner (2016) and are set up identically to the CESM1 experiments just described. They consist of a 780-yr-long year 2000 radiative forcing control simulation and an 800-yr-long sea ice albedo forcing simulation. The CCSM4 RCP8.5 forcing simulations are similar to the CESM1 experiments; however, only five realizations are available compared to 30 realizations for CESM1. This means that internal variability plays a larger role in the CCSM4 RCP8.5 results, as will be noted in section 5b.

For reference, in the appendix we briefly compare selected aspects of the long-term sea ice albedo forcing experiments for CESM1 and CCSM4, including the seasonal cycle of sea ice loss, warming, and surface energy budget changes. In the plots below, the significance of model responses is tested using the Student’s t test, assuming that each year is independent. This likely overestimates significance but is adequate to assess relative significance of the response in different fields. Statistical significance for the variability plots (Fig. 10) is calculated using an f-test for equal variance. The number of degrees of freedom used for the f-test is the total number of days divided by 10.

3. Sea ice and sea surface temperature response

The time series of the annual mean Arctic SIA for the CESM1 simulations are shown in Fig. 1a. The blue curve shows the ensemble mean SIA for the 30 realizations of the CESM1 historical and RCP8.5 forcing simulations from 1920 to 2100. Because much of the internal variability is averaged out, the SIA time series is a relatively smooth curve that represents the forced response. The annual mean SIA is relatively constant at about 11 million km$^2$ until toward the end of the twentieth century when it starts to decrease significantly to less than 4 million km$^2$. At year 2000 in one of the ensemble members, the year 2000 control simulation (red curve) is branched off and the sea ice continues to slowly melt, with SIA decreasing from about 10 to 9 million km$^2$ over about 200 years as the coupled system adjusts to equilibrium. The sea ice albedo forcing simulation (green curve) is branched off at year 301 of the year 2000 control. In the first year, over 2 million km$^2$ of sea ice melts; the ice continues to melt as SIA equilibrates at about 6 million km$^2$. The seasonal cycle of sea ice loss (see appendix) shows a sea ice–free Arctic (<0.5 million km$^2$) in August, September, and October, reflecting the large shortwave forcing and albedo feedbacks in the polar Arctic. Wintertime sea ice loss is more modest but the sea ice experiences thinning throughout the year.

The annual mean, Northern Hemisphere mean SST time series is shown in Fig. 1b. As expected, the SSTs warm very strongly in response to greenhouse gas–dominated radiative forcing in the historical and RCP8.5 forcing simulations. In addition, temperatures continue to increase by approximately 0.4°C in the year 2000 control simulation before equilibrating. In the sea ice
albedo forcing simulation, there is about 0.3°C warming immediately after the perturbation is imposed and then a slow warming as the system adjusts to a new equilibrium on a global scale. Figure 1c shows the SSTs averaged between 0° and 40°N in these three experiments to represent low-latitude Northern Hemisphere warming away from the Arctic. Even at these lower latitudes, sea ice albedo forcing drives a small warming of about 0.3°C that reflects the “mini” global warming (Deser et al. 2015) induced by a warmer ocean and greenhouse warming related to water vapor feedback. One task of our analysis is to try to separate out the impact of this additional warming from the sea ice loss response.

Figure 2 plots the Arctic SIA response against the low-latitude SSTs (averaged from 0° to 40°N) in each of the simulations for the annual mean, September–November (SON), and December–February (DJF). Each blue dot represents the ensemble mean for each year of the 30 transient CESM1 historical and RCP8.5 simulations, and the red and green dots represent the time mean of the year 2000 control simulation and sea ice albedo forcing simulations respectively. Only the last 425 years for the control and 275 years for the sea ice albedo forcing simulation were used, because this is when the simulations had reached equilibrium. In the annual mean (Fig. 2a), there is a linear relationship between the SIA and lower latitude SSTs: for every 1 million km² of sea ice loss there is an increase in SSTs at the lower latitudes of approximately 0.49°C. This linear relationship has also been seen between SIA and global temperatures in most climate models (Winton 2011; Mahlstein and Knutti 2012). The equilibrated year 2000 control simulation lies very close to the RCP8.5 forcing simulation in the early 2030s. This means that for CESM1 under RCP8.5 forcing, there is the equivalent of just over 30 years of SST warming and sea ice loss in the adjustment to equilibrium in the year 2000 control simulation. In the sea ice albedo forcing simulation there is less warming compared to the amount of sea ice loss, so it does not lie on the same line as the transient RCP 8.5 simulation. For every
1 million km$^2$ of sea ice loss, there is approximately 0.12°C warming in the SST at lower latitudes, or about 4 times less than in the RCP8.5 simulations.

The scatterplots of SIA versus the SSTs averaged from 0°–40°N for SON and DJF for the RCP8.5 do not show as straightforward a linear relationship. Toward the end of the twenty-first century during SON, there is less SIA decrease per increase in SST as result of the Arctic becoming ice free in these months. In DJF, the opposite happens: the sea ice loss per degree of low-latitude warming increases in the middle of the twenty-first century. This could be the result of the ice edge entering the Arctic Ocean at this time making more ice available to melt, as described by Eisenman (2010) to explain why the rate of Arctic ice loss is faster in summer than in winter. In other words, the reason for increased rate of sea ice loss in winter could be that the sea ice edge becomes more “summer like” once the ice edge retreats into the Arctic Ocean.

A summary of the changes in Arctic SIA and SST from 0° to 40°N for the annual mean and for DJF for each experiment is presented in Table 1. To easily compare the sea ice albedo forcing experiments with the RCP8.5 experiments, we have used 10-yr averages of the ensemble mean of epochs when total Arctic SIA matched as closely as possible their amounts in the year 2000 simulation and the sea ice albedo forcing simulation. These epochs are 2027–36 and 2063–72 for the annual mean and 2027–36 and 2057–66 for DJF. (For CCSM4, we used 20-yr averages to compensate for the relatively small number of transient RCP8.5 realizations. The epochs are 2030–49 and 2062–81 for the annual mean and 2032–51 and 2052–71 in DJF for CCSM4.) The difference between these two epochs defines the climate response in the RCP8.5 experiments. The reason that the separation between epochs is shorter for DJF than for the annual mean is that sea ice albedo forcing drives less sea ice loss in winter than in summer due to a lack of shortwave forcing in winter (Deser et al. 2015). The amount of low-latitude warming in the sea ice albedo forcing experiments in CESM1 is about 25% of that in the RCP8.5 forcing experiments in

| Table 1. The change in Northern Hemisphere SIA ($\delta$I) and SST averaged between 0° and 40°N ($\delta$T) and their ratios for the different experiments in either the annual mean or the DJF mean. |
|---------------------------------|-----------------|-----------------|
| Arctic SIA $\delta$I (10$^6$ km$^2$) | 0°–40°N SST $\delta$T (°C) | $\delta$T/$\delta$I |
| Annual CESM1 RCP8.5 forcing [(2063–72) – (2027–36)] | $-2.90$ | $1.44$ | $-0.495$ |
| Annual CESM1 sea ice albedo forcing | $-2.94$ | $0.358$ | $-0.122$ |
| DJF CESM1 RCP8.5 forcing [(2057–66) – (2027–36)] | $-2.36$ | $1.16$ | $-0.494$ |
| DJF CESM1 sea ice albedo forcing | $-2.28$ | $0.360$ | $-0.157$ |
| Annual CCSM4 RCP8.5 forcing [(2062–81) – (2030–49)] | $-2.16$ | $0.927$ | $-0.429$ |
| Annual CCSM4 sea ice albedo forcing | $-2.12$ | $0.221$ | $-0.104$ |
| DJF CCSM4 RCP8.5 forcing [(2052–71) – (2032–51)] | $-1.25$ | $0.558$ | $-0.447$ |
| DJF CCSM4 sea ice albedo forcing | $-1.17$ | $0.227$ | $-0.194$ |
the annual mean (about 30% in DJF). The CCSM4 results are similar, except that we see less ice loss in response to the same albedo forcing, especially during DJF. We see more low-latitude warming relative to the RCP 8.5 experiments than Deser et al. (2015) did, possibly a result of the Antarctic sea ice melting, which was not included in Deser et al. (2015), contributing to the global warming response, or a result of the different methods used to melt the sea ice.

Figure 3 shows the spatial maps of the response of the DJF sea ice concentration (SIC) and sea ice thickness (SIT) in both the RCP8.5 forcing experiments and the sea ice albedo forcing experiment and their differences. The responses in the RCP8.5 experiments are calculated using the epochs in Table 1. Despite the different forcing mechanisms, the two experiments show a similar spatial pattern of sea ice loss. For SIC, the region that shows that biggest difference between the two experiments is the Barents–Kara Sea region, where the RCP8.5 experiment has greater losses (about a 20% larger change in SIA averaged over 70°–80°N, 40°–80°E). The RCP8.5 experiment also has a larger decrease in SIC in the Bering Strait region and a smaller decrease in SIC throughout most of the rest of the Arctic Ocean. For SIT, both experiments show similar thinning of the sea ice throughout the entire Arctic Ocean, but the sea ice albedo forcing experiment generally shows more thinning (the differences are less than 0.15 m).

4. Using pattern scaling to isolate the response to sea ice loss

Figures 4a and 4b, which will be described in more detail in the next section, show the response of the annual mean and zonal mean temperature for the RCP8.5
and sea ice albedo forcing experiments, respectively. Figure 4a shows the temperature response as a difference for the RCP8.5 epochs whose net sea ice loss (i.e., reduction of sea ice area) is quite similar to that found in the sea ice albedo forcing experiment (section 2 and Table 1). A key point to notice is that the amplitude and structure of the lower tropospheric warming in the two simulations is quite similar. This suggests that aspects of the atmospheric response to sea ice loss might be controlled by the total amount of sea ice loss and might be less sensitive to details such as the spatial pattern of the sea ice loss (Fig. 3) and the cause of the sea ice perturbation. We also notice that there is moderate tropical upper tropospheric warming in the sea ice albedo forcing experiment, a feature also found more strongly in the RCP8.5 forcing experiment. This is part of the "mini" global warming induced by sea ice loss in the coupled climate system (Deser et al. 2015). This response is described classically as a response to low-latitude surface warming, which is stronger under greenhouse warming, and can be considered separately from the sea ice response. This motivates our approach of decomposing the total response due to the separate controls of sea ice loss and low-latitude surface ocean warming.

We propose a pattern scaling approach in which Arctic sea ice area ($I$) and low-latitude SST ($T$) are the controlling variables. Defining $Z$, which represents a field, such as 500-hPa geopotential height at a given Northern Hemisphere geographic location, that could be affected by both sea ice loss and low-latitude surface warming, we write $Z = Z(T, I)$. We consider how $Z$
varies in response to small variations in $T$ and $I$. Symbolically we write

$$
\frac{\partial Z}{\partial T} \bigg|_I + \frac{\partial Z}{\partial I} \bigg|_T \delta I \Delta T
$$

The first partial derivative on the right-hand side of the second equation represents the response of $Z$ to a change in low-latitude temperature while the sea ice area is held constant and the second partial derivative represents the response of $Z$ to a change in sea ice area while the low-latitude temperature is kept constant. To estimate these derivatives we use the output of the two sets of experiments, according to Eq. (1):

\[
\begin{align*}
\delta Z_A &= \frac{\partial Z}{\partial T} \bigg|_I \delta T_A + \frac{\partial Z}{\partial I} \bigg|_T \delta I_A, \\
\delta Z_R &= \frac{\partial Z}{\partial T} \bigg|_I \delta T_R + \frac{\partial Z}{\partial I} \bigg|_T \delta I_R.
\end{align*}
\]

Here the subscript $A$ indicates the response (long-term equilibrium difference from control) in the sea ice albedo forcing experiment and the subscript $R$ the response (ensemble mean epochal difference) in the RCP8.5 forcing experiment. We can diagnose all quantities in these equations except the two partial derivatives. Since we have two equations and two unknowns we are able to solve for these quantities by inversion:

\[
\begin{pmatrix}
\frac{\partial Z}{\partial T} \\
\frac{\partial Z}{\partial I}
\end{pmatrix}
= \frac{1}{\delta I_A \delta T_R - \delta I_R \delta T_A} \begin{pmatrix}
\delta I_R & \delta I_A \\
\delta T_R & -\delta T_A
\end{pmatrix}
\begin{pmatrix}
\delta Z_A \\
\delta Z_R
\end{pmatrix}.
\]

The two quantities are simply linear combinations of the responses in each of the two experiments, weighted by the amount of sea ice loss and low-latitude surface warming in each experiment. Note that the partial derivatives on the left-hand side of Eq. (4) and the quantities $\delta Z_R$ and $\delta Z_A$ on the right-hand side of Eq. (4) are spatial fields. The denominator and $2 \times 2$ matrix on the right-hand side of Eq. (4) are not spatial fields but simply the four numerical values obtained from the sea ice loss and low-latitude warming in the two experiments.

We note that the expressions above are somewhat ambiguous to interpret. For example, although we use the SSTs from 0°–40°N as the $T$ to perform this calculation, it does not necessarily mean that any part of the response that appear in this term is directly caused by the low-latitude warming. Instead, it means that that part of the response scales with the SSTs from 0° to 40°N in these experiments as distinct from the scaling with Arctic sea ice loss. Thus, the first term on the right-hand side of Eq. (1) can be thought of as the part of the response due to low-latitude surface warming not related to sea ice loss. Similarly, the second term on the right-hand side of Eq. (1) can be thought of as the part of the response due to sea ice loss that is not related to low-latitude surface warming. This is distinct from the result of the sea ice forcing experiments because of the warming induced at low latitudes in the coupled system.

For the CESM1 results, the values for $\delta Z_R$, $\delta I_R$, and $\delta T_R$ will be calculated using the epochal difference between two time periods used in Table 1, so that the simulations have similar amounts of sea ice loss with $\delta I_R \approx \delta I_A$. Defining $\varepsilon = \delta T_A / \delta T_R < 1$ as the ratio of the sea ice albedo forcing’s warming to the RCP8.5 forcing’s warming at this time (i.e., the magnitude of the RCP8.5 response is negligible), then

\[
\frac{\delta Z}{\delta T} \delta T_R = \frac{\delta Z_R - \varepsilon \delta Z_A}{1 - \varepsilon}, \quad \frac{\delta Z}{\delta I} \delta I_R = \frac{\delta Z_A - \varepsilon \delta Z_R}{1 - \varepsilon}.
\]

The first equality represents the diagnosed part of the RCP8.5 response associated with low-latitude temperature change and the second equality the response associated with sea ice loss. These expressions provide four useful special cases: if $\delta Z_R \ll \delta Z_A$ (i.e., the magnitude of the RCP8.5 response is negligible), then

\[
\frac{\partial Z}{\partial T} \delta T_R = \frac{\delta Z_A}{1 - \varepsilon}, \quad \frac{\partial Z}{\partial I} \delta I_R = \frac{\delta Z_A}{1 - \varepsilon}.
\]

This implies that if the RCP8.5 forcing response is negligible the low-latitude surface warming and sea ice loss parts of the decomposition are approximately equal and opposite and inflated by a factor of $1 / (1 - \varepsilon)$. If $\delta Z_A = \varepsilon \delta Z_R$, then

\[
\frac{\partial Z}{\partial T} \delta T_R = \delta Z_R, \quad \frac{\partial Z}{\partial I} \delta I_R = 0.
\]
\[
\frac{\partial Z}{\partial T} T_R = 0, \quad \frac{\partial Z}{\partial I} \delta I_R = \delta Z_A \tag{8}
\]

then all the RCP8.5 response can be attributed to sea ice loss, leaving nothing to explain from low-latitude surface warming. Finally, if \( \delta Z_R = -\delta Z_A \),

\[
\frac{\partial Z}{\partial T} T_R = -2\delta Z_A \quad \frac{\partial Z}{\partial I} \delta I_R = \frac{1 + \varepsilon}{1 - \varepsilon} \delta Z_A \tag{9}
\]

then both the sea ice and low-latitude surface warming parts of the decomposition have opposite sign and are inflated, with low-latitude surface warming part having a greater magnitude.

Throughout this study, we use total SIA area change \( \delta I \) and low-latitude warming \( \delta T \) for a given seasonal mean to diagnose the response in that mean (e.g., DJF mean values for \( \delta I \) and \( \delta T \), to diagnose DJF \( \delta Z \)). A process-based approach might introduce lags between the sea ice loss and the atmospheric response and test for regional effects. For example, to test whether it is important to explicitly account for how summertime sea ice loss can influence the wintertime atmosphere (e.g., Lainé et al. 2016), we have tried using June–August or September–November sea ice area to diagnose DJF responses and found that the results of the decomposition are very similar—for most variables and regions, only the magnitude of the calculated terms change (not shown). Similarly, a test of the impact of sea ice loss in the Barents and Kara Sea region, which has been identified as being important for the circulation response (e.g., Petoukhov and Semenov 2010), again yielded similar results apart from the change in magnitude for each of the terms in the decomposition (not shown).

While we have emphasized the DJF season, the suggestion that the circulation response to the sea ice loss peaks in late winter (Kim et al. 2014; Sun et al. 2015) motivates us to examine the February–March (FM) response. The results are similar but indeed point to the midlatitude RCP8.5 response being more dominated by sea ice loss to some extent (not shown). For example, the decrease in zonally averaged midtropospheric wind speed at high latitudes in the RCP8.5 response is about 0.5 m s\(^{-1}\) in FM compared to 0.3 m s\(^{-1}\) in DJF (Fig. 6). This suggests that in late winter sea ice may play a somewhat larger role than low-latitude warming in CESM1.

We have also considered whether the relationships between \( \delta I \) and \( \delta T \) with \( \delta Z \) might change with time. In the North Atlantic in the RCP8.5 simulations, weakening of the Atlantic meridional overturning circulation (AMOC) results in reduced warming there (not shown). In the sea ice albedo forced simulations, the AMOC first weakens and then recovers to nearly the strength in the control simulation [not shown, but similar behavior was seen in Fig. 1 of Blackport and Kushner (2016)]. For the analysis presented in the next section we use the long-term average after the AMOC has recovered to perform all of the calculations. If we use an earlier time period in the sea ice albedo forcing simulation, when the AMOC is still weak (years 50–200), we get nearly identical results with only small differences in maps over the North Atlantic. The results also do not depend strongly on the epochs chosen in the RCP8.5 experiment, with the exception of using epochs in the latter part of the twenty-first century for DJF. During these times, \( \varepsilon \) starts increasing (i.e., the amount of sea ice loss per unit low-latitude surface warming increases and becomes closer to what it is in the sea ice albedo forcing simulations), which makes the calculation less robust. In addition, the pattern of sea ice loss is quite distinct in the two experiments when epochs in the latter parts of the twenty-first century are used.

Finally, caution should be used in applying this decomposition to the stratosphere, where projected anthropogenic greenhouse gas increases cool the stratosphere through their direct radiative impacts. This process does not operate in the sea ice albedo forcing experiments with fixed greenhouse gas concentrations, even in the presence of the sea ice–induced low-latitude warming. The situation in the RCP8.5 simulations is further complicated by the recovery of stratospheric ozone in the models, which is projected to warm the stratosphere, although to a lesser extent than the greenhouse gas cooling (Forster et al. 2011).

5. Results of the decomposition

a. CESM1 results

In Figs. 4–14 we show the results of the decomposition described in section 4 for a number of different fields related to atmospheric circulation and variability in a standard format. For each field, we will show four panels as follows:

- Panel (a): The response (epochal difference for the epochs in Table 1) in the RCP8.5 forcing experiment.
- Panel (b): The response (long-term mean of sea ice albedo forcing minus the year 2000 control) in the sea ice albedo forcing experiment.
- Panel (c): the low-latitude surface warming-related response in the RCP8.5 experiment [corresponding to the first term on the right-hand side of Eq. (3)].
- Panel (d): the sea ice loss-related response in the RCP8.5 experiment [corresponding to the second term on the right-hand side of Eq. (3)].
By construction, in all figures panel (c) plus panel (d) equals panel (a). Note that (a) and (b) are determined directly from the model output in the two experiments, while (c) and (d) are calculated using the methods described in section 4 (note that no statistical significance test was developed for these derived quantities).

Figure 4 shows the decomposition for the annual mean, zonal mean temperature, using the annual mean \( \delta I \) and \( \delta T \). In the RCP 8.5 forcing experiment (Fig. 4a), there is warming everywhere in the troposphere with the largest increases in temperature being in the Arctic lower troposphere and in the tropical upper troposphere. In the sea ice albedo forcing experiment (Fig. 4b), there is a similar structure and magnitude of the warming signal in the Arctic planetary boundary layer, as well as additional warming in the Arctic lower troposphere. Since the epochs in the RCP8.5 were chosen to have similar sea ice loss to the sea ice albedo forcing response, the near-surface warming provides a good match between the two experiments. This is borne out in the decomposition in Figs. 4c and 4d. Figure 4c suggests that the low-latitude surface warming part (Fig. 4c) is responsible for the warming throughout the troposphere including the tropical warming, while the sea ice loss part (Fig. 4d) is responsible for all of the warming in the Arctic lower troposphere, as in the \( \delta Z_R = \delta Z_A \) case [see Eq. (8)], but none of the tropospheric warming outside this region, as in the \( \delta Z_A = \varepsilon \delta Z_R \) case [see Eq. (7)]. This clear separation, which is a consequence of annual mean Arctic lower tropospheric warming scaling with annual mean sea ice loss and the remaining signal scaling with low-latitude warming, gives us some confidence that we can use this method to decompose other fields.

The DJF zonal mean temperature (Fig. 5) decomposition is similar to the annual mean in the troposphere, with most of the Arctic amplified lower tropospheric warming attributable to sea ice loss. The decomposition in Fig. 5 is different from Fig. 4 in that the stratospheric responses are distinctive. As mentioned in section 4, the application of this decomposition in the stratosphere should be treated cautiously, but there is
potentially useful information in this analysis. In the RCP8.5 simulation (Fig. 5a), there is tropical stratospheric cooling (several degrees) and little change in the Arctic stratosphere, whereas in Fig. 5b there is polar stratospheric warming (about one degree) and little change at lower latitudes. Our tentative interpretation of this is that in RCP8.5 greenhouse gas increases have cooled the stratosphere from the tropics to the high latitudes, and that this is partially offset at the pole by stratospheric warming induced by sea ice loss (which has been seen in other simulations; e.g., Sun et al. 2015). In the decomposition, polar stratospheric warming that scales with sea ice loss (Fig. 5d) appears to cancel the polar stratospheric cooling that scales with low-latitude surface warming (Fig. 5c). This corresponds to the $\delta Z_R \ll \delta Z_A$ case [see Eq. (6)] where the RCP8.5 forcing response is negligible compared to the sea ice albedo forcing response. However, we question the validity of the result from the decomposition that the influence of sea ice loss warms the low-latitude stratosphere (a similar signal with weaker amplitude seen in Figs. 4c,d).

The DJF zonal mean wind response in the RCP 8.5 experiment (Fig. 6a) is very weak in the troposphere, with only a small (0.3 m s$^{-1}$) decrease on the poleward side of the eddy-driven jet between 60$^\circ$ and 80$^\circ$N. The sea ice albedo forcing experiment produces a weakening and southward shift in the jet consistent with other sea ice perturbation studies (e.g., Peings and Magnusdottir 2014; Deser et al. 2015) but somewhat inconsistent with our own previous study using CCSM4 (Blackport and Kushner 2016); this case will be discussed in the next subsection. Because the RCP8.5 experiment produces very little response and the sea ice albedo forcing experiments produce an equatorward shift of the eddy-driven jet, the low-latitude surface warming part of the decomposition consequently corresponds to a roughly equal and opposite poleward shift in the eddy-driven jet as shown in Fig. 6c [another case with $|\delta Z_R| \ll |\delta Z_A|$; see Eq. (6)]. The cancellation is consistent with Barnes and Polvani (2015), who show that the CMIP5 models that have more (less) Arctic amplification tend to have more of a decrease (increase) in the strength of the tropospheric jet and an equatorward (poleward) shift of the
Consistent with the stratospheric temperature response in Fig. 5, we find a decrease in the strength of the stratospheric polar vortex in the sea ice albedo forcing experiments and in the estimated response to sea ice loss. This result is in agreement with Peings and Magnusdottir (2014), who found a similar result forcing CAM5, the atmospheric model used in CESM1, with lower sea ice concentrations.

In summary, wintertime sea ice loss in isolation, with low-latitude surface warming suppressed, drives an equatorward shift of the tropospheric jet and a weakening of the stratospheric polar vortex associated with a warming of the polar stratosphere (Figs. 4d and 5d). The latter signal occurs despite the fact that CAM5 is a low-top model (Peings and Magnusdottir 2014; Kim et al. 2014; Sun et al. 2015). Within this framework it is difficult to say more about how this impact can be separated from other impacts of climate change in the polar stratosphere.

The results of the decomposition for midtropospheric extratropical warming and circulation as represented by Z500 in DJF are shown in Fig. 7. In the RCP 8.5 forcing experiment (Fig. 7a), there are increases in height throughout the entire Northern Hemisphere accompanying tropospheric greenhouse warming. The smallest increases are over the North Atlantic and northeast Pacific Oceans. In the sea ice albedo forcing experiments (Fig. 7b), there are large increases over Greenland, no increase over the North Atlantic and western Europe, and small decreases over the northeast Pacific. However, there are also small increases over the rest of the Northern Hemisphere as a result of the small amount of global warming. The decomposition in Fig. 7d removes the small global warming signal from the sea ice loss signal. What remains is the increase over Greenland, a larger decrease over the North Pacific, and a small decrease over western Europe. Compared to the total
RCP8.5 response (Fig. 7a), in the low-latitude surface warming part of the response (Fig. 7c) there are larger increase over the North Pacific and over Eurasia, with little change over the North Atlantic. This suggests that the sea ice loss is responsible for the decrease in geopotential heights in the North Pacific, and that the low-latitude surface warming part of the response increases heights in this region, cancelling the response to sea ice loss. It also suggests that both sea ice loss and the low-latitude surface warming are contributing to the relatively small changes in Z500 over the North Atlantic sector.

Figure 8 shows the sea level pressure (SLP) response in the two experiments and the decomposition into the parts of the response. In the RCP8.5 forcing experiment, there is a decrease in SLP over most of the Arctic Ocean with the largest response over the Barents Sea and the Chukchi Sea, as well as a weak decrease over North America. In the sea ice albedo forcing experiment, however, there is a high pressure response over northern Eurasia and a deepening of the Aleutian low. This high pressure response has been seen in many studies (e.g., Honda et al. 2009; Mori et al. 2014) in response to sea ice loss in the Barents and Kara Seas and can cause a cooling response in East Asia, as a result of cold air advection. Even though both the albedo forcing and RCP8.5 forcing experiments show decreased sea ice loss in this region, the SLP responses are opposite. Based on this, it no surprise that the decomposition tells us that the low-latitude surface warming and sea ice loss parts of the response are generally opposite, similar to the $\delta Z_R = -\delta Z_A$ case [see Eq. (9)]. The low-latitude warming and sea ice loss parts both show a reduction in SLP over the Chukchi, Beaufort, and Bering Seas, as well as over parts of the Canadian Arctic. They have opposite signs over a broader hemispheric region.
including the North Pacific, the rest of the Arctic Ocean, and Eurasia, consistent with the zonal mean wind response, which also produced an opposite response in the troposphere.

Applying this analysis to 2-m temperature over land (Fig. 9) reveals an advection pattern that can be inferred from the SLP response. The sea ice loss produces a moderate cooling response over East Asia as result of cold air advection from the Arctic, consistent with previous studies (e.g., Honda et al. 2009; Mori et al. 2014). However, the warming over this same region connected with the low-latitude surface warming is strong, cancelling out any cooling due to sea ice, so there is still significant warming over East Asia in the RCP8.5 experiment. Even in the albedo forcing experiment, the small amount of low-latitude surface warming is enough to cancel any cooling, leaving instead a lack of warming in that region. Over North America, both the sea ice and low-latitude surface warming contribute to the warming at higher latitudes over Canada, but the influence of the sea ice loss does not reach past the northern United States.

Examining Figs. 7–9 together, we see that many of the sea ice–driven changes inferred from our experiments are equivalent barotropic in circulation and project onto planetary wave 1, with a positive center of action over Eurasia and a negative center over the northeast Pacific at the surface (Fig. 8d) and the midtroposphere (Fig. 7d). We expect that this pattern reflects regional changes in the regional eddy-driven circulation arising from the sea ice perturbation. These equivalent barotropic circulation features are associated with surface temperature advection signals that presumably drive some of the pattern of Eurasian cooling and western North American warming. However, the response is less vertically coherent over the main sea ice loss regions of

![Fig. 9. As in Fig. 5 but for 2-m land temperature (°C).](image-url)
the central Arctic and Hudson Bay. In these regions, the strong regional heating from sea ice loss might directly drive strong changes to the diabatic circulation and lead to less barotropic responses. The combination of direct diabatically driven responses and indirect eddy driven responses are expected to be model dependent, as we will mention in the next subsection.

The analysis applied to winter precipitation (Fig. 10) shows that in the RCP8.5 forcing experiment there are increases in precipitation over most of the middle and high latitudes. In the sea ice albedo forcing simulations, there are more limited increases over the Arctic Ocean where there is sea ice loss, over the west coast of North America and over western Europe. The increases over the Arctic Ocean and over the west coast of North America are seen in the RCP8.5 experiments as well, meaning that sea ice loss contributes to the projected increase in precipitation over these regions. The rest of the increases, mainly over the North Pacific and northern Eurasia, cannot be attributed to sea ice loss but can be attributed to that part of the global warming response associated with low-latitude surface warming. Some of the precipitation responses can be linked to the equivalent barotropic circulation response in SLP and Z500 in Figs. 7 and 8. For example, over the northeast Pacific the increased precipitation over the west coast of North America appears to be connected with the deepening of the Aleutian low.

There have been a number of studies that have investigated how the variability of temperature and other variables are changing in response to greenhouse-dominated radiative forcing and sea ice loss (Screen 2014; Screen et al. 2015; Blackport and Kushner 2016). Here we investigate the role of sea ice loss in the change of temperature variability by performing the decomposition for the subseasonal 2-m temperature variability.

**FIG. 10.** As in Fig. 5, but for precipitation (mm day$^{-1}$).
In the RCP8.5 simulations, the variability is calculated by taking the daily averaged 2-m temperature for each of the 30 realizations over the same 10-yr epochs used in the previous plots, and subtracting the ensemble mean (this removes the seasonal cycle and the ensemble mean trend). The DJF seasonal mean of each of the 300 years is subtracted to obtain the subseasonal anomalies from which the standard deviation is calculated at each geographic point. The difference of the standard deviations of subseasonal 2-m temperature between the two RCP8.5 epochs is plotted in Fig. 11a. Similarly, the subseasonal 2-m temperature standard deviations for the albedo forcing experiments are calculated by subtracting the daily climatological values of the daily temperature and seasonal average DJF temperatures from 200 years of daily data. The difference in this statistic for the sea ice albedo forcing experiments is plotted in Fig. 11b, and the decomposition of the responses is plotted in Figs. 11c and 11d.

In the RCP8.5 experiment, there is a decrease in temperature variability throughout most of the middle and high latitudes, indicating that the Northern Hemisphere winters will become less extreme, a result that is robust across all CMIP5 models (Screen 2014). We also see a reduction in variability in the albedo forcing experiment, consistent with other studies that have looked at the temperature variability response to sea ice loss (Screen et al. 2015; Blackport and Kushner 2016). This suggests that sea ice loss plays a large role in the reduction of variability. Once the decomposition is calculated, we find that sea ice loss is responsible for much of reduced variability over southern Canada, northern United States, Russia, and northern Europe. It is not surprising that sea ice loss is responsible for the reduced
temperature variability in the model, as sea ice loss is the primary contributor to Arctic amplification, which causes reduced variability due to a decrease in temperature gradients (Screen 2014; Schneider et al. 2015). Also, because sea ice has a lower heat capacity than the ocean, melting ice will cause the regions where the ice was lost to be more maritime, also reducing the variability. Although the sea ice loss contributes to much of the reduced variability, the low-latitude surface warming part of the response also shows reduced variability, especially over Europe and central Asia.

b. CCSM4 results

We check the robustness of these results with CCSM4, which has weaker Arctic amplification due to shortwave feedbacks associated with optically thicker clouds (Kay et al. 2012). To compensate for having only five realizations of the RCP8.5 experiments (compared to 30 for CESM1) we average two 20-yr epochs (2027–46 and 2047–66) to calculate the response (compared to 10-yr epochs for CESM1). As expected from Kay et al. in the CCSM4 albedo forcing experiment, there is about 50% less sea ice loss in DJF (30% less in the annual mean; see Table 1 and appendix for more details) and less warming. Despite these differences, the amount of low-latitude surface warming per million kilometers of sea ice loss is similar to CESM1 (Table 1). Because we are using only a third of the number of years to calculate the response in the RCP 8.5 experiment, combined with the weaker forcing, the CCSM4 results are not as robust as the CESM1 results. In particular, changing the epochs used to calculate the response in the RCP 8.5 experiment does change some of the results. Nevertheless, there a few robustly estimated differences between the two models that will be highlighted below.

As in CESM1’s RCP8.5 simulation (Fig. 6), the extratropical tropospheric DJF zonal mean wind response in the CCSM4 RCP8.5 simulation (Fig. 12) is weak and in the CCSM4 case statistically insignificant. In the sea ice albedo forcing simulation, the response generally has the same sign as in CESM1 but is proportionally weaker, accounting for the weaker forcing from sea ice loss. Unlike in CESM1, the stratospheric wintertime zonal wind response is weak in in the sea ice albedo forcing experiment. This could be associated with the relatively strong CCSM4 control run polar vortex compared to the CESM1 control run (not shown), making it less susceptible to changes associated with altered planetary wave propagation. These differences could be related to the inclusion of a turbulent mountain stress parameterization in CAM5 (not included in CAM4), which been shown to increase planetary wave propagation into the stratosphere (Richter et al. 2010). This leads to CAM5 having relatively large stratospheric variability and number of sudden stratospheric warmings for a low-top model (Peings and Magnusdottir 2016). Likely linked to the weaker stratospheric response in CCSM4 is the relatively weak wave-1 circulation response (see, e.g., Fig. 13d). The decomposition between the two parts of the response in CCSM4 shows similar competing southward and northward jet shifts to those seen in CESM1. The responses in the troposphere in the RCP8.5 simulations are not statistically significant and changing the epochs used can change the response (not shown). However, the main features of the response decomposition in Figs. 12c and 12d are present for various epochs.

The results for the DJF SLP (Fig. 13) responses in both the RCP8.5 and the sea ice albedo forcing experiments are generally weaker in CCSM4 than in CESM1. This is at least in part due to the weaker response to sea ice albedo forcing in the CCSM4 experiment. Although the sea ice albedo forcing experiment response is robust, many of the features seen in the RCP8.5 forcing experiment (and therefore in both terms in the decomposition as well) are not, as they are not statistically significant or robust for different epochs analyzed. Notably, for the different epochs analyzed, there is no high sea level pressure response over northern Eurasia due to sea ice loss [this is also found in Blackport and Kushner (2016)]. This also results in there being no advective cooling in the East Asia like there was in CESM1 (cf. Figs. 14 and 9). A robust feature of CCSM4 and CESM1 is the opposite responses due to sea ice loss and low-latitude surface warming in the SLP response in the North Pacific (cf. Figs. 13 and 8). This leads to a similar Pacific sector precipitation response in CCSM4 compared to CESM1 (not shown). Many of the remaining features of the SLP response change for different epochs and thus are not robust.

6. Summary and discussion

To isolate the role of sea ice loss in the projected atmospheric response to greenhouse-dominated anthropogenic forcing, we have compared RCP8.5 projections to albedo forcing experiments, which have different amounts of low-latitude surface warming relative to the amount of sea ice loss. We have used a pattern scaling approach in which the projected response to anthropogenic forcing can be decomposed into a part that scales with total Arctic sea ice loss and a part that scales with low-latitude ocean surface warming. We have applied this approach to a number of fields to determine how important sea ice loss is in the atmospheric response to greenhouse gas-dominated radiative forcing in CESM1.
and CCSM4. We found that the results of our scaling approach were similar for different seasonal lags, for scaling by different regions, and for different epochs.

For zonal temperature, Arctic lower tropospheric warming scales with sea ice loss and the rest of the warming with low-latitude surface warming. The boreal winter Northern Hemisphere extratropical circulation response to sea ice loss was associated with a weakening of the zonal winds, an equatorward shift in the jet stream, a high SLP response over northern Eurasia, and a deepening of the Aleutian low. In addition, the sea ice loss part was associated with a weaker polar vortex, but we pointed out some ambiguity in the interpretation of the stratospheric response in these simulations. For many of these circulation changes, the low-latitude surface warming part was opposite to sea ice loss response, leading to cancellation and a relatively weak projected response in the RCP8.5 simulation.

We have found important examples where sea ice loss and low-latitude warming are more regionally distinct or reinforce rather than cancel each other. For the regional response in precipitation, sea ice loss was responsible for an increase of wintertime precipitation over the west coast of North America that is also seen in the RCP8.5, while the low-latitude warming is associated with increases over other sectors such as Eurasia. We also find that sea ice loss and low-latitude warming both act to decrease wintertime subseasonal temperature variability at middle and high latitudes. This response is seen in both the sea ice albedo forcing and the RCP8.5 forcing simulations and both sea ice loss and low-latitude surface warming play a role in it.

The circulation responses in the sea ice albedo forcing experiments and in the sea ice part of the responses have been documented in other studies including those that use prescribed noninteractive SST and sea ice (e.g., Peings and Magnusdottir 2014). The fact that we do not see these same responses in the RCP8.5 experiments, and that such effects are somewhat weaker in the sea ice albedo forcing experiments than in the decomposition, suggests that, in the coupled system, extratropical greenhouse warming responses associated with
low-latitude surface warming can be just as important as, if not more important than, the sea ice loss of Arctic origin. In other words, just because sea ice loss can impact the lower latitudes in many ways, as seen in previous studies, does not necessarily mean it will be the dominant response to climate change, as discussed in Barnes and Screen (2015). This is in agreement with Barnes and Polvani (2015), who found that many of the proposed impacts of Arctic amplification are not seen in the CMIP5 models under RCP8.5 forcing, although there is a significant amount of spread.

The results of this study might suggest that the circulation response to sea ice loss in the coupled model would be weaker than in the experiments that use prescribed SST and sea ice because of the partial cancellation brought about by the “mini global warming.” This, however, was not found in Deser et al. (2015, 2016), who found a stronger circulation response in a coupled model sea ice loss experiment compared to one without an interactive ocean under the same amount of sea ice loss, as a result of a larger temperature response that penetrates higher up into the atmosphere. They attribute this to an increase in poleward heat transport associated with a warmer and moister tropical upper troposphere. We do not see evidence of more warming in the Arctic free troposphere in the RCP8.5 response than in the albedo forcing experiments, even though it has much more tropical warming. One factor that may contribute to the stronger response in the coupled sea ice loss experiments is that the fixed SST and sea ice experiments used in Deser et al. (2015, 2016) only prescribe increased SSTs in the regions where sea ice is lost. In the coupled model experiments, however, there are SST increases well outside these regions that might be attributed to sea ice loss. Indeed, when we perform the decomposition on the SST, we find that there is
warming of the high- and midlatitude oceans that can be attributed to sea ice loss (not shown). These could help to produce a stronger temperature response in the coupled model. As part of a future study, we plan on testing this and other ideas presented here by performing new experiments using prescribed SST and SIC that are calculated using the decomposition approach described in section 4.

We note that although we do not see many of the impacts of sea ice loss in the RCP8.5 experiment, this is likely model dependent. Barnes and Polvani (2015) find that there is a significant amount of spread in the jet stream’s response to greenhouse-dominated radiative forcing and that some of this spread can be explained by the amount of Arctic amplification. Models that have more Arctic amplification see a response in the jet that looks more like the sea ice part of the response, whereas models with less Arctic amplification look more like the low-latitude surface warming part of the response. The models analyzed here tend to be in the middle of the range of responses, but in models with more sea ice loss or in models that have a larger Arctic amplification response sea ice could play a larger role than in the models analyzed in this study. In addition, even with the same Arctic amplification or lower-latitude warming, different models with different parameterizations, resolutions, or stratospheric representations could respond with a different amplitude or even sign. For example, Sun et al. (2015) found a stronger circulation response to sea ice loss using a model with a better representation of the stratosphere, compared to its low-top counterpart.

To reconsider the larger picture, we emphasize that the sea ice albedo forcing Earth system model experiments used here are an artificial means to isolate the part of the sea ice loss process associated with sea ice albedo.
shortwave feedbacks and summertime sea ice loss (Blackport and Kushner 2016; Bitz et al. 2006). These experiments show that the tropospheric near-surface warming in the Arctic can be physically attributed to sea ice loss, once the warming is scaled by the sea ice loss (Figs. 4 and 5). In addition, these experiments, combined with radiative forcing experiments, provide a possible causal chain to explain aspects of the tropospheric circulation response to anthropogenic greenhouse forcing, without the need to run additional experiments. Under greenhouse warming, anthropogenic radiative forcing quickly increases atmospheric water vapor and drives tropical upper troposphere amplified warming and global stratospheric cooling. In isolation, this will tend to drive the jet stream poleward and strengthen the wintertime Arctic stratospheric vortex (Figs. 5c and 6c). However, the greenhouse gas and water vapor enhanced downwelling longwave radiation simultaneously leads to sea ice loss and Arctic amplification, boosted by positive shortwave sea ice albedo feedbacks (and possibly other positive surface energy budget feedbacks in the Arctic). This additional warming contributes, primarily via oceanic warming in the coupled system (Tomas et al. 2016), to the original global warming effect, and so in terms of the thermal and radiative response provides a positive feedback mechanism for greenhouse forcing circulation change. But the direct dynamical effect of sea ice loss is a negative circulation feedback that pushes the jet equatorward and dynamically warms the polar stratosphere (Figs. 5d and 6d). Such negative feedbacks associated with the eddy-driven response to sea ice forcing have been well documented before in the context of North Atlantic Oscillation trends (e.g., Magnusdottir et al. 2004; Deser et al. 2004) and appear here to be acting in the context of pan-Arctic sea ice loss and the zonal mean circulation. In summary, in this interpretation, greenhouse gas warming induced sea ice loss provides a negative feedback that weakens the impact of the extratropical circulation response to greenhouse warming.

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APPENDIX

Comparison of the Response to Sea Ice Albedo Forcing between CESM1 and CCSM4

SIA values in the control simulations of CESM1 and CCSM4 (Fig. A1a) show similar amounts of sea ice in the annual mean but CESM1 has a larger-amplitude seasonal cycle. In most months of the year, the response in SIA is larger in CESM1 than in CCSM4. The only months that do not show this are August and September, but this is only because there is less ice available to melt.
in the CESM1 control simulations during these months. In DJF, the CESM1 control simulation has slightly more ice, but the sea ice albedo forcing simulation has considerably more ice melt than CCSM4 (2.28 vs 1.17 million km$^2$ lost). Much of this difference comes in December when there is a 2 million km$^2$ difference in the amount of ice loss between the two models. All of these differences in the sea ice response between the two models are consistent with there being stronger shortwave feedbacks in CESM1 (Kay et al. 2012).

Arctic sea ice thickness (Fig. A1b) is similar in the two control simulations, but CESM1 has thinner sea ice throughout the summer and fall, consistent with greater downwelling surface shortwave radiation. The responses in both models involves approximately 0.5 m of thinning throughout most of the year, but an increase in thickness in late summer when all but the thickest ice near the north coast of Greenland and the Canadian Arctic Archipelago has melted. This peak in ice thickness in the sea ice albedo forcing simulation is shifted a month earlier in CESM1 because the ice melts faster and then by September even the thickest ice has been able to thin to about 0.5 m.

The control simulation’s seasonal cycle of Arctic 2-m temperature (averaged from 65° to 90°N; Fig. A1c) is colder in CESM1 than in CCSM4 in the control simulation, particularly in DJF. Under sea ice albedo forcing simulations, however, the two models show similar Arctic temperatures in DJF. This is associated with approximately 70% more warming in CESM than in CCSM4 (5.8°C vs 3.3°C), which is consistent with the sea ice response. Both models have qualitatively similar seasonal cycles in surface heat flux response (Fig. A2), but CESM1 generally has a larger response, consistent with the sea ice and temperature responses. In June, when the shortwave response is strongest, CESM1 has about 100% more shortwave absorbed at the surface (note that the shortwave response is multiplied by 0.1).

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