The Role of Extratropical Ocean Warming in the Coupled Climate Response to Arctic Sea Ice Loss

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ABSTRACT

The role of extratropical ocean warming in the coupled climate response to Arctic sea ice loss is investigated using coupled atmosphere–ocean general circulation model (AOGCM) and uncoupled atmospheric-only (AGCM) experiments. Coupled AOGCM experiments driven by sea ice albedo reduction and greenhouse gas–dominated radiative forcing are used to diagnose the extratropical sea surface temperature (SST) response to sea ice loss. Sea ice loss is then imposed in AGCM experiments both with and without these extratropical SST changes, which are found to extend beyond the regions where sea ice is lost. Sea ice loss in isolation drives warming that is confined to the Arctic lower troposphere and only a weak atmospheric circulation response. When the extratropical SST response caused by sea ice loss is also included in the forcing, the warming extends into the Arctic midtroposphere during winter. This coincides with a stronger atmospheric circulation response, including an equatorward shift in the eddy-driven jet, a deepening of the Aleutian low, and an expansion of the Siberian high. Similar results are found whether the extratropical SST forcing is taken directly from the AOGCM driven by sea ice loss, or whether they are diagnosed using a two-parameter pattern scaling technique where tropical adjustment to sea ice loss is removed. These results suggest that AGCM experiments that are driven by sea ice loss and only local SST increases will underestimate the Arctic midtroposphere warming and atmospheric circulation response to sea ice loss, compared to AOGCM simulations and the real world.

1. Introduction

Recent decades have seen the Arctic warm more than twice as fast as the global average temperature—a process known as Arctic amplification (Holland and Bitz 2003; Screen and Simmonds 2010; Walsh 2014; Cohen et al. 2014). This has motivated investigation into the impacts of rapid Arctic warming and sea ice loss on the large-scale atmosphere circulation, including its contribution to extreme weather events at lower latitudes. However, there remains large uncertainties due to the short observational record, difficulty in diagnosing cause and effect, and disagreements among modeling studies (Francis and Vavrus 2012; Screen and Simmonds 2013; Cohen et al. 2014; Barnes and Polvani 2015; Barnes and Screen 2015; Screen et al. 2018). To isolate the impact that Arctic sea ice loss has on the atmosphere, numerous modeling experiments forced with observed or projected sea ice loss have been performed. This is typically done by forcing an atmospheric general circulation model (AGCM) with prescribed sea ice loss while keeping the sea surface temperatures (SST) constant. Some common responses in these experiments include Arctic lower-tropospheric warming and a weakening and equatorward shift in the eddy-driven jet (e.g., Deser et al. 2010; Peings and Magnusdottir 2014; Nakamura et al. 2015; Sun et al. 2015).

Using an AGCM with fixed SST neglects potentially important thermodynamic and dynamical feedbacks with the ocean. As a result, there has been increased attention recently on examining the impacts of sea ice loss on the atmosphere using full atmosphere–ocean coupled models (AOGCM) or slab ocean models (Deser et al. 2015, 2016; Semmler et al. 2016; Tomas et al. 2016; Blackport and Kushner 2016, 2017; Oudar...
et al. 2017; Smith et al. 2017; McCusker et al. 2017; Screen et al. 2018; Hay et al. 2018). In particular, Deser et al. (2015, 2016) and Smith et al. (2017) examine the impact of ocean–atmosphere coupling on the atmospheric response to sea ice loss by comparing the response in uncoupled AGCM experiments with AOGCM experiments with the same sea ice forcing. Smith et al. (2017) found that in the AOGCM, the warming spreads into the ocean, amplifying the temperature response. They also found opposite North Atlantic Oscillation (NAO) responses in the two experiments, which were attributed to the different background states rather than the additional warming. Deser et al. (2015, 2016) found that while the uncoupled AGCM warms only the Arctic lower troposphere, coupling to the ocean extends this warming into the Arctic midtroposphere (AMT) and to lower latitudes. The zonal-mean temperature response in the coupled ocean–atmosphere model was found to resemble the response to increased greenhouse gases, but with a smaller magnitude, so they refer to it as a “mini global warming.” In addition, they found ocean–atmosphere coupling increased the magnitude of the atmospheric circulation response to sea ice loss by approximately 50%, but the overall structure remained similar. This amplified circulation response occurs despite the presence of a tropical upper-troposphere warming response, which may be expected to oppose the atmospheric circulation response to Arctic warming and dampen the response (Butler et al. 2010; Blackport and Kushner 2013). The stronger circulation response was attributed to the enhanced warming in the AMT that was not seen in the uncoupled AGCM experiment. Deser et al. (2015, 2016) speculated that increased heat transport from the tropics caused this AMT warming.

While AGCM experiments have shown that sea ice loss drives strong warming primarily in the Arctic lower troposphere, reanalysis has shown that Arctic amplification extends into the AMT (Graversen et al. 2008; Screen and Simmonds 2010; Screen et al. 2012). Identifying the causes of warming in the AMT is important, as it has been shown that warming higher in the Arctic atmosphere can have a greater impact on the atmospheric circulation (Sellevold et al. 2016; Baker et al. 2017). A number of studies have concluded that this warming in the AMT can be attributed to increased poleward heat transport caused by SST changes outside the Arctic (Chung and Räisänen 2011; Screen et al. 2012; Laliberté and Kushner 2013; Perlwitz et al. 2015). By using AGCM experiments forced by observed changes in SST and sea ice together with experiments forced with only sea ice loss and local SST changes, Screen et al. (2012) concluded that sea ice loss is the primary driver of temperature trends near the surface, while SST changes outside the Arctic are responsible for the temperature trends in the AMT. They find that AMT warming trends driven by remote SST changes are largest in September and October, but statistically significant trends are found year-round, with the exception of springtime. Using a similar set of AGCM experiments, Perlwitz et al. (2015) come to the same conclusion, finding that the majority of Arctic warming above the near surface in October–December is driven by long-term trends and fluctuations in SSTs outside the Arctic.

In this study, we expand on the results of Screen et al. (2012), Perlwitz et al. (2015), and Deser et al. (2015, 2016) by investigating how sea ice loss impacts the AMT and associated circulation changes via changes in extratropical SST. We hypothesize a two-step process by which Arctic sea ice loss contributes to warming in the AMT and amplifies the atmospheric circulation response to sea ice loss beyond what has been found in previous AGCM experiments. First, the warming from sea ice loss spreads out beyond the regions where sea ice disappears and into the extratropical ocean. This occurs through the advection of warm air and downward heat fluxes into the ocean, but can also be influenced by ocean dynamics (Wang et al. 2018). This additional extratropical warming induced by sea ice loss is then transported into the AMT by the atmospheric circulation. Unlike AGCM experiments, where some of this warming will be absent as the SST cannot respond, coupled AOGCMs can capture this process. This hypothesis suggests that some of the remote SST changes prescribed in AGCM experiments done by Screen et al. (2012) and Perlwitz et al. (2015) may in fact be the result of sea ice loss, and, therefore, the influence of sea ice loss on AMT warming and atmospheric circulation may be underestimated.

To test this, we use a series of modeling experiments using both coupled AOGCM and uncoupled AGCM experiments, which will be described section 2. We will use the coupled AOGCM experiments to diagnose the extratropical SST response connected with sea ice loss. Next, we will attempt to isolate the impact of these SST changes associated with sea ice loss using AGCM experiments forced by sea ice loss both with and without these SST changes. Using the SST and sea ice output from coupled climate model experiments to drive AGCM experiments is common practice; however, these either include no changes in SST (e.g., Deser et al. 2010) or changes in SSTs only in regions where there is significant sea ice loss (Peings and Magnusdottir 2014; Sun et al. 2015; Deser et al. 2015, 2016). In section 3, we will show that prescribing the SST response associated with sea ice loss acts to amplify the AMT warming and
enhance the atmospheric circulation response. Using additional experiments, we will attribute most of the impact to the SST response away from regions where the sea ice is lost, but still associated with sea ice loss. Finally, we will summarize and discuss the results in section 4.

2. Experiment design

a. Model description

The model used is the Community Earth System Model version 1 (CESM1), which is developed at the National Center for Atmospheric Research (NCAR). The atmospheric component of the model is the Community Atmosphere Model version 5 (CAM5), which uses a finite-volume dynamical core and has a horizontal resolution of 0.90° latitude and 1.25° longitude with 30 vertical levels up to approximately 3 hPa. The ocean model is the Parallel Ocean Program, version 2 (POP2), which has a horizontal resolution of approximately 1° with 60 vertical levels. The land model is the Community Land Model version 4 (CLM4), and the sea ice model is the Community Ice Code 4 (CICE4); these are run on the same grid as the atmosphere and ocean, respectively. The AGCM simulations use CAM5 coupled to CLM4 with prescribed SST, sea ice concentrations (SIC), and sea ice thickness (SIT). All simulations were forced with a repeated seasonal cycle of SST, SIC, and SIT for 100 years under constant year 2000 radiative forcing. Monthly averaged values for these boundary conditions were either taken directly or diagnosed from the AOGCM simulations described above and interpolated to submonthly values. A summary of the boundary conditions used in each AGCM simulation is outlined in Table 1 and described in detail below.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>SIC/SIT</th>
<th>SST</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTLALB</td>
<td>Coupled year 2000 control</td>
<td>Coupled year 2000 control</td>
</tr>
<tr>
<td>IALB</td>
<td>Sea ice albedo forcing</td>
<td>Coupled year 2000 control</td>
</tr>
<tr>
<td>IALBLEB40</td>
<td>Sea ice albedo forcing</td>
<td>Coupled year 2000 control</td>
</tr>
<tr>
<td>CTLRCP</td>
<td>RCP8.5 2027–36</td>
<td>RCP8.5 2027–36</td>
</tr>
<tr>
<td>IRCP</td>
<td>RCP8.5 2057–66</td>
<td>RCP8.5 2057–66</td>
</tr>
<tr>
<td>IRCPTSIL</td>
<td>RCP8.5 2057–66</td>
<td>RCP8.5 2027–36 + T_{SIL}</td>
</tr>
<tr>
<td>IRCPTSIL_IL</td>
<td>RCP8.5 2057–66</td>
<td>RCP8.5 2027–36 + T_{SIL}</td>
</tr>
<tr>
<td>IRCPTSIL_AIL</td>
<td>RCP8.5 2057–66</td>
<td>RCP8.5 2027–36 + T_{SIL}</td>
</tr>
<tr>
<td>IALBTSIL</td>
<td>Sea ice albedo forcing</td>
<td>Coupled year 2000 control + T_{SIL}</td>
</tr>
</tbody>
</table>

We initially performed a set of three AGCM simulations using boundary conditions from the albedo forcing and year 2000 control AOGCM simulations:

- The first simulation, termed CTL_{ALB}, was forced with the long-term averaged seasonal cycle of SST, SIC, and SIT from the AOGCM year 2000 control simulation.
• To isolate the impacts of only sea ice loss, a second simulation, termed \(I_{ALB}\), was performed that had the same SST boundary conditions as \( CTL_{ALB}\), but the Northern Hemisphere SIC and SIT were taken from the albedo forcing simulation, which had reduced sea ice.

• A third simulation, termed \(I_{ALB}T_{ALB40}\), was performed to isolate the impact on the atmosphere of the extratropical SST response to sea ice loss in the coupled system. For \(I_{ALB}T_{ALB40}\), the SIC and SIT are from the same albedo forcing simulations as in \(I_{ALB}\) and so represent the same sea ice loss associated with the albedo forcing. But \(I_{ALB}T_{ALB40}\) uses different SST than \(I_{ALB}\). For \(I_{ALB}T_{ALB40}\), the SSTs from \( CTL_{ALB}\) are used south of 40°N, as in \(I_{ALB}\), but the SSTs north of 40°N are taken from the albedo forcing simulation. Thus, the only difference in forcing between the \(I_{ALB}T_{ALB40}\) and \(I_{ALB}\) simulations is the extratropical SST change poleward of 40°N, which captures the impact of albedo-forced sea ice loss on extratropical ocean temperatures.

While these three simulations are a direct way to capture the impact of the extratropical SST response associated with sea ice loss, some of this response is associated with the “mini global warming” response (Deser et al. 2015). The albedo forcing experiment shows a larger global warming response, compared to similar AOGCM experiments forced with Arctic sea ice loss (Screen et al. 2018), which we recall could also be influenced by the induced Antarctic sea ice loss in the simulations used here. We can account for the associated tropical adjustment and diagnose the SST response more directly associated with sea ice loss using a two-parameter pattern scaling approach, which we will briefly outline below. For a more detailed description of this approach, see Blackport and Kushner (2017). We start by considering how the SST at each grid point will respond to a change in sea ice area and low-latitude sea surface temperature:

\[
\delta \text{SST} = \text{SST}(T + \delta T, I + \delta I) - \text{SST}(T, I)
\]

\[
\approx \left. \frac{\partial \text{SST}}{\partial T} \right|_I \delta T + \left. \frac{\partial \text{SST}}{\partial I} \right|_T \delta I.
\]

SST refers here to the sea surface temperature at each grid point, which is a function of the total amount of Arctic sea ice area \(I\) and low-latitude (averaged from 0° to 40°N) sea surface temperature \(T\). The first partial derivative on the right-hand side of Eq. (1) represents the response of the sea surface temperature to a change in low-latitude temperature while keeping Arctic sea ice constant. Similarly, the second partial derivative represents the response of the sea surface temperature to a change in Arctic sea ice while keeping low-latitude temperature constant. To estimate the SST response associated with sea ice loss in the absence of tropical warming, we calculate the second term on the right-hand side of Eq. (1). To calculate this partial derivative, we will use the output from the two sets of AOGCM experiments. We can write out the change in SST at each grid point for each experiment as

\[
\delta \text{SST}_A = \left. \frac{\partial \text{SST}}{\partial T} \right|_I \delta T_A + \left. \frac{\partial \text{SST}}{\partial I} \right|_T \delta I_A
\]

and

\[
\delta \text{SST}_R = \left. \frac{\partial \text{SST}}{\partial T} \right|_I \delta T_R + \left. \frac{\partial \text{SST}}{\partial I} \right|_T \delta I_R.
\]

where the \(R\) and \(A\) subscripts represent the RCP8.5 and albedo forcing experiments, respectively. As each quantity in Eqs. (2) and (3) is known except for the partial derivatives, we can solve for the partial derivatives:

\[
\begin{pmatrix}
\left. \frac{\partial \text{SST}}{\partial T} \right|_I \\
\left. \frac{\partial \text{SST}}{\partial I} \right|_T
\end{pmatrix} = \frac{1}{\delta I_A \delta T_R - \delta I_R \delta T_A} \begin{pmatrix}
-\delta I_R & \delta I_A \\
\delta T_R & -\delta T_A
\end{pmatrix} \begin{pmatrix}
\delta \text{SST}_A \\
\delta \text{SST}_R
\end{pmatrix}.
\]

In Eq. (4), the partial derivatives and \(\delta \text{SST}\) terms are spatial fields, while the \(\delta I\) and \(\delta T\) terms are single numerical values that are derived from the AOGCM experiments (Blackport and Kushner 2017).

This procedure provides a separate estimate of the impact of sea ice loss on the ocean surface temperature, accounting for tropical adjustments under the “mini” global warming arising from sea ice loss. This motivates us to carry out a complementary set of three AGCM simulations to probe the impact of SST adjustment on the atmospheric response:

• The first simulation, termed \(CTL_{RCP}\), was forced with the SST averaged over 2027–36 from the ensemble mean of the RCP8.5 simulations of the CESM Large Ensemble.

• To isolate the impacts of only sea ice loss, a second simulation, termed \(IR_{RCP}\), was performed with the same SST boundary conditions as \(CTL_{RCP}\), but using the SIC and SIT averaged over 2057–66 from the RCP8.5 simulations. The altered sea ice represents sea ice loss after 30 additional years of projected greenhouse...
warming [2027–36 and 2057–66 were the focus epochs of Blackport and Kushner (2017)]. These epochs were chosen to match the total December–February (DJF) Arctic sea ice in the coupled model albedo forcing and year 2000 control simulations, respectively.

- A third simulation, termed I_RCP_T_SIL, was performed; as in the first set of simulations, this simulation attempts to isolate the impact of the SST response to sea ice loss in the coupled system. The I_RCP_T_SIL simulation has the same sea ice as I_RCP, but was forced with the change in SST associated with sea ice loss [the second term on the right-hand side of Eq. (1); hereafter T_SIL] calculated from Eq. (4) for the Northern Hemisphere. To calculate the monthly SST forcing required for the simulation, the calculation was performed for each month using the corresponding month's sea ice area and low-latitude SST (i.e., to calculate T_SIL for January, January δ_I and δ_T were used). The responses used in the calculation of Eq. (4) are defined to be the difference between the averages over the 10-yr epochs 2057–66 and 2027–36, which again were the focus epochs of the analysis by Blackport and Kushner (2017).

These two sets of three AGCM simulations each allow us to examine the impact that extratropical ocean warming from sea ice loss has on the atmosphere; however, they have a number of potentially important differences. First, the I_ALB_T_ALB40 simulation includes additional SST warming associated with the long-term adjustment of the climate system that is part of the “mini global warming” response, while the I_RCP_T_SIL simulation attempts to capture only the SST changes more directly associated with sea ice loss. Additionally, although the Northern Hemisphere means of the DJF SSTs, SIC, and SIT used in the CTL simulations are approximately the same, the spatial patterns differ. This is because they were determined from transient simulations in the case of CTL_RCP and equilibrium simulations for CTL_ALB. Similarly, the spatial pattern of the prescribed sea ice loss differs in the I_RCP and I_ALB simulations, although the total change in DJF sea ice area (SIA) is approximately the same (by virtue of the choice of epochs in the RCP8.5 experiments). These differences could potentially be important, as it has been shown that the response is dependent on the background state for both sea ice forcing (Screen and Francis 2016; Osborne et al. 2017) and SST forcing (Zhou et al. 2017). With dependence on the background in mind, a seventh AGCM simulation was performed to impose the SST perturbation from the second set of AGCM simulations on the background state from the first set of AGCM simulations:

- This simulation, termed I_ALB_T_SIL, has the same background SST, SIT, and SIC, along with the same SIC and SST forcing as I_ALB, but with the T_SIL SST pattern used for the SST forcing (as found in the difference between I_RCP_T_SIL and I_RCP) instead of the SST pattern taken directly from the albedo forcing experiments (as found in the difference between I_ALB_T_ALB40 and I_ALB).

Some previous studies that examined the atmospheric response to sea ice loss in AGCM experiments included SST increases only in regions where sea ice disappeared (Screen et al. 2013; Peings and Magnusdottir 2014; Deser et al. 2015; Sun et al. 2015). To test how important these local SST changes are compared to remote SST changes also caused by sea ice loss, we finally performed two additional simulations:

- One of these, termed I_RCP_T_SIL_IL, is identical to the I_RCP_T_SIL simulation, except it only includes SST changes where the change in SIC was greater than 10%, following Screen et al. (2013). The “IL” subscript tag refers here to including SST adjustments only in the ice loss region.

- The other of these, termed I_RCP_T_SIL_AIL, complements I_RCP_T_SIL IL and thus only includes the changes in SST where the changes in SIC are less than 10%. The “AIL” subscript tag refers here to including SST adjustments only away from the ice loss region.

3. Results

a. Sea ice and sea surface temperature response in the AOGCM experiments

The DJF SIC and SIT responses in the albedo forcing experiments—which are also used to force the I_ALB, I_ALB_T_ALB40, and I_ALB_T_SIL simulations—are shown in Figs. 1a and 1b. Reductions in SIC are found in the marginal ice zone, with the largest changes occurring in the Barents–Kara Sea region. The SIT shows reductions over all regions, generally of less than 1 m, except for along the northern coasts of Greenland and the Canadian Arctic Archipelago. Figures 1d and 1e show the response in SIC and SIT for the RCP8.5 simulations from the CESM1 Large Ensemble, shown as the difference between the 2057–66 and 2027–36 epochs. These 10-yr epochs were chosen so that the SIA in the RCP8.5 simulations matched the SIA in the year 2000 and albedo forcing simulations, so the responses are similar; however, they are not identical. There are larger reductions in SIC in the Barents–Kara Sea, smaller reductions over the Pacific side of the Arctic Ocean, and smaller reductions in SIT in the RCP8.5 simulations, compared to the albedo forcing simulations.
The largest DJF SST warming response in the albedo forcing simulations (Fig. 1c) is in the regions with reductions in SIC, in particular over the Atlantic side of the Arctic Ocean. Warming extends beyond the regions where the sea ice was lost and into the North Atlantic and North Pacific Oceans. However, there are some regions over the midlatitude oceans that show a lack of warming or even small regions of cooling, in particular over the Pacific Ocean. Figure 1f shows the DJF SSTs directly associated with sea ice loss as calculated by the pattern scaling method in Eq. (4). Near the regions where the sea ice is lost, the warming is similar to the response in the albedo forcing experiment (Fig. 1c). Farther away from these regions, there is less warming compared to the albedo forcing experiment, as some of this warming was diagnosed to be associated with warming at low latitudes. Some insight into the causes of the SST patterns in Figs. 1c and 1f can be gained by examining the surface heat flux response (not shown). Over the regions with sea ice loss, the warming is caused by an increase in shortwave flux into the ocean. In nearby regions in the North Atlantic and North Pacific Oceans, the warming is associated with increased turbulent heat flux into the ocean from the atmosphere, consistent with the ocean coupling allowing the warming to spread out (Smith et al. 2017). The cooling in the Pacific Ocean in Fig. 1f, the warming over the eastern Pacific, and the small region of cooling in the middle of the North Atlantic are likely a result of changes in the atmosphere and ocean circulation, which Wang et al. (2018) have pointed out may be of tropical origin.

While a more complete analysis of the ocean response in the coupled model simulations is beyond the scope of this study, it is noteworthy that the midlatitude SST
responses described here may involve several sources of cross-latitude coupling.

b. Vertical structure of the atmospheric temperature response in the AGCM experiments

Figure 2 shows the seasonal cycle of the polar cap temperature (averaged from 65° to 90°N) response in the AGCM experiments. In response to only sea ice loss in the IALB simulation (Fig. 2a), the warming peaks in late fall and early winter and is trapped near the surface (below 800 hPa). When the SST warming from northward of 40°N was prescribed in the IALBTALB40 simulation, the response is similar, but the warming extends higher into the troposphere (Fig. 2b). The difference between the IALBTALB40 and IALB simulations (i.e., the response to the extratropical SST changes only) is shown in Fig. 2c. Near the surface, the extratropical SST changes have the largest impact in summer and fall. However, in the AMT, the warming from the extratropical SST forcing peaks in winter, despite the warming near the surface being weaker than it was in summer and fall. Similar results are found in the IRCP and IRCERTALB experiments (Figs. 2d–f), with surface warming peaking in summer and fall and the midtroposphere warming peaking in winter. In the troposphere, the largest difference between the two sets of experiments occurs in the AMT during the summer and fall. During this time, the warming from the extratropical SST forcing from the IRCERTALB experiment does not penetrate beyond 700 hPa, while the warming extends throughout the Arctic troposphere in the IALBTALB40 experiment. The stratospheric response to sea ice loss differs in the two experiments, with the IALB experiment showing a slight but not statistically significant cooling, while IRCP shows warming. These could be a result of the small differences in regional sea ice forcing between the two sets of experiments, which has been shown to be important for the stratospheric response (Sun et al. 2015).
The seasonality of the polar cap temperature response and the differences between the two sets of experiments seen in the right-hand column of Fig. 2 are consistent with the extratropical SST warming from sea ice loss driving the AMT warming. During summer and fall, sea ice loss, and the SST warming it causes, occurs almost entirely over the polar cap. This results in strong surface warming caused by the increased SST. In the IALBTALB40 experiment, there is weak, but not negligible, SST warming outside the Arctic in the summer; however, this part of the response was diagnosed as scaling with lower-latitude temperature instead of sea ice loss by the pattern scaling and therefore is absent in the IRCPTSIL experiment (not shown). These differences in SST forcing can explain the differences in AMT warming in summer and fall between the top and bottom rows of Fig. 2. In contrast, during winter, the sea ice loss occurs at lower latitudes, resulting in SST warming that is directly associated with sea ice loss extending into the midlatitudes and only weak SST warming over the polar cap. In response to this SST warming that extends beyond the Arctic, we see AMT warming in both sets of experiments. The slightly weaker AMT warming in IRCPTSIL can be attributed to some of the ocean warming being subtracted out as a result of the pattern scaling that accounts for low-latitude warming effects from sea ice loss. Neither experiment shows AMT warming in spring, despite there being SST forcing outside the Arctic during this time. The reason for this is unknown, but it is consistent with Screen et al. (2012), who also found a lack of AMT warming in spring in response to observed global SST increases.

Because the AMT warming from extratropical SST is strongest in winter, we will focus on the DJF response for the rest of the study. In response to only reductions in sea ice concentration and thickness in the IALB experiment, the DJF warming is trapped near the surface (Fig. 3a), with nearly all the warming occurring below 800 hPa. Table 2 shows the DJF Arctic (poleward of 65°N) 2-m and midtropospheric (averaged between 700 and 400 hPa) temperature response in each AGCM experiment. In the IALB simulation, despite the near surface warming by 4.57°C, the AMT shows no warming (~0.02°C), consistent with previous AGCM studies (Screen et al. 2012; Perlwitz et al. 2015). When the SSTs from poleward of 40°N are prescribed in the IALBTALB40 simulation, the warming extends to lower latitudes and higher into the atmosphere (Fig. 3b), with the AMT warming increasing to 0.43°C. The difference between the IALBTALB40 and IALB (i.e., the response to only the extratropical SST warming) is shown in Fig. 3c. The extra warming caused by the SST increases is not confined to the lower troposphere, as the warming in the AMT is comparable to the near surface (0.45° vs 0.62°C), with a peak around 850 hPa. A similar response to the SST warming is found in the IRCPTSIL experiment (Fig. 4a), with differences in magnitude that are consistent with the differences in forcing.

Next, we examine the contributions of the SST warming in regions where sea ice is lost, compared to the
SST warming away from these regions (but that we still attribute to sea ice loss). Some previous studies that prescribe sea ice loss in AGCM experiments do include warming in SST in the regions where the sea ice decreases (e.g., Screen et al. 2013; Peings and Magnusdottir 2014; Deser et al. 2016). The zonal-mean temperature response to SST warming in only regions where the sea ice is lost is very small (Fig. 4b), with an AMT warming of only 0.01°C. In response to only the SST warming away from the sea ice, the warming extends higher into the troposphere and into the AMT (0.19°C), indicating that the remote SST warming is much more important for the response (Fig. 4c). Note, however, that the addition of the two responses does not add up to the total response. For example, the AMT warming response seen in response to SST changes both in and away from the regions where sea ice is lost is 0.33°C, whereas the sum of the responses is 0.20°C. The origin of this nonlinearity in the temperature response, which is also found in regional sea ice loss AGCM experiments (Screen 2017), is an intriguing open question.

c. Atmospheric circulation response

We now investigate the impact of the extratropical SST warming associated with sea ice loss on the atmospheric circulation. Figure 5 shows the response of the DJF zonal-mean zonal wind. In response to sea ice loss without any changes in SST, the I ALB experiment (Fig. 5a) shows little response in the troposphere, while in the RCP experiment (Fig. 5d), there is a weakening and equatorward shift in the eddy-driven jet. These different responses could be associated with the differences in the spatial pattern of the forcing causing different stratospheric responses (Sun et al. 2015). The change in the jet seen in the RCP experiment is qualitatively similar to the response seen in Peings and Magnusdottir (2014), who also prescribed sea ice loss derived from the RCP8.5 experiments and used the same AGCM as used in this study.

When the extratropical SST forcing is included in the forcing in the I ALBTALB40 experiment, the response is very different (Fig. 5b). The jet is weakened at high latitudes, with an easterly response in zonal wind of approximately 1 m s⁻¹ at 300 hPa and 55°N. Similarly, the IRCPTSIL (Fig. 5e) experiment also shows a significantly larger response of zonal winds, compared to RCP.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>AMT response (°C)</th>
<th>Arctic 2-m temperature response (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>I ALB – CTLALB</td>
<td>−0.02</td>
<td>4.57</td>
</tr>
<tr>
<td>I ALB TALB40 – CTLALB</td>
<td>0.43</td>
<td>5.19</td>
</tr>
<tr>
<td>I RCP – CTLRCP</td>
<td>0.13</td>
<td>4.21</td>
</tr>
<tr>
<td>I RCP T SIL – CTLRCP</td>
<td>0.47</td>
<td>4.60</td>
</tr>
<tr>
<td>I RCP T SIL IL – CTLRCP</td>
<td>0.14</td>
<td>4.30</td>
</tr>
<tr>
<td>I RCP T SIL AIL – CTLRCP</td>
<td>0.32</td>
<td>4.48</td>
</tr>
<tr>
<td>I ALB T SIL – CTLALB</td>
<td>0.23</td>
<td>4.96</td>
</tr>
</tbody>
</table>

Table 2. The DJF AMT (averaged over 65°–90°N and 400–700 hPa) and 2-m temperature response (averaged over 65°–90°N) for each experiment.
In both sets of experiments, the impact of the SST warming related to sea ice loss (Figs. 5c,f) is to weaken the westerly winds at higher latitudes and increase them at lower latitudes. The change in zonal wind speed is larger and at lower latitudes in the IALBTALB40 experiment, which is consistent with the SST forcing being stronger, particularly at lower latitudes. Despite the modest magnitude of the additional AMT warming due to the SST warming from sea ice loss compared to the AMT warming directly from sea ice loss, the zonal-mean wind response to the SST warming is similar to, or larger than, the zonal-mean wind response to sea ice loss alone. This is linked to the warming occurring at lower latitudes and higher in the atmosphere (Figs. 2 and 3). This is in agreement with idealized model experiments that have shown that the jet speed and latitude are not very sensitive to warming at the Arctic surface and that generally, the sensitivity increases as the warming moves equatorward toward the jet and higher into the atmosphere (Baker et al. 2017).

Features of the DJF sea level pressure (SLP) response are shown in Fig. 6. Both experiments with only prescribed sea ice loss (Figs. 5c,f) show a low pressure response over the Arctic Ocean, the Canadian Arctic Archipelago, and the Hudson Bay region (Figs. 6a,d). Outside these regions, the responses are quite different, but they are generally very weak (1 hPa) and not statistically significant. When the impact of changes to the extratropical SSTs is included, the SLP response becomes larger, and the two experiments show more similarities (Figs. 6b,e). Both IALBTALB40 and IRCPTSIL show a wave-1-like response, with a low pressure response over the North Pacific (i.e., a deepening of the Aleutian low) and Canada and an expansion of the Siberian high. Unlike
the response to only sea ice loss, these responses closely resemble the SLP response to sea ice loss in the coupled model experiments from which the prescribed sea ice SST forcing was derived [not shown, but see Fig. 7 of Blackport and Kushner (2017)], highlighting the importance of the extratropical SST in the AOGCM response.

Finally, some aspects of the robustness of the response were examined in more detail with the IALBTSIL experiment, which has the same extratropical SST forcing as IRCPTSIL, but with the same sea ice loss and background state as I ALBTA L B40. The DJF zonal-mean temperature, zonal-mean wind, and SLP responses to the SST forcing ($I_{ALBTSIL} - I_{ALB}$) are shown in Fig. 7. The zonal-mean temperature response is similar to the responses seen in $I_{ALBTA L B40}$ (Fig. 3c) and $I_{RCP T S I L}$ (Fig. 3d), with the warming not being trapped near the Arctic surface; however, the AMT warming is slightly weaker (0.25°C, compared to 0.45°C and 0.34°C in the $I_{ALBTA L B40}$ and $I_{RCP T S I L}$ experiments, respectively). Consistent with the weaker temperature response, the zonal wind response is also weaker, but still displays the same weakening of the jet at high latitudes and strengthening at lower latitudes. For SLP, the deepening of the Aleutian low is also seen in IALBTSIL; however, the high pressure response over northern Eurasia is significantly weaker and not statistically significant, indicating that this aspect of the response is not as robust.

4. Summary and discussion

We have examined the impact that extratropical ocean warming induced by sea ice loss has on the atmospheric response. Using coupled AOGCM simulations driven by sea ice albedo reduction and greenhouse gas–dominated radiative forcing, we diagnosed the
extratropical SST response associated with sea ice loss using two different methods. For the first method, we directly used the SST response north of 40°N from the simulations forced with sea ice albedo reduction. The second method used a two-parameter pattern scaling method (Blackport and Kushner 2017) to diagnose the part of the SST response that scales with sea ice loss, in the absence of low-latitude warming, in the two coupled ocean–atmosphere model experiments. In both cases, sea ice loss was found to warm the ocean not only in regions where sea ice had been lost, but also in regions that extend into the midlatitudes. This warming that extended into the midlatitudes was weaker using the pattern scaling method, as more of this warming was diagnosed to scale with low-latitude warming instead of sea ice loss. We note that the extratropical SST warming response we found is similar to the response found in previous sea ice loss experiments that used a dynamical ocean, but it is distinct from slab ocean model experiments (e.g., Deser et al. 2016; Wang et al. 2018). This suggests that the extratropical SST response is not simply a spreading out of the warming, which slab ocean models should capture, but that ocean dynamics plays a crucial role in the patterns of the response.

To examine the impact that this extratropical SST warming has on the atmospheric response, an AGCM was driven with reduced sea ice concentration and thickness both with and without the SST warming. In the experiments driven by only sea ice loss, the warming response was confined to the lower troposphere. In contrast, the additional warming contributed by the extratropical SST changes associated with sea ice loss was not trapped near the surface, as the AMT warming response was approximately the same as the surface warming response. Most of this AMT warming was attributed to SST warming outside the regions where sea ice loss occurred, indicating that only including SST forcing where sea ice is lost in AGCM experiments (e.g., Screen et al. 2013; Peings and Magnusdottir 2014) is not sufficient to capture this additional warming. Similar results were found in two experiments that used slightly different forcing and background states. The mechanism that connects the extratropical SST to the AMT is unknown, but may involve airmass transport along moist isentropic surfaces by synoptic-scale transient eddies (e.g., Laliberté and Kushner 2013). Using AGCM experiments, previous studies have suggested AMT warming is driven entirely by remote SST changes and not by sea ice loss (Screen et al. 2012; Perlwitz et al. 2015). Our results suggest that the warming from remote SST changes and sea ice loss cannot be so simply separated: in the coupled ocean–atmosphere system, sea ice loss can drive AMT warming through remote SST changes associated with sea ice loss.

Including the extratropical SST response from sea ice loss in the forcing was found to amplify the atmospheric circulation response to sea ice loss. In fact, the circulation response to extratropical SST warming from sea ice loss was found to be just as strong as, or stronger than, the response to the sea ice loss alone. Robust responses that were seen in each of the simulations, independent of the details in forcing and background state, include an equatorward shift and slight weakening of the eddy-driven jet and a deepening of the Aleutian low. An expansion of the Siberian high was also found in two of the simulations but not in the third, indicating some
dependence on the background state of the sea ice or SSTs. These results suggest that by neglecting extratropical SST changes, AGCM experiments may under estimate the atmospheric circulation response to sea ice loss compared to the real world, and coupled atmosphere–ocean modeling experiments are needed, in agreement with Deser et al. (2016). However, these stronger impacts of sea ice loss on the atmospheric circulation may not be seen in response to greenhouse gas forcing, as the response to warming at lower latitudes (which was not investigated in this study) may oppose the impacts of sea ice loss. Indeed, in a previous study using the same coupled ocean–atmosphere experiments used here, we found that a near cancellation of the atmospheric circulation response to sea ice loss and low-latitude warming results in a very weak response to greenhouse gas warming (Blackport and Kushner 2017). A recent study using a different model and experiment design found that this “tug of war” (Barnes and Polvani 2015) between the sea ice loss and tropical warming was more regionally dependent, featuring cancellation in the Atlantic sector but reinforcement in the Pacific sector (McCusker et al. 2017).

The wintertime atmospheric circulation response to sea ice loss found in previous AGCM experiments is generally found to be relatively weak, with disagreements between studies attributed to model differences, difference in the magnitude and location of the forcing, internal variability, and background state (e.g., Petoukhov and Semenov 2010; Screen et al. 2014; Peings and Magnusdottir 2014; Sun et al. 2015; Smith et al. 2017). The two AGCM experiments forced with only sea ice loss in this study also show weak, but different, atmospheric circulation responses, despite only small differences in forcing and background state. In contrast, recent studies examining the atmospheric circulation response to sea ice loss in coupled AOGCMs show a stronger and more consistent response between studies, despite different models and different methods to melt the sea ice (Screen et al. 2018). Our AGCM experiments forced with sea ice, together with the extratropical SST changes that we associate with sea ice loss, are able to reproduce the main features of the responses found in these AOGCM experiments. This indicates the important role that the extratropical ocean plays in determining the response to sea ice loss in the coupled climate system. In particular, the deepening of the Aleutian low that we find in response to the extratropical SST associated with sea ice loss is also found in many of the coupled ocean–atmosphere model experiments, but is not seen, or is very weak, in the AGCM experiments in this and other studies. These results are consistent with a positive feedback between the atmospheric circulation and SSTs in the North Pacific, which was also discussed by Osborne et al. (2017). We note that the tropical response to sea ice loss, which was not addressed in this study, may also impact the Aleutian low response and other aspects of the response in the extratropics either directly or by modifying the midlatitude SST response (Tomas et al. 2016; Gan et al. 2017; Wang et al. 2018). If the extratropical SST response reflects dynamical control by the tropical Pacific Ocean (Wang et al. 2018), our approach of simple scaling of the climate response with tropical mean temperature response might require further refinement.

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