Separating the influences of low-latitude warming and sea-ice loss on Northern Hemisphere climate change

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ABSTRACT

Analyzing a multi-model ensemble of coupled climate model simulations forced with Arctic sea-ice loss using a two-parameter pattern-scaling technique to remove the cross-coupling between low- and high-latitude responses, the sensitivity to high-latitude sea-ice loss is isolated and contrasted to the sensitivity to low-latitude warming. In spite of some differences in experimental design, the Northern Hemisphere near-surface atmospheric sensitivity to sea-ice loss is found to be robust across models in the cold season; however, a larger inter-model spread is found at the surface in boreal summer, and in the free tropospheric circulation. In contrast, the sensitivity to low-latitude warming is most robust in the free troposphere and in the warm season, with more inter-model spread in the surface ocean and surface heat flux over the Northern Hemisphere. The robust signals associated with sea-ice loss include upward turbulent and longwave heat fluxes where sea-ice is lost, warming and freshening of the Arctic ocean, warming of the eastern North Pacific relative to the western North Pacific with upward turbulent heat fluxes in the Kuroshio extension, and salinification of the shallow shelf seas of the Arctic Ocean alongside freshening in the subpolar North Atlantic. In contrast, the robust signals associated with low-latitude warming include intensified ocean warming and upward latent heat fluxes near the western boundary currents, freshening of the Pacific and Arctic Oceans, salinification of the North Atlantic, and downward sensible and longwave fluxes over the ocean.
1. Introduction

Arctic sea-ice loss observed by satellites over the last 40 years is one of the most obvious manifestations of greenhouse warming, and many record low extents have occurred in recent years. The question as to how a shrinking ice cover and corresponding increase in open ocean affects the atmosphere, and the mid-latitude climate in particular, remains a topic of debate [e.g., Cohen et al. (2014); Vihma (2014); Barnes and Screen (2015); Overland et al. (2015); Cohen et al. (2020); Blackport et al. (2019); Blackport and Screen (2020)]. It is difficult to disentangle the response to sea-ice loss from other concurrent forcings, from general greenhouse warming, and from natural variability. Because there is only a relatively short observational period, and the response to sea-ice loss has a low signal-to-noise ratio (Screen et al. 2014), the only way to definitively link Arctic sea-ice loss and climate responses is through dedicated modelling experiments with sufficient sampling to separate the signal from climate noise.

Earlier efforts to isolate the atmospheric response to sea-ice loss used atmospheric general circulation models forced with either high or low sea-ice cover and diagnosed the response as the difference between the resulting atmospheric patterns (Deser et al. 2010; Screen et al. 2012; Peings and Magnusdottir 2014; Sun et al. 2015). A robust warming of the lower troposphere as in Arctic amplification is found, but there tends to be a model dependence (Screen et al. 2012), and robust signals generally do not extend far beyond the high latitudes.

More recently, several studies have sought to study the coupled climate response to sea-ice loss for individual models (Deser et al. 2015, 2016; Petrie et al. 2015; Blackport and Kushner 2016, 2017; Smith et al. 2017; England et al. 2020a,b) and others have compared that response to the response to an increase in greenhouse gases under constant sea-ice area (Oudar et al. 2017; McCusker et al. 2017; Sun et al. 2018). The inclusion of a fully dynamical ocean is shown to be
crucial to obtaining the “mini-global warming” response to sea-ice loss (Deser et al. 2015; Tomas et al. 2016), whereby the zonal-mean temperature response to sea-ice loss resembles the response to greenhouse warming with amplified warming over both poles and in the tropical upper troposphere. Using an atmospheric model coupled to a slab-ocean model that includes only thermodynamic ocean feedbacks (Tomas et al. 2016; Cvijanovic et al. 2017; Wang et al. 2018; Chemke et al. 2019) creates unrealistically large warming of the Northern Hemisphere mid-latitudes (Tomas et al. 2016).

Screen et al. (2018) showed that there were common features amongst these different coupled climate model simulations when scaling the atmospheric response per unit sea-ice loss. This, alongside an indirect comparison of atmospheric response to Arctic sea-ice loss making use of CMIP5 models (Screen and Blackport 2019) showed consistency and discrepancy between coupled climate models, and here we seek to understand this.

Hay et al. (2018) was the first multi-model comparison to use two-parameter pattern scaling, as introduced by Blackport and Kushner (2017), by combining greenhouse gas forcing simulations with sea-ice loss simulations to obtain the sensitivity to sea-ice loss (SIL) and to low-latitude warming (LLW). Hay et al. (2018) focused on the near-surface atmospheric response in two coupled models and found that the sensitivity to SIL was more robust than the sensitivity to LLW.

While the use of coupled models has indicated that the ocean’s circulation is crucial to reduce the uncertainty in the atmospheric response to Arctic sea-ice loss, comparatively little research has focused on the ocean’s response itself [e.g Tomas et al. (2016); Sévellec et al. (2017); Wang et al. (2018); Liu and Fedorov (2019); Liu et al. (2019)]. The Atlantic Meridional Overturning Circulation (AMOC) is found to weaken (Sévellec et al. 2017; Liu and Fedorov 2019; Liu et al. 2019), and associated northward heat transport is reduced (Tomas et al. 2016; England et al. 2020b).
We make use of preexisting simulations to examine the inter-model and inter-method differences in the Northern Hemispheric atmospheric and oceanic responses to Arctic SIL and to LLW in five different coupled climate models: CESM1 (Blackport and Kushner 2017), WACCM4 (England et al. 2020b), CanESM2 (McCusker et al. 2017), CNRM-CM5 (Oudar et al. 2017), and GFDL-CM3 (Sun et al. 2018). This paper presents a generalization of the two model analysis of Hay et al. (2018), and presents for the first time a multi-model intercomparison of the sensitivities of the surface ocean to Arctic SIL.

This paper is organized as follows: in Section 2, we present an overview of the experiments and briefly review the two-parameter pattern scaling technique of Blackport and Kushner (2017), which yields the sensitivity patterns that we use to characterize the models. In Section 3, the atmospheric and oceanic sensitivity patterns are presented and a robust sensitivity to SIL is found in the atmosphere, while a robust sensitivity to LLW is found in the ocean. Section 4 features a more in-depth discussion of some aspects of the results, including the role of SIL in Eurasian cooling and how SIL and LLW can induce negative feedbacks on one another, each within the context of the existing literature. Finally, the results are summarized in Section 5.

2. Data and Methods

a. Models and Simulations

Due to the uncoordinated nature of these experiments, a given response in this ensemble of opportunity will depend not only on the model but also on the experimental protocol used. We characterize the robustness of the responses across model-experiments (MEs), defined as the combination of a particular model and a particular experimental protocol used to drive it (e.g. a radiative perturbation or a sea-ice perturbation, a transient versus equilibrated/timeslice simulation).
A response is the difference between a control ME and a perturbation ME. To better separate the
tropically driven component of the response from the part driven by sea-ice loss, we use two-
parameter pattern-scaling (Blackport and Kushner 2017). This yields sensitivity patterns to low-
and high-latitude drivers that are obscured by the cross-coupling in the coupled model responses.

By cross-coupling, we mean, for example, that warming at low latitudes in the model in response
to sea-ice loss can induce a response in the extratropics. With pattern scaling, we seek to separate
the direct response to each of these drivers.

The MEs, i.e. the coupled models and experimental protocols, used in this study are presented
in Table 1. For ease of reference to the previous literature, the nomenclature used in the original
publications is included in the table (second column). Five models are listed (first column and
described below) with the approaches to isolating the response to sea-ice loss from greenhouse
warming (fourth and fifth column), which are also discussed in Screen et al. (2018); Sun et al.
(2020). Another key distinction between approaches is whether transient or equilibrated solutions
are examined (sixth column). Each of the MEs provides a response which is dominated by sea-ice
loss, by the combined effect of both warming and sea-ice loss, and some provide a response to
warming in the absence of sea-ice loss.

1. **CESM1:** The configuration of the Community Earth System Model version 1 (CESM1) used
for the CESM Large Ensemble (LENS) (Kay et al. (2015), and references therein), with a
nominal resolution of 1° in both the atmosphere and the ocean, is also used here. The 40-
member ensemble mean of the CESM LENS simulations provides the forced response to
external RCP8.5 forcing, which is dominated by greenhouse warming. The sea-ice loss sim-
ulations used here are those outlined in Blackport and Kushner (2017): a control simulation
with constant year-2000 forcing is branched (at nominal year zero) from member “101” of
the LENS simulations, and years 301 to 735 are analyzed; another simulation is branched at year 301, starting from the year-2000 control at its year 301, that has the albedo of sea ice permanently reduced, and years 501 to 840 are analyzed. The albedo change applies to both Arctic and Antarctic sea ice. The difference between these two simulations gives the response to sea-ice loss in isolation. The time periods chosen for CESM LENS are those that match the Arctic sea-ice area found in the control and perturbed albedo experiments.

2. **WACCM4**: Whole Atmosphere Community Climate Model, version 4, provides improved stratospheric representation relative to the atmospheric component of CESM1, but uses previous versions of atmospheric physical parameterizations and a coarser nominal horizontal resolution of 2°. The ocean model is very similar to CESM1 LENS and the simulations are carried out at a nominal 1° ocean resolution. A small ensemble of three CMIP5 RCP8.5 simulations are used to describe projected greenhouse gas forcing (Marsh et al. 2013). The CONTROL and ARCTIC simulations used here are described in England et al. (2020b): These are analogous to the simulations performed with CESM1, however they differ in length, the control climate state, and in the sea-ice loss protocol. The longwave forcing method of Deser et al. (2015) is used to emulate the sea-ice extent in the Arctic to the mid-20th century in CONTROL, and the end of the 21st century for ARCTIC. For each, the last 250 years of a 350-year simulation are retained for analysis. The time periods chosen from the CMIP5 RCP8.5 WACCM simulations are those from which the sea-ice targets are taken.

3. **CanESM2**: These simulations, outlined in McCusker et al. (2017), use the second generation of the Canadian Earth System Model at spectral horizontal resolution T63. These MEs consists of four simulations in which the Arctic sea-ice mass is nudged grid-cell-by-grid-cell to match that from equilibrated preindustrial CO$_2$ ($I_{PI}$) or 2×CO$_2$ ($I_{2X}$) sea-ice distributions of
the free-running CanESM2 model. For each of these sea-ice distributions, the radiative forcing is set to either preindustrial (PI) or $2 \times CO_2$ (2X), and the model is integrated for 201 years when the radiative forcing and sea-ice distributions are matched, and 300 years when they are mismatched, of which the last 200 years are retained for analysis. The greenhouse warming-forced response is given by the difference between the $C_{2X}I_{2X}$ and $C_{PI}I_{PI}$ simulations, while the response to sea-ice loss can be isolated at both cooler and warmer background climates by using the other two simulations. Additionally, the response to warming in the absence of ice loss at high or low ice cover can be explicitly determined for this ME, unlike for the CESM1 and WACCM4 MEs.

4. **CNRM-CM5**: These simulations are outlined in Oudar et al. (2017). The atmospheric and ocean components of this version of the model have a horizontal resolution of $1.4^\circ$ and $1^\circ$, respectively. These MEs consist of four simulations that are analogous to those performed with CanESM2: they utilize a non-solar flux correction to the ocean model to control the seasonal cycle of Arctic sea ice to match that from the late 20th century or the late 21st century resulting from historical or RCP8.5 radiative forcing, while the radiative forcing itself is set to either late 20th century or late 21st century. Each integration is 200 years in length and the last 100 years are retained for analysis. Again, this experimental design allows us to also explicitly assess the response to greenhouse warming in the absence of sea-ice loss.

5. **GFDL-CM3**: Version 3 of the Geophysical Fluid Dynamics Laboratory Coupled Model has an atmospheric resolution of approximately 200km, while the ocean component has a latitudinal resolution of $1^\circ$ and enhanced longitudinal resolution reaching $1/3^\circ$ near the equator. The experiments outlined here are those of Sun et al. (2018): five ensemble members of an integration representing 1970 to 2090 in which ice is unconstrained and radiative forcing is
given by historical and RCP8.5 forcing provides the forced response to greenhouse warming. Another experiment of five ensemble members whereby ice is constrained using a volume-nudging method similar to the mass-nudging method of McCusker et al. (2017) to match its mean distribution from 1970-1990. This experiment is integrated from 1990 to 2090 to provide the transient response to RCP8.5 radiative forcing in the absence of sea-ice loss. The difference between the ensemble means then gives the transient response to sea-ice loss.

The way the responses are diagnosed from the MEs are shown in Table 2, retaining the nomenclature from the original publications. We separately group the simulations that are run to equilibrium with unchanging sea ice and radiative forcing and those that are integrated with transient forcing. We plot annual mean Arctic sea-ice area versus low-latitude sea surface temperature (area-weighted average over 0 to 40°N) in Figs. 1a and b. The transient simulations in Fig. 1b show 15-year running means as well as the periods chosen in determining the response. Each response that we diagnose is then shown in Fig. 1c on a plot of Arctic sea-ice loss and low-latitude warming, where the responses are colour-coordinated to distinguish between the three types of responses: warming under near-constant sea-ice area, for which 2.0-3.0°C of low-latitude warming is obtained with less than one million km² of Arctic sea-ice loss; sea-ice loss under near-constant low-latitude warming, for which less than 0.5°C of low-latitude warming is obtained with 3.0-8.0 million km² of sea-ice loss; and full greenhouse warming, for which about 1°C of low-latitude warming is obtained per 2 million km² of Arctic sea-ice loss.

b. Pattern Scaling

There is no consistent framework upon which these MEs are designed and as a result there are differing amounts and patterns of sea-ice loss and low-latitude warming. We show, for each season, the sea-ice fraction in each ME’s low sea-ice simulation relative to the 15% sea-ice extent
contour in its high sea-ice simulation in Fig. 2. This figure highlights the disparate patterns of
sea-ice loss across the MEs considered in this study. To address the inconsistencies in the MEs
and their responses arising from differences in experimental design, we use two-parameter pattern-
scaling (Blackport and Kushner 2017). This linear decomposition removes the signal of LLW in
the pattern of sensitivity to SIL, thus yielding the direct sensitivity to SIL while accounting for the
cross-coupling from LLW. It yields also a direct sensitivity to LLW by removing the signal of SIL.

Besides attempting to consistently scale for the amount of Arctic sea-ice loss, pattern scaling
also seeks to simultaneously scale for differences in tropical warming responses that might arise
from radiative feedbacks (e.g. in the case of the CESM1 albedo-reduction ME due to the additional
effects of Antarctic sea-ice loss on the tropics (England et al. 2020a).)

The method used is the same two-parameter pattern-scaling of Blackport and Kushner (2017)
and Hay et al. (2018); we present here a brief review. From the empirical observation that the
pattern of the forced response to greenhouse warming in climate models is proportional to the
global mean temperature (Santer et al. 1990; Tebaldi and Arblaster 2014) and is independent of
the details of the forcing, Blackport and Kushner (2017) generalized pattern scaling to multiple
climate variables. As in that study, we use low-latitude (0-40°N) sea surface temperature ($T_l$) and
Arctic sea-ice area ($I$) as the scaling variables. The response pattern, $\delta Z_m$, of some variable $Z_m$ can be decomposed as:

$$\delta Z_m = \left. \frac{\partial Z}{\partial I} \right|_{T_l} \delta I_m + \left. \frac{\partial Z}{\partial T_l} \right|_{I} \delta T_{l,m}. \tag{1}$$

For a given set of MEs and a given season, we choose two response patterns and invert (1), solving
for the sensitivities $\left. \frac{\partial Z}{\partial I} \right|_{T_l}$ and $\left. \frac{\partial Z}{\partial T_l} \right|_{I}$. These sensitivities give the patterns of response that scale with
LLW in the absence of SIL, and with SIL in the absence of LLW, respectively. They are further in-
terpreted as characterizing the model response, independently of the experimental protocols in the
original MEs. For the models CESM1, WACCM4, and GFDL-CM3, we have \( m = 2 \) each and so obtain one pair of sensitivity patterns for each of these models. For CanESM2 and CNRM-CM5, we have \( m = 5 \) (the responses to sea-ice loss under different warm and cold climate conditions, the responses to increasing greenhouse gases under high and low sea-ice cover, and the combined response to sea-ice loss and increased greenhouse gases) and so we can obtain two independent pairs of sensitivity patterns each.

The method is demonstrated in Fig. 3 for the zonal-mean temperature response in boreal winter in CNRM-CM5, using the nomenclature of Oudar et al. (2017). Five response patterns for this ME are shown: in Fig. 3a, “ICE+GHG”, which is the response to greenhouse warming; Fig. 3b, “GHG Effect 21”, the response to greenhouse warming at fixed late 20th century ice cover; Fig. 3d, “ICE Effect 20”, the response to sea-ice loss under late 20th century greenhouse gas (GHG) concentrations; Fig. 3f, “GHG Effect 20”, the response to greenhouse warming at late 21st century ice cover; and Fig. 3h, “ICE Effect 21”, the response to sea-ice loss under late 21st century GHG concentrations. Fig. 3a shows the usual pattern of projected greenhouse warming with an overall warming of the troposphere at all Northern Hemisphere latitudes with amplified warming in the Arctic lower troposphere (Arctic amplification) and in the tropical upper-troposphere, alongside cooling in the stratosphere. In Figs. 3b and f, when GHG forcing is applied with sea ice held fixed, the pattern closely resembles that in Fig. 3a but it lacks surface Arctic amplification and instead shows Arctic warming with a weak maximum in the mid-troposphere. Interestingly, in Fig. 3f, there is a hint that greenhouse gas radiative forcing under prescribed high ice conditions gives rise to surface cooling, something that was found after the pattern-scaling decomposition for CESM1 in Hay et al. (2018). This weak but significant cooling response to GHG forcing with fixed ice cover may arise as an artifact of the sea-ice nudging. In Figs. 3d and h, when ice loss alone is imposed, an Arctic amplification pattern is seen, alongside a small amount of warming in the
tropical mid-troposphere, particularly in “ICE Effect 21”. This pattern was termed a “mini-global warming” by Deser et al. (2015) due to its resemblance to the pattern of warming in the tropics in Fig. 3a.

After the pattern-scaling decomposition we obtain the sensitivity patterns of this ME to LLW, \( \frac{\partial T_{\text{zon}}}{\partial T_l} |_{T_l} \), for each background climate in Figs. 3c and g and the sensitivity patterns to SIL, \( \frac{\partial T_{\text{zon}}}{\partial I} |_{T_l} \), for each background climate in Figs. 3e and i. We note the similarities of the sensitivities between background climates. Pattern-scaling has removed the mid-tropospheric tropical warming seen in Fig. 3h from the modelled sea-ice loss response pattern, and the Arctic amplification is slightly shallower, suggesting that some Arctic midtropospheric warming scales with LLW. The pattern of sensitivity to LLW comprises all the tropical warming as well as warming of the mid-troposphere in the Arctic, with cooling found both at the surface and in the stratosphere. Because the stratospheric response to climate change is dominated by direct CO\(_2\) radiative forcing, a process that is not at play in simulations forced by sea-ice loss even with low-latitude warming, caution is required in interpreting sensitivity patterns there. The amount of warming that occurs near the Arctic surface in Fig. 3b or Fig. 3h is greater than that in Fig. 3a, thus it appears that in certain models a modest negative feedback near the Arctic surface arises, i.e. a minimum in surface warming or even a cooling, which scales with LLW.

3. Results

a. Patterns of Atmospheric Sensitivity

We first extend the analysis of Hay et al. (2018), which included the same pattern-scaling decomposition for two of the MEs considered here, CESM1 and CanESM2, to include more MEs and zonal-mean climate variables as well as the near-surface variables investigated in that study.
The left-hand column of Fig. 4 presents the mean boreal winter (DJF) map of the sensitivity to Arctic SIL, per two million km$^2$ of ice loss, for surface temperature, sea level pressure, precipitation, and 850hPa zonal wind for all MEs, while Fig. 5 shows the same for zonal-mean zonal wind and and zonal-mean temperature. The right-hand column of Fig. 4 and 5 shows the sensitivity to LLW, per degree of warming. Scaling per two million km$^2$ of ice loss and per degree of warming arises from the observation that these models show roughly one degree of tropical warming per two million km of sea-ice loss in response to greenhouse warming (Fig. 1). To obtain Fig. 4 and 5, we perform the pattern-scaling decomposition for each ME using the bolded and italicized combinations in Table 2. This results in patterns for each of the sensitivities and for each of the MEs. These patterns are then regridded onto a $1^\circ$ by $1^\circ$ global grid, and averaged across the set of MEs. The hatched areas in Fig. 4 and 5 represent those regions where at least four of the five models agree on the sign of the sensitivity pattern, and the number in the bottom right-hand corner indicates the median spatial correlation of all MEs with each of the rest of the set for the Northern Hemisphere extratropics (defined here to include all regions north of 30$^\circ$N). The number between panels indicates the spatial correlation between the mean sensitivity to SIL and to LLW. A more negative number indicates that the sensitivity to SIL acts as a negative feedback on the sensitivity to LLW.

For comparison of how the sensitivity patterns differ from their response pattern counterparts, Fig. 6 shows the same variables’ shown in Fig. 4 and 5 responses to sea-ice loss in the upper row, responses to greenhouse warming with fixed sea ice (for the models that provide this) in the middle row, and responses to greenhouse warming with freely evolving sea ice in the lower row.

The surface temperature sensitivity to Arctic SIL in DJF, shown in Fig. 4a, consists of a strong warming directly over the Arctic Ocean and Hudson Bay of more than two degrees per million km$^2$ of ice loss, as well as a more modest warming over adjacent high-latitude land areas and
over the mid-latitudes of North America. A weak cooling pattern over eastern Eurasia is seen in
the multi-model mean, but it is not robust, save for a region of Eastern China and Japan. This
cooling is a part of a robust east-west gradient of cooling to warming over the Pacific ocean, and
is associated with the SST sensitivity pattern that resembles a positive Pacific Decadal Oscillation
(PDO) pattern. Compared to the response patterns shown in Fig. 6, pattern-scaling attributes much
of the mid-latitude warming to LLW and not directly to SIL.

The pattern of surface temperature sensitivity to LLW (Fig. 4b) correlates negatively with that
the pattern of sensitivity to SIL ($r = -0.57$) and consists of a robust warming everywhere south of
the Arctic circle, excluding the subpolar North Atlantic. There is enhanced warming over Siberia
and the Rocky mountains. The temperature sensitivity over the Arctic is not robust, with three of
the five models showing some cooling, as was seen in the zonal-mean temperature sensitivity to
LLW in CNRM-CM5 in Fig. 3c. We note that there is little consistency over the subpolar North
Atlantic in either of the two sensitivity patterns. This region also does not show a robust sign in the
modelled responses (i.e., before we scale the patterns) to greenhouse warming, warming without
ice loss, or sea-ice loss (Figs 6a, g, and m). This lack of consistency is related to inter-model
differences in the location of the North Atlantic ‘warming hole’, differences that are reflected in
the ocean response below.

To obtain a more complete picture of the robustness across MEs, Fig. 7 presents each of the
inter-ME spatial correlations for all the seasons and the annual mean. The result is a distribution
of inter-ME correlations for the sensitivities to SIL and LLW in blue and orange, respectively. The
black bars represent the median of the correlations. A cluster of dots near $r = 1.0$ implies that
all MEs agree on the sensitivity pattern, a large spread in dots implies that some MEs are highly
correlated with one another while some are not (or are anti-correlated), and a clustering of dots
near a lower $r$ implies that the sensitivity patterns are not robust.
The pattern correlations in Fig. 7a show that the patterns of extratropical surface temperature sensitivities to both SIL and LLW are robust throughout the year except in boreal summer, when correlations between sensitivities to SIL are at a minimum. We expect that the sensitivity to SIL will be most robust in DJF (Hay et al. 2018; Screen and Blackport 2019) due to the greatest turbulent heat fluxes in response to sea-ice loss occurring then, despite there being the largest ice losses in late summer and early autumn (Deser et al. 2010). Consequently, sea-ice-loss forcing in JJA is weaker than in DJF and there is less consistency in model responses and sensitivities. We note that the larger differences in sea-ice area shown in Fig. 2 in JJA, with one ME nearly completely ice-free, may also be a confounding factor in the sensitivity comparison in this season.

The pattern of sea level pressure sensitivities are shown in Figs. 4c and d, and the responses in Figs. 6b and h. The sensitivity patterns are found to have opposite sign across most regions, with a correlation of $r = -0.61$, suggesting that sea-ice loss acts as a negative feedback on the sea level pressure response arising from the tropics. A robust sensitivity to SIL emerges with a deepening of the Aleutian Low and lowered pressure from the Canadian Basin to Hudson Bay, alongside strengthening of the Siberian High. The latter response has been associated with sea-ice loss in the Barents and Kara Seas in atmosphere-only modelling studies (Mori et al. 2014), but whether the same dynamical mechanism is at play in coupled modelling studies remains an open question. There is less consistency in the latitudinal extent of the change to the Aleutian Low, which contributes to some of the spread in correlations in Fig. 7b. Differences in the Aleutian Low response were attributed to differences in tropical precipitation in Deser et al. (2016), which may indeed be the case here, as the precipitation sensitivity to SIL can be seen in Fig. 8 and we note little inter-ME agreement in the tropics.

The Siberian High response to LLW is a weakening with increasing temperature, and this weakening directly opposes the strengthening of the Siberian High associated with the warm-Arctic-
cold-Eurasia (WACE) pattern (Mori et al. 2014). This regional tug-of-war suggests that changes to the Siberian High associated with the WACE pattern would be difficult to observe in a warming world in the presence of both tropical and Arctic warming. Apart from this signal, the entirety of the Western Hemisphere’s sensitivity pattern is not robust to LLW. The Aleutian Low either deepens or weakens depending on the ME in question, as was seen in Hay et al. (2018), and this is the reason for the negative pattern correlations seen in DJF in Fig. 7b. The sensitivity pattern in the Atlantic sector is not particularly robust to either forcing.

There is a large spread in the inter-ME correlations of LLW sensitivity patterns in DJF, and to a lesser extent in the other seasons, and the largest spread in correlations between the SIL sensitivity patterns in in JJA. We find, as in the surface temperature sensitivity, that the most robust sensitivity to LLW occurs in JJA, which is the same season for which we find the least robust sensitivity to SIL, as indicated by the median in correlations.

The precipitation sensitivity patterns are shown in Figs. 4e and f and in Fig. 8, and the response patterns are shown in Figs. 6c, i, and o. Although the sensitivity patterns are highly structured, pattern-scaling helps to normalize and remove some of the noisiness from the response over the Arctic Ocean. There are regions of consistently signed sensitivities to both SIL and to LLW. The global precipitation sensitivity patterns for individual models (Fig. 8) confirm the overall consistency between the models summarized in these panels in the Northern Hemisphere for the annual mean, but the sensitivity to SIL is model-dependent in the tropics, as mentioned above.

In Fig. 4e and in each individual model in Fig. 8, an increase in precipitation over the Arctic Ocean is seen as a result of increased evaporation from the ocean surface under SIL (Bintanja and Selten 2014). Additionally, consistent with the strengthening of the Aleutian Low and the inter-ME spread in that sensitivity pattern, an increase in precipitation on the western coast of North America is seen. However, its exact latitudinal position is dependent on the ME. In Fig.
f, the sensitivity to LLW shows a robust increase of precipitation over much of the mid-latitudes, including Western Europe, northern Eurasia, and both coasts of North America, and a drying over the subtropical eastern Atlantic and Mediterranean. We find median spatial correlations are lower overall than they were for surface temperature and sea level pressure, ranging from $r = 0.1$ to $r = 0.5$ for each collection of sensitivities. The spatial correlations across MEs are generally greater for the response to LLW than to SIL (Fig. 7c).

The sensitivity to SIL of the 850hPa zonal wind over the Pacific Ocean (Fig. 4g) is largely consistent with the sea level pressure sensitivity. It includes a general weakening on the poleward flanks of the climatological maximum jet, indicating a southward shift in the Pacific storm track, as expected in response to a decrease in the equator to pole temperature gradient via the thermal wind relationship and under a reduction in baroclinicity. A similar pattern is seen in the Atlantic, and though it is not as robust in the sensitivity pattern as it is in the response pattern (Fig. 6d), resembles that found in Peings et al. (2019) for simulations dominated by the effects of Arctic amplification rather than upper tropospheric warming. In most seasons we see a large spread in correlations, and, once again the median value of correlation is lowest in JJA. Looking at the sensitivity of zonal-mean zonal wind in a latitude-height cross-section (Fig. 5c), a robust weakening and equatorward intensification is seen, consistent with the dominant pattern in U850. This is in agreement with other studies (Deser et al. 2015; Blackport and Kushner 2017).

The midlatitude U850 and tropospheric zonal-mean zonal wind sensitivity to LLW (Figs. 4h and 5d) is, on average, of opposite sign ($r = -0.71$ and $r = -0.36$, respectively) to the SIL sensitivity, so that LLW results in poleward intensification of the jet, as expected for an increase in the equator to pole temperature gradient aloft, and in good agreement with the idea of a “tug-of-war” between high and low latitude forcing (Harvey et al. 2014; Barnes and Screen 2015; Peings et al. 2019). The competing influences of high- and low-latitude forcing are more clear in the sensitivity patterns
than they are in the response patterns of Fig. 6e and k because pattern scaling has removed the
effects of cross-coupling. This sensitivity is not robust in DJF at higher altitudes in the zonal
mean, due to differences in the Atlantic and Pacific basins. We do, however, find better inter-ME
agreement for a poleward intensification of the jet in other seasons, whereas in DJF, the most
robust part of the low-latitude response is the strengthening of the subtropical jet.

Lastly, we examine the mean sensitivity patterns of zonal-mean temperature in Figs. 5a, b. The
sensitivity to SIL is that of Arctic amplification and is robustly seen in all models. The latitudinal
and vertical extent of the warming is from 45°N and up to 400hPa is significantly reduced com-
pared to the response in Fig. 6f. A weak but robust cooling pattern in the mid troposphere and
warming above 200 hPa in the mid-latitudes is also found. The sensitivity to LLW is a robust
warming throughout the troposphere, intensified at the tropical upper troposphere. In the Arctic
lower troposphere we note again that the MEs disagree on on the sign of the sensitivity. Strato-
spheric cooling of the Arctic is robustly associated with LLW, while the stratospheric sensitivity
to SIL is not robust, in agreement with the indirect method of Screen and Blackport (2019). Inter-
ME spatial correlations in zonal-mean temperature are all closely clustered above $r = 0.8$ (Fig. 7f)
except for the SIL sensitivities in JJA that have a large spread in correlations.

Overall, in DJF (and to a lesser extent in MAM, SON, and ANN), the patterns of atmospheric
sensitivity to SIL are more consistent between MEs than to LLW, for the near-surface climate
variables (surface temperature, sea level pressure, and 850hPa zonal wind), as evidenced by the
median of the correlations shown by the black bars in Fig. 7. This reverses in JJA when the
sensitivity to LLW is more robust in all variables. Therefore, the conclusions from Hay et al.
(2018) generally hold in this larger pool of models and sea-ice loss protocols.

Before carrying out an analysis of the ocean’s surface sensitivity, we decompose surface heat flux
responses over the ocean (Fig. 9). A positive flux is defined as in to the atmosphere. The sensible
heat flux sensitivity to SIL (Fig. 9a) is positive where sea ice is lost in the Arctic, and negative equatorward of the climatological sea-ice margins. Robustness is confined to the Arctic Ocean itself, as well as weak but robust positive heat flux south of the Kuroshio Extension. The former arises when sea ice is lost, exposing the ocean to the atmosphere and driving a greater transfer of heat from the ocean to atmosphere. Equatorward of the sea-ice margin, atmospheric warming from SIL is leading to anomalous warming of the ocean from downward heat fluxes (Deser et al. 2010).

On the other hand, the latter result of upward sensible heat south of the Kuroshio is curious because the surface ocean (see Fig. 10 below) and overlying atmosphere (Fig. 4a) both exhibit cooling as the sensitivity to SIL in the same region. Inter-ME agreement in DJF is not as great as for some of the other surface variables, with the median $r = 0.48$. Here, the details in the differences of the patterns of sea-ice loss (Fig. 2) may be driving the lower pattern correlation. Over the ocean, the sensible heat flux sensitivity to LLW (Fig. 9b) is downward everywhere, suggesting that the atmospheric warming of tropical origin consistently drives anomalous ocean warming throughout the Northern Hemisphere. While the inter-ME median $r$ is similar, the sign of the sensitivity is robust nearly everywhere. The sensitivities are moderately negatively correlated with each other.

We find a similar pattern in the latent heat flux sensitivity to SIL (Fig. 9c), with an upward flux where sea-ice is lost and downward flux equatorward of the sea-ice loss region, but inter-ME agreement is somewhat weaker, with median $r = 0.36$. Regions of agreement are confined to higher latitudes in the sensitivity to SIL. On the other hand, the sensitivity to LLW (Fig. 9d) is robustly positive over all regions but the Arctic Ocean and the subpolar North Atlantic.

The longwave flux sensitivity to SIL (Fig. 9e) exhibits a pattern of positive Arctic amplification, while the sensitivity to LLW (Fig. 9f) is a robust moderately negative signal that is likely associated with poleward heat and moisture transport (Lee et al. 2019). The negative heat flux sensitivities to
LLW over the Arctic Ocean are consistent with the cooling signal some MEs exhibit there in the surface temperature sensitivity.

Overall, in contrast to the atmospheric variables presented in Figs. 4 and 5, we find that, for surface fluxes, sensitivities to SIL are overall less robust than sensitivities to LLW. We can infer that the heat flux responses are more strongly controlled by the pattern of sea-ice loss, whereas the nature of the atmospheric response depends less on the details of the sea-ice loss. However, as for the atmospheric response, there is a generally a negative correlation between the mean sensitivity to SIL and to LLW.

b. The ocean’s sensitivity

The pattern-scaling decomposition (Hay et al. 2018), as well as other studies comparing the atmospheric responses to sea-ice loss in coupled models (Screen et al. 2018; Screen and Blackport 2019), indicate that coupling to the ocean is an important part of the surface climate response. To elucidate the role of the ocean, we decompose the surface ocean response using pattern scaling, as we did for the atmosphere. We present the results for sea surface temperature (SST) in Fig. 10 and for sea surface salinity (SSS) in Fig. 11. Due to the greater inconsistency in the sensitivity patterns on regional scales compared to the atmosphere, we present not only the multi-ME mean as above, but also the individual MEs.

McCusker et al. (2017) demonstrated, with the CanESM2 MEs we use here, that there is separability and additivity of atmospheric response patterns to a doubling of CO$_2$ without sea-ice loss and to Arctic sea-ice loss under constant background CO$_2$. We have confirmed additivity in a second model by using the analogous CNRM-CM5 simulations for the atmospheric response. For example, McCusker et al. (2017) found that 98% (90%) of the surface temperature (sea level pressure) variability is explained by the sum of the sea-ice loss and CO$_2$ forced patterns, and in
CNRM-CM5 we find the percentage of variability explained to be 99% for both surface tempera-
ture and sea level pressure. We applied the same analysis to the surface ocean responses and find
that 79% of the SST variability and 92% of the SSS variability is explained by the sum of the
responses in CanESM2, while in CNRM-CM5 the percentage of variability explained is 92% and
87%, respectively. These results justify the applicability of the linear pattern-scaling framework
to our surface ocean analysis.

The sensitivity of annual mean SST to SIL for individual MEs and for the multi-ME mean, as
shown in the left hand column of Fig. 10 and in Fig. 10k, respectively, indicate that the ocean’s
surface warms robustly where sea ice is lost and solar radiation is absorbed by the ocean surface,
as well as where turbulent fluxes are in to the ocean (Figs. 9a, c). The sensitivity pattern in
the Pacific presents as warming along the west coast of North America alongside a cooling that
extends out across the western Pacific that resembles the positive phase of the PDO. Such a pattern
is consistent with what is seen in the surface temperature sensitivity (Fig. 4a), but opposite that
seen in the sensible and latent heat flux sensitivities (Figs. 9a, c), as previously noted. This robust
pattern, found in all MEs to varying degrees, if driven by similar processes that drive variability
in the PDO, results as a complex interplay of both local and remote atmospheric and oceanic
processes (Newman et al. 2016). All MEs show at least some cooling in the North Atlantic that
scales with SIL, although the location is not robust and is likely influenced by the mean states of
the models, including biases and regions of deep ocean convection. Subpolar cooling relative to
the global mean has been identified as signature of AMOC weakening (Rahmstorf et al. 2015),
and indeed all models considered here also simulate a weakening of the overturning circulation as
a response to sea-ice loss (not shown), in agreement with Sévellec et al. (2017); Liu and Fedorov
(2019); Liu et al. (2019).
The second column of Fig. 10 shows each ME’s sensitivity to LLW, and the multi-ME mean is shown in Fig. 10l. As in the atmosphere, there is disagreement in the sign of the response in the Arctic, but otherwise there is a robust pattern of warming in most locations. The North Atlantic subpolar gyre region exhibits a “warming hole” in all models but CanESM2. The warming in the mean sensitivity pattern is intensified in both the Gulf Stream and the Kuroshio current relative to other regions, but we note considerable variability between MEs (e.g. CanESM2 does not exhibit intensified warming of the Kuroshio region, and CNRM-CM5 does not exhibit intensified warming of the Gulf Stream.)

We calculate inter-ME spatial correlations for the Northern Hemisphere oceans north of 30°N, as in Section 3a (Fig. 12a), but we also subdivide that area into the three ocean basins. Annual mean correlations calculated over the North Atlantic and the North Pacific, for 30°N–60°N, and the Arctic Ocean, north of 60°N, are shown in Fig. 12c. There is little seasonal variation in the inter-ME correlations (Fig. 12a), especially compared to atmospheric surface temperatures (Fig. 7a) due to slower timescales of evolution in the ocean. The exception to small inter-seasonal variability is a peak in correlations amongst the SIL sensitivity patterns in JJA and SON in the Atlantic compared to the other seasons (not shown). We reason that this is because the greatest forcing by sea-ice loss on the ocean coincides with the seasons in which sea-ice loss is greatest and is driven by the shortwave flux, unlike for the atmosphere whose response lags by at least a season and is driven by turbulent heat fluxes.

The median of inter-ME correlations for the Northern Hemisphere sensitivity to SIL are $r = 0.40$, while in the Pacific $r \approx 0.75$ year-round. In the Atlantic, the most robust season is JJA where the median $r = 0.7$, while the colder seasons have lower correlations of $r = 0.25$. The median of inter-ME correlations in the sensitivity to LLW is greater in all seasons for the Northern Hemisphere. In the Atlantic and Pacific basins, the dots are clustered into two groups, where the lower correlations
are the result of the CanESM2 sensitivity pattern in the Pacific, and the GFDL-CM3 sensitivity pattern in the Atlantic, as noted above. In contrast to the atmospheric sensitivity patterns, but in agreement with the surface flux variables, the sensitivity patterns to LLW are on average more robust than the SIL sensitivity patterns. In the Arctic, we see a considerable spread in correlations, which might arise because these patterns are more closely tied to the details of ice-loss here, which are in turn dependent on the ME (Fig. 2). This region is directly affected by the sea-ice loss protocol, particularly in the MEs that introduce perturbations into the ocean either directly (e.g., CNRM-CM5) or indirectly through nudging the sea ice (WACCM4, CanESM2, GFDL-CM3).

The patterns of SSS sensitivities (Fig. 11) show that the sensitivity to LLW is more consistent across MEs (Fig. 11k) than the sensitivity to Arctic SIL (Fig. 11l). The agreement in the sensitivity to SIL is mainly confined to the Arctic, where there is a robust freshening of the central Arctic ocean and salinification around the continental shelves, which may be indicating increased brine rejection as a consequence of more seasonal sea-ice formation along the shelves. Outside the Arctic, we note a freshening in the extension of the Gulf Stream. On the other hand, the sensitivity to LLW indicates a clear and consistent freshening in the Pacific and Arctic Oceans that extends into the subpolar North Atlantic, and salinification of the rest of the Atlantic and the Mediterranean. This pattern reflects both observed and projected changes in salinity, which can be understood through the intensification of the hydrological cycle and manifests as an amplification of the mean pattern of salinity (Durack and Wijffels 2010; Skliris et al. 2014). Freshening of the Arctic Ocean in the sensitivity to LLW occurs in absence of increased precipitation (see Fig. 4f), suggesting that it may be explained through an ocean pathway where the currents carry relatively fresh Pacific Water through the Bering Strait, where they remain near the surface (Aagaard and Carmack 1989; Serreze et al. 2006), or through increased precipitation over land, which scales with LLW (Fig. 4f), increasing river runoff into the Arctic Ocean (Nummelin et al. 2016).
Correlations over the Northern hemisphere as a whole (Fig. 12b) show minimal seasonality and reflect the relative robustness of the LLW sensitivity pattern compared to the SIL sensitivity pattern. The median correlations in the Pacific are small for both sensitivity patterns, so while the sign of the pattern is consistent in the sensitivity to LLW, this reveals that the details of the pattern of freshening are not particularly robust. In the Arctic, despite a consistent sign in the sensitivity, as for SST, the low correlations amongst the patterns of sensitivity to SIL reflects the differing patterns of sea-ice loss and sea-ice loss protocols.

4. Discussion

Some aspects of the results warrant a more in-depth discussion. Firstly, whether cooling of Eurasia is a forced response to Arctic sea-ice loss in model experiments remains a topic of debate (Cohen et al. 2014; McCusker et al. 2016; Sun et al. 2016; Blackport et al. 2019; Mori et al. 2019; Cohen et al. 2020; Labe et al. 2020), with these results suggesting it may play a minor role in the sensitivity to SIL in some models, as seen in Fig. 4a. However, this pattern of cooling is not present in the coupled model response to Arctic sea-ice loss (see Fig. 6a) and emerges only after we remove the back-effect of LLW. In particular, Figs. 4 c, d suggest that forced LLW acts to weaken the Siberian High in opposition to the forced response to SIL, which strengthens the Siberian High and cools by advection. Deser et al. (2016) found that coupling reduced Eurasian cooling compared to equivalent experiments run with an atmosphere-only model, a result also found comparing the atmosphere-only runs in England et al. (2018) with the coupled runs of England et al. (2020b,a). However, the Siberian High was found to be stronger in the coupled runs, and so the reduced cooling is not a dynamical response to LLW found in the coupled runs. Rather, it is likely related to the background mini global warming (which scales with LLW), or the stronger Arctic warming from the coupling advected over the continent. Beyond regional
circulation changes, broader hemispheric warming coherent with LLW appears, in the models, to overwhelm any Eurasian cooling induced by SIL. Thus, if the models represent the global warming process accurately, such cooling is not likely to be observed as a long-term mean climate response under greenhouse warming. While these results on their own do not rule out the possibility that sea-ice loss could enhance cold extremes, only that we should not expect cooling on average, other modelling studies have pointed to the reduced risk of cold extremes in response to sea-ice loss because of reduced subseasonal variability (Screen et al. 2014, 2015; McCusker et al. 2016; Blackport and Kushner 2017; Collow et al. 2019). An additional note on this topic is that the model with the strongest cooling sensitivity in Eurasia is W ACCM4 (not shown), suggesting the potential importance of stratospheric dynamics in the development of that pattern of response, as has been previously found (Zhang et al. 2018). Lastly, recent work (He et al. 2020; Labe et al. 2020) has pointed out that the depth of Arctic amplification, which is not captured by our surface-parameter based pattern scaling approach, may be crucial to simulating Eurasian cooling. However, these studies do not suggest that the depth of Arctic warming is necessarily linked to sea-ice loss, and we can not rule out that the upper tropospheric warming arises from low-latitude warming.

Eurasian cooling from SIL is an example of “tug-of-war”-type responses where sea-ice loss acts as a negative feedback on global warming (Harvey et al. 2014; Barnes and Polvani 2015; McCusker et al. 2017). Many of the mean sensitivity patterns were found to be negatively correlated with one another, such as the sea level pressure responses over Eurasia just described, and similarly opposite responses in lower tropospheric winds in the west Pacific and in the zonal-mean zonal winds. For example, all MEs indicate a strengthening of the Siberian High in the sensitivity to SIL and a weakening of U850 winds in the Pacific (a weakening/equatorward shift of the Pacific Storm track), a weakening of the Siberian High and a poleward shift of the Pacific storm track in the sensitivity to LLW. Other negative feedbacks are found only in certain models, such as the
finding that the sensitivity of the Aleutian Low in CESM1 strengthens with SIL and weakens with LLW. While we do find overall that sensitivity patterns tend to be negatively correlated, we note that not all aspects of the patterns are robust in the multi-ME mean.

Lastly, the curious cooling of atmospheric surface temperature directly above the Arctic Ocean that scales with LLW is found in three of the five MEs considered here: CESM1, WACCM4, and CNRM-CM5. For the former two MEs, this counterintuitive result suggests that, per unit sea-ice area loss, there is more warming in the dedicated sea-ice loss experiment than in the transient RCP8.5 simulations. Because the sea-ice distribution in WACCM4 is matched to that from the RCP8.5 simulation, it cannot be attributed to large differences in sea-ice area (although small differences could be a factor), discrepancies in sea-ice thickness may arise since it is not explicitly controlled for (England et al. 2020b). The two models that have Arctic warming that scales with LLW use similar nudging protocols to lose sea ice and better match both thickness and area. It is therefore uncertain whether this Arctic cooling signal is a real physical response, an artifact of the ice loss protocol, or an artifact of the pattern-scaling decomposition. All of these possibilities warrant further investigation with, for example, an approach involving application of multiple ice-loss protocols, such as nudging and ghost forcing, in a single model to better elucidate mechanisms.

5. Conclusion

We have explored available simulations from five different coupled models designed to isolate the coupled climate response to sea-ice loss as well as those that simulate the response to greenhouse warming. First, we have decomposed the patterns of response into sensitivity patterns that scale with Arctic SIL and with LLW separately. Our results largely confirm the conclusions of Hay et al. (2018), which focused on just two of the model experiments used here. We find that near-surface atmospheric sensitivity to SIL is surprisingly robust despite differences in sea-ice loss
protocols, background climate, and model physics. It consists of warming directly over the Arctic
Ocean and over the high-latitude land masses that extends up to 400hPa, a dipole in sea level pres-
sure with lower pressure over North America and high pressure over Eurasia, a weakening and
equatorward shift of the storm tracks and jet that is more robust in the Pacific than in the Atlantic
sector, and increased precipitation over the newly open Arctic and along the west coast of North
America. The atmospheric sensitivity to SIL is generally more robust than the sensitivity to LLW
in the cold season and near to the surface. However, in the warm season, the sensitivity to SIL is
not robust amongst MEs, whereas the sensitivity to LLW is.

For the first time, we have applied pattern-scaling to the surface ocean response (SST and SSS)
and while there are greater inter-ME differences, we find a more robust sensitivity to LLW than
to SIL in virtually all seasons and ocean basins, in general agreement with the sensitivities of
surface heat fluxes, where the pattern of sea-ice loss appears to be more important than in the
atmospheric response. For the sensitivity to LLW, we find warming of the North Pacific and
Atlantic intensified along the western boundary currents as well as freshening of the Pacific and
Arctic Oceans (including the subpolar North Atlantic) and salinification of the Atlantic Ocean.

Taken together, the results for the climate response to SIL presented here provide a partially
cohesive picture. The atmospheric response is generally quite robust, but it is difficult with the
simulations at hand to determine whether some of the discrepancies that we are seeing are the result
of differences in model physical parameterizations, of sensitivity to the mean-state (Smith et al.
2017), of including a mix of transient and equilibrated simulations, or of including Arctic-only and
global sea-ice loss (with the latter being the case for the CESM1 low albedo simulation). While
Sun et al. (2020) have shown that the sea-ice loss protocol is unlikely to play a major role in the
atmosphere or in the SST responses (although albedo forcing will likely result in a weaker response
in DJF), whether the sea-ice loss protocol is playing an important role in other aspects of the
ocean response remains an open question. Fortunately, the proposed Polar Amplification Model Intercomparison Project [PAMIP Smith et al. (2019)] will include coordinated coupled modelling experiments with consistent sea-ice loss protocols and target sea-ice distributions, which should help elucidate some of the questions left open in this work.

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References


Oudar, T., E. Sanchez-Gomez, F. Chauvin, J. Cattiaux, L. Terray, and C. Cassou, 2017: Respective roles of direct GHG radiative forcing and induced Arctic sea ice loss on the North-


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Table 1. Summary of individual simulations and MEs used in this study, including the nomenclature from the original publication, how many ensemble members and years of simulation we have, the radiative forcing and targeted ice area/volume, and whether the integrations are performed as transient or time-slice experiments. The symbols shown in the model column are the same as those used in Fig. 1. Subscripts on the simulations names indicate the publications: \(^1\)Blackport and Kushner (2017), \(^2\)Kay et al. (2015), \(^3\)England et al. (2020b), \(^4\)Marsh et al. (2013), \(^5\)McCusker et al. (2017), \(^6\)Oudar et al. (2017), \(^7\)Sun et al. (2018).

Table 2. A summary of the response pattern names used in this paper and how they are calculated, including the names of the simulations as they are in the original publication, as well as the time periods chosen here. The right two columns show the amount of annual mean low-latitude warming in °C and Arctic sea-ice loss in million km\(^2\) between the differenced simulations. These are the values that are used to obtain the matrices used to calculate the sensitivities. The bolded responses are those that represent the response to Arctic sea-ice loss in isolation in a cool background climate, whereas the italicized responses are those to both radiative forcing and sea-ice loss. The symbols shown in the model column are the same as those used in Fig. 1.
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### TABLE 2

A summary of the response pattern names used in this paper and how they are calculated, including the names of the simulations as they are in the original publication, as well as the time periods chosen here. The right two columns show the amount of annual mean low-latitude warming in °C and Arctic sea-ice loss in million km² between the differenced simulations. These are the values that are used to obtain the matrices used to calculate the sensitivities. The bolded responses are those that represent the response to Arctic sea-ice loss in isolation in a cool background climate, whereas the italicized responses are those to both radiative forcing and sea-ice loss. The symbols shown in the model column are the same as those used in Fig. 1.

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<td>(\text{GHG Effect 20} = \text{ICE21}(101:200) - \text{CTL20}(101:200))</td>
<td>2.30</td>
<td>0.7</td>
</tr>
<tr>
<td></td>
<td>(\text{GHG Effect 21} = \text{CTL21}(101:200) - \text{ICE20}(101:200))</td>
<td>2.52</td>
<td>0.55</td>
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<tr>
<td>GFDL-CM3</td>
<td>(\Delta \text{RCP8.5} = \text{RCP8.5}(2070:2090) - \text{RCP8.5}(1990:2010))</td>
<td>2.04</td>
<td>6.02</td>
</tr>
<tr>
<td></td>
<td>(\triangleleft) (\Delta \text{ICE1990} = \text{ICE1990}(2070:2090) - \text{ICE1990}(1990:2010))</td>
<td>0.18</td>
<td>0.2</td>
</tr>
<tr>
<td></td>
<td>(\Delta \text{ICE} = [\text{RCP8.5}(2070:2090) - \text{RCP8.5}(1990:2010)] - [\text{ICE1990}(2070:2090) - \text{ICE1990}(1990:2010)])</td>
<td>0.21</td>
<td>5.82</td>
</tr>
</tbody>
</table>
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Fig. 1. In (a) is shown the amount of annual mean Arctic sea-ice and Northern Hemisphere low-latitude (0-40°N) SST for each of the time slice or equilibrated simulations, as indicated in Table 1. In (b) is shown the same thing for the transient experiments. For these, we also show the evolution in time for each 15-year epoch of each simulation, while the time periods chosen for analysis (see Table 2) are outlined in black. The models are color- and shape-coded according to the legend, and the colors of the lines connecting the various simulations indicate what response (ice loss alone, warming alone, or both ice loss and warming together) obtained by differencing those specific simulations. In (c) is shown the amount of Arctic sea-ice loss and low-latitude warming in these various responses. The shape of the symbol corresponds to the model and the color of the symbol groups the responses according to the type of response. 43

Fig. 2. In the upper row is shown the 15% contour of Arctic sea-ice concentration in the control simulation (thick red contour) and the sea-ice fraction in the simulation with low sea-ice albedo (filled contours) in DJF (a), MAM (b), JJA (c), and SON (d) for the CESM1 MEs. The next rows as in the first but for the analogous WACCM4, CanESM2, CNRM-CM5, and GFDL-CM3 MEs, respectively. 44

Fig. 3. An example of the pattern-scaling decomposition for zonal-mean temperature using all of the CNRM-CM5 ME simulations in DJF: in (a) the ICE+GHG response is shown; in (b), (d), (f), and (h) “GHG Effect 21”, “ICE Effect 20”, “GHG Effect 20”, and “ICE Effect 21” responses, $\delta T_{\text{zon}}^{\text{con}}$ are shown, respectively. In (c) and (e) the patterns of sensitivity to LLW, $\frac{\partial T_{\text{zon}}^{\text{con}}}{\partial T_{l}} \bigg|_{T_l}$, per degree of warming, and SIL, $\frac{\partial T_{\text{zon}}^{\text{con}}}{\partial I_{l}} \bigg|_{T_l}$, per 2 million km$^2$ of sea-ice loss, utilizing the response patterns in (a) and (b) or (d) are shown, and (g) and (i) are as in (c) and (e) but for the calculation utilizing the response patterns in (a) and (f) or (h). 45

Fig. 4. In the left-hand column is shown the DJF multi-ME mean of the pattern of sensitivity to SIL, $\frac{\partial Z}{\partial I_{l}} \bigg|_{T_l}$, per 2 million km$^2$ of Arctic sea-ice loss, for (a) surface temperature, (c) sea level pressure, (e) precipitation, and (g) 850 hPa zonal wind. (b), (d), (f), and (h) as in (a), (c), (e), and (g) but for the pattern of sensitivity to LLW, $\frac{\partial Z}{\partial T_{l}} \bigg|_{I_{l}}$ per degree of warming. In each panel, the hatched areas represent locations where at least 4 of the 5 MEs agree on the sign. The bolded number in the lower right hand corner of each panel represents the median pattern correlation across the different MEs, while the number between panels represents the pattern correlation between the mean sensitivity patterns. 46

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Fig. 7. In (a) the pattern correlations over the Northern Hemisphere extratropics are shown (defined as all regions north of 30°N). Each dot represents the inter-ME correlation between individual sensitivity patterns of surface temperature obtained through pattern-scaling for each season (DJF, MAM, JJA, SON, as well as in the annual mean, ANN). The blue dots represent correlations between the SIL sensitivity patterns and the orange dots represent correlations between the LLW sensitivity patterns. (b) through (f) as in (a) but for sea level pressure,
precipitation, 850 hPa zonal wind, zonal-mean zonal wind, and zonal-mean temperature. The black bars represent the median of the correlations.

Fig. 8. In (a) is the sensitivity of annual mean precipitation to SIL in CESM1 and in (b) is the sensitivity to LLW in CESM1. (c) – (j) as in (a) and (b) but for CanESM2, CNRM-CM5, GFDL-CM3, and WACCM4. In (k) is the multi-ME mean sensitivity to SIL and in (l) is the multi-ME mean sensitivity to LLW. Hatching represents where at least 4 of the 5 MEs agree on the sign of the sensitivity.

Fig. 9. As in Fig. 4 but for DJF sensible heat flux in (a) and (b), for latent heat flux in (c) and (d), for surface net shortwave flux in (g) and (h), and for surface net longwave flux in (i) and (j). A flux from the ocean to the atmosphere is defined as positive.

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Fig. 11. As in Fig. 10 but for sea surface salinity.

Fig. 12. In (a) and (b) as in Fig. 7 but for annual mean SST and SSS, respectively. In (c) inter-ME annual mean spatial correlations are shown for the Atlantic, Pacific, and Arctic basins. (d) as in (c) but for SSS. Blue markers are correlations between SIL sensitivity patterns, while orange makers are correlations between LLW sensitivity patterns.
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