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# Pacific Decadal Oscillation modulates the Arctic sea-ice loss influence on the mid-latitude atmospheric circulation in winter

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# Abstract

The modulation of the winter impacts of Arctic sea ice loss by the Pacific Decadal Oscillation (PDO) is investigated in the IPSL-CM6A-LR ocean-atmosphere general circulation model. Ensembles of simulations are performed with constrained sea ice concentration corresponding to pre-industrial, present-day and future states, and initial conditions sampling warm and cold phases of the PDO. Using a general linear model, we estimate the simulated winter impact of sea ice loss, PDO and their combined effects. In response to sea ice loss, the Arctic lower troposphere warms and a negative North-Atlantic oscillation like pattern appears together with a weak deepening of the Aleutian Low. The two patterns are associated with a weakening of the poleward flank of the eddy-driven jet, while in the stratospheric the polar vortex weakens. Besides, a warm PDO phase induces a large positive Pacific North America pattern, as well as a small negative Arctic oscillation pattern associated with a weakening of the stratospheric polar vortex. However, the effects of PDO and Arctic sea ice loss are not additive. The Arctic sea ice teleconnections in both troposphere and stratosphere are reduced by the PDO, most importantly in the stratosphere. The results are discussed and compared to those obtained with the same model in atmosphere-only simulations, where sea ice loss does not significantly alter the stratospheric polar vortex.

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# Introduction

Since the late 1970s, the Arctic sea ice extent has exhibited a significant decline in all seasons, which is due to human influence (IPCC, 2021 report: Masson-Delmotte et al., 2021) and is expected to continue. Climate models project a summer ice-free Arctic Ocean by 2050, although this date varies depending on the climate scenario considered (SIMIP Community, 2020). Many studies have shown that the Arctic sea ice loss is likely to change the mid-latitude climate, but its extent is still a matter of debate (Cohen et al., 2014; Blackport and Screen, 2020).

Studies with observations have linked the loss of Arctic sea ice in late fall to a negative North Atlantic Oscillation (NAO) in winter (King et al., 2016; Garcia-Serrano et al., 2015; Simon et al. 2020). However, there are many confounding factors at play and the observational period is too short to accurately assess the amplitude of the sea ice loss impact. On the one hand, most atmospheric models forced by a reduction of Arctic sea ice cover simulate a negative NAO-type response in winter (Sun et al., 2015; Peings and Magnusdottir, 2014; Liang et al., 2021; Levine et al. 2021; Smith et al., 2022). Nevertheless, this result is not completely robust as some studies reported a positive NAO (Screen et al., 2014; Cassano et al., 2014) or a weak response that does not project onto the NAO (Screen et al., 2013; Blackport and Kushner, 2016; Dai and Song, 2020). Some of the differences across models can be explained by different regional expressions of Arctic sea ice loss (Levine et al., 2021). On the other hand, all coupled models show a negative NAO response (Deser et al., 2015; Screen et al., 2018; Simon et al., 2021) but fewer studies exist. Furthermore, when comparing observational and modeling studies, the amplitude of the negative NAO response is much weaker for models than in observations (Liang et al., 2021). Understanding these differences within models and between models and observations is an active topic of research (Cohen et al, 2020). For instance, among the coupled model studies, there are very contrasting impacts of the sea ice loss on the Aleutian low. Screen et al. (2018) found a strengthening of the Aleutian low in six sensitivity experiments involving different models or methodologies to melt the sea ice while Cvijanovic et al. (2017), Simon et al. (2021) and Seidenglanz et al. (2021) found a weakening of the Aleutian low or a ridge in the North Pacific, and Blackport and Screen (2019) found no clear Aleutian Low response. A weakening of the Aleutian low in late winter has been associated with less vertical propagation of planetary waves into the stratosphere and to an acceleration of the polar vortex (Nakamura and Honda, 2002; Garfinkel et al., 2010; Smith et al., 2010). Therefore, whether the Arctic sea ice loss affects the polar vortex is still an open question (Cohen et al., 2020). Indeed, some studies found a weakening of the polar vortex in response to Arctic sea ice loss (Kim et al., 2014; Peings & Magnusdottir, 2014; King et al., 2016; Kretschmer et al., 2016; Screen, 2017; Zhang et al., 2016; Hoshi et al.; 2019) while others found no robust winter stratospheric circulation response (Smith et al., 2022).  A lack of stratospheric polar vortex changes could be potentially related to canceling effects from sea ice loss in the Atlantic and Pacific sectors (Sun et al., 2015).

The various responses to Arctic sea ice loss among the previous studies suggest that there might be concomitant signals that interfere with the Arctic sea ice loss impacts (Ogawa et al.,

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2018). Labe et al. (2019) found that for melting sea ice, December Wave 1 components of the 300 hPa geopotential height were reinforced under the East phase of the Quasi-Biennial Oscillation. Gastineau et al. (2017) and Simon et al. (2020) using multivariate regressions found that early winter snow cover in Eurasia and sea ice in the Arctic could constructively interfere to weaken the polar vortex. Peings et al. (2019) and Blackport and Screen (2020) showed that Ural Blocking can more effectively drive a weakening of the polar vortex than a concomitant sea ice reduction. Arctic-midlatitude linkages may also be affected by sea surface temperature (SST) variability, as discussed by Ogawa et al. (2018), Cohen et al. (2020), Dai and Song (2020) and Simon et al. (2020). The Atlantic Multidecadal Variability (AMV) could regulate the Arctic sea ice loss impact on Arctic Oscillation (AO)-like through the stratospheric pathway (Li et al., 2018) or on Pacific-North America atmospheric circulation through horizontal wave propagation (Osborne et al., 2017). Liang et al. (2021) showed that the North Atlantic horseshoe SST pattern (Czaja and Frankignoul, 1999; 2002) and Arctic sea ice concentration act oppositely on the atmosphere in late winter. Also, Park et al. (2016) revealed that the North Pacific SST could modulate the effect of the Arctic Oscillation on winter temperature in East Asia. Using a composite analysis, Screen and Francis (2016) investigated observations and atmospheric model simulations forced with different Pacific Decadal Oscillation (PDO; Mantua et al., 1997) patterns and sea ice extensions. They found that during the warm phase of the PDO, the contribution of sea ice loss to Arctic amplification was smaller than during the cold PDO phase. Many of the model results discussed above are based on individual models, a small selection of models, and/or use one particular methodology. It's therefore essential to extend the analyses to other models or new methodological approaches.

In the present paper, we focus in particular on how persistent PDO-like SST anomalies could modulate the influence of Arctic sea ice loss on the Northern Hemisphere atmospheric circulation. We will be revisiting the previous results of Screen and Francis (2016) with the novelty to account for atmospheric-ocean feedback using a coupled model and under the light of a new method based on general linear models to assess the interaction between sea ice loss and the PDO. The results agree with Screen and Francis (2016) in that the sea ice loss induces a weakening of PDO teleconnections. In addition, the presented method allows accurate quantification of the interactions.

# Methodology

We use the Institut Pierre Simon Laplace coupled model (IPSL-CM6A-LR; Boucher et al., 2020) which contributed to the 6th phase of the international Coupled Model Intercomparison Project (CMIP6; Eyring et al., 2016). The IPSL-CM6-LR uses the atmospheric component LMDZ6A (Hourdin et al., 2020) which includes the land model ORCHIDEE version 2 (Cheruy et al., 2020). It has a 79-layer vertical discretization ranging from about 10 m to 80 km above surface (top at 1 Pa) and a horizontal resolution of 144 × 143 points (2.5° in longitude and 1.25° in latitude). The ocean component is the Version 3.6 stable of NEMO (Nucleus for European Models of the Ocean), which includes the ocean physics module OPA (Madec et al., 2017), the

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sea ice dynamics and thermodynamics module LIM3 (Vancoppenolle et al., 2009; Rousset et al., 2015), and the ocean biogeochemistry module PISCES (Aumont et al., 2015). All NEMO components share the same tripolar grid, eORCA1xL75, with a horizontal resolution of about 1º except in the tropics where the latitudinal resolution decreases to 1/2°. There are 75 vertical levels with 1 m resolution near the surface and 200 m in the abyss.

The experiments are part of the PAMIP (Polar Amplification Model Intercomparison Project) panel of CMIP6, and are described in detail in Smith et al. (2019). Three sets of simulations are performed with the coupled model using an online restoring to constrain the SIC. The specific names of these experiments are pa-pdSIC, pa-piArcSIC and pa-futArcSIC (tier 2) in Smith et al. (2019). The present-day ensemble, hereafter called PD, uses the observed SIC climatology from 1979-2008 in HadISST (Rayner et al., 2003). The pre-industrial ensemble, called PI, uses an Arctic SIC retrieved from the CMIP5 simulations, with a global mean surface temperature that is 0.57°C colder than for the reference period 1979-2008. The future ensemble, called FUT, is calculated in a similar way, but using the CMIP5 scenario simulations to produce the SIC corresponding to a global mean surface temperature 2°C warmer. The SIC field used to constrain the coupled model simulations is called the target SIC in the following. Details on the calculation of their boundary conditions are given in Smith et al. (2019). Complementary experiments to determine the uncoupled atmospheric response have also been conducted and analyzed (see discussion). The specific names of these experiments are pdSST-pdSIC, pdSST-piArcSIC and pdSST-futArcSIC (tier 1) in Smith et al. (2019). These experiments are atmosphere-only simulations, using the same SIC as the one used as the target in the coupled simulations. The simulations use a repeated climatological SST calculated from 1979-2008 in HadISST, except for the grid points where the amplitude of SIC anomalies is at least 10% compared to present-day conditions. In that case, the SST is modified using the procedures described previously for sea ice concentration.

All experiments used the CMIP6 external forcing corresponding to the year 2000. The experiments have a duration of 14 months (from 2000 April 1st to 2001 May 31st). Unless stated otherwise, the first two months of spin-up are excluded to avoid potential initialisation adjustments, so that time series of 12 months are finally analyzed. As previously suggested, a large number of members are needed to characterize the response to sea ice changes (Peings et al., 2021). Therefore, we performed initial-conditions ensembles of 200 members for each Arctic sea ice experiment. This makes a total of 600 14-month simulations for the coupled and also for the atmosphere-only configurations. For the coupled model simulations, the initial conditions were chosen from the available ensemble of 32 historical CMIP6 simulations with the IPSL-CM6A-LR (Bonnet et al., 2021) in the 1990-2009 period. For the atmosphere-only simulations, the initial conditions are similarly sampled from the available ensemble of AMIP runs (22 members) realized in CMIP6 with IPSL-CM6A-LR.

To constrain the sea ice in the coupled model simulations, we use a method analogous to a nudging of the SIC, already used in Acosta Navarro et al. (2022) with the EC-Earth model.

We apply a heat flux anomaly, called F, calculated as:

$$F = \alpha \, H \, \Delta SIC \qquad (1)$$

where H is the online sea ice thickness at a given grid point; $\Delta SIC$ is the difference of actual SIC for the grid point and the target SIC; and $\alpha$ is a relaxation coefficient. Given the short period of the simulations (14 months), we aim at reproducing the target SIC field within a few days. We

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found that a relaxation constant of 3500 W /m² m leads to little difference between the simulated and target sea ice (see Figure 1). This corresponds to a time constant of about 1 day for typical values of the latent heat of fusion and ice density. To achieve an effective nudging at short time scale, an additional flux anomaly is applied under the ice, as SST is either nudged with a relaxation coefficient of 100 W / m² K (if $\Delta SIC < 0$) or prescribed to the freezing point (if $\Delta SIC > 0$). The difference between two sets with different concentrations of sea ice reveals the impact of changing sea ice.

Figure 1 shows the Arctic SIC simulated in the coupled "pre-industrial" (PI), "present-day" (PD) and "future" (FUT) simulations. As described in Smith et al. (2019), the winter sea ice loss in FUT is mostly located in the Barents-Kara, Labrador and Chukchi Seas compared to PI. The upper panel of Fig. 1 shows the simulated ensemble mean Arctic sea ice area and compares it to the target one. From August to February, the simulated SIC of the three coupled experiments is in good agreement with the target SIC. However, they underestimate by ~0.5 to 1 $10^6$ $km^2$ the sea ice area from April to July, with differences smaller in FUT (red lines) than in PI (green lines). The size of the confidence intervals of the ensemble mean, assuming Gaussian distribution, is small for all months, which implies that the nudging method has effectively reduced the large internal variability of the Arctic sea ice obtained in IPSL-CM6A-LR (Jiang et al., 2021).

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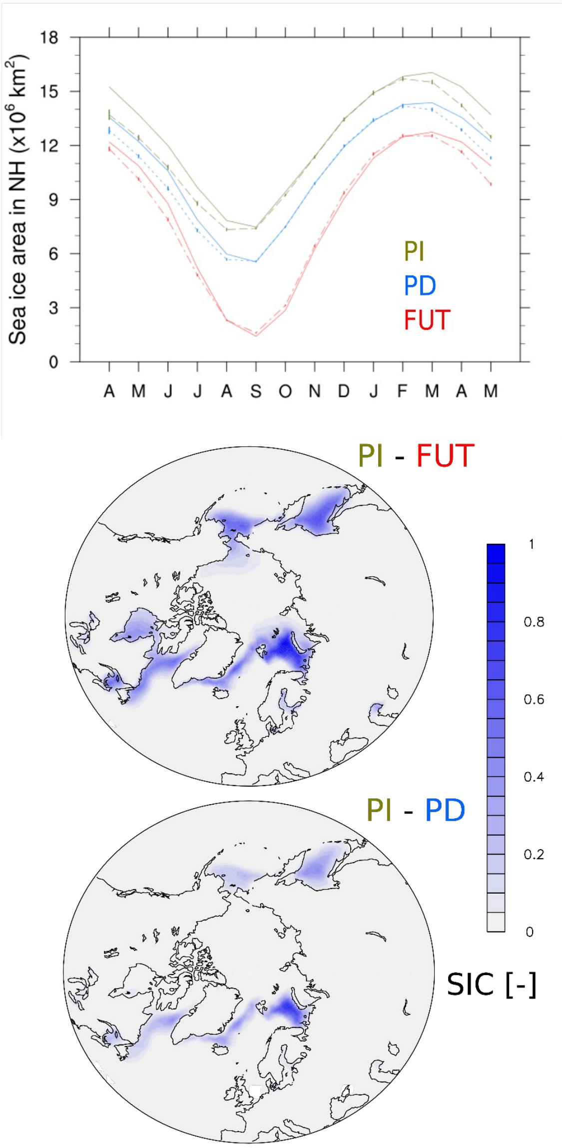


Figure 1: (Top) Arctic sea ice area (in $10^6$ km$^2$) for the ensemble mean of coupled model simulations using constrained SIC for (red, dash-dotted line) FUT, (blue, dotted line) PD and (green, dash line) PI. The corresponding target sea ice is shown with solid lines. Vertical bars represent the 95% confidence interval for the ensemble mean. (Center) Simulated Arctic sea ice concentration changes in the coupled model ensembles for PI minus FUT and (Bottom) PI minus PD averaged from December to February.

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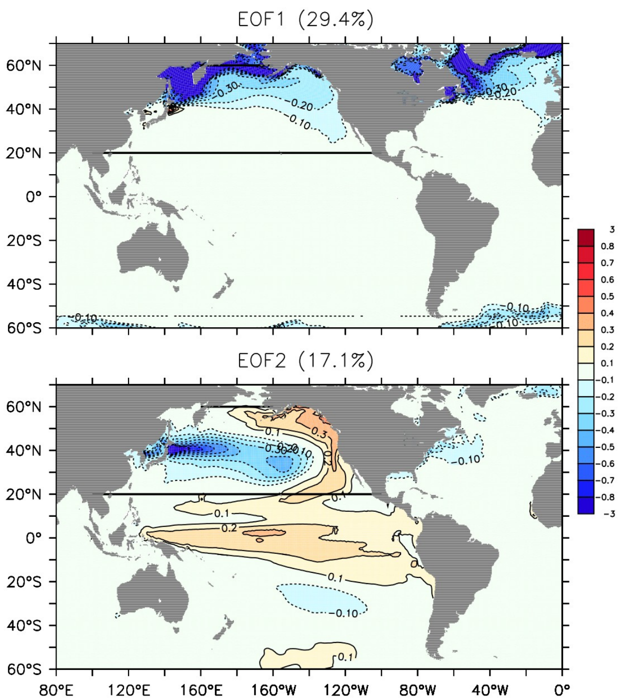


Figure 2: (Top) First and (bottom) second empirical orthogonal function of the yearly averaged SST between 20°N and 60°N in the Pacific ocean in the ensembles of coupled simulations.

To characterize the Pacific Ocean decadal variability, an empirical orthogonal function (EOF) analysis of the yearly sea surface temperature (SST) between 20°N and 60°N in the Pacific ocean (Fig. 2, black lines) is performed using the concatenated outputs of the ensembles PI, PD and FUT. This EOF analysis uses the member dimension instead of the time dimension, as classically used. The EOFs are defined as the regression of the SST onto the standardized principal components (PCs). The first EOF (Fig. 2 top) shows large loadings in the Chukchi, Okhotsk and Bering seas where sea ice was removed in PD and FU conditions (see Fig. 1). It is associated with anomalies of the same sign in the North Atlantic at the edges of the Arctic sea ice cover. The first PC explains 29.4% of the variability of the concatenated PI, PD and FUT members. It shows the dominant influence of the mean sea ice changes, with standardized values around 1, 0 and -1 for simulations PI, PD and FUT, respectively (not shown). The second EOF explains 17.1% of the variance and shows a horse-shoe shaped anomaly in the eastern Pacific that typically characterizes the PDO (Fig. 2, bottom). The anomalies in the eastern Pacific are associated with an equatorial Pacific SST of the same sign, reflecting the role of the El Nino Southern Oscillation (ENSO) in generating the PDO. Conversely, anomalies with the opposite sign are located in the western and central North Pacific, with maximum amplitude off Japan. This pattern is similar to the observed Pacific Decadal Oscillation in the warm phase but for the midlatitude horseshoe extending too much in the western Pacific, together with the equatorial SST anomaly, as found in many other climate models (Sheffield et al., 2013; Coburn

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and Pryor, 2021). Hereafter, the PDO index is defined as the standardized second principal component. A positive PDO index corresponds to a warm PDO phase and a negative PDO index to a cold PDO phase.

In order to investigate the simultaneous atmospheric influence of the sea ice changes and the PDO, we use an analysis of the covariance based on a general linear model. This methodology benefits from the use of the three ensemble simulations together (600 members) and avoids building composites dependent on the arbitrary choice of a threshold. Hereafter, we only focus on the atmospheric anomalies in winter, defined as the 3-month mean in December-February-March. The atmospheric variables from the concatenated 600 members are regressed using the PDO index as a covariate and the sea ice state as a categorical independent variable with three levels. We use the PI conditions as the reference. We also consider the interactions between the sea ice and the PDO, as we find that it significantly improves the explained variance of the general linear model in many locations (not shown).

At each grid point, the general linear model is defined as follows:

$$Y(n) = \beta_0 + \beta_{PD}\,[PD](n) + \beta_{FUT}\,[FUT](n) + \beta_{PDO} PDO(n) + \beta_{PD:PDO}[PD](n)\,PDO(n) + \beta_{FUT:PDO}[FUT](n)\,PDO(n) + \varepsilon \qquad (2)$$

where Y(n) designates the dependant variable, an atmospheric variable in simulation $n$ ;
[PD](n) is a dummy variable with a value of 1 if the simulation $n$ is from PD ensemble, and 0 otherwise (same for [FUT](n) with FUT);
$PDO(n)$ is the PDO index for simulation n ;
$\beta_0$is the intercept;
$\beta_{PD}$ is the regression coefficient determining the effect of the sea ice in PD (FUT) when compared to PI (same for $\beta_{FUT}$ with FUT);
$\beta_{PDO}$ is the regression coefficient determining the effect of the PDO;
$\beta_{PD:PDO}$ is the regression coefficient determining the interaction between the PDO and the PD sea ice (same for $\beta_{FUT:PDO}$ with FUT). It evaluates to what extent their contributions are non-additive;
$\varepsilon$ is a residue.

Statistical significance is estimated using a two-tailed Student's t-test for each of the regression coefficients, assuming all members independent. To account for the overestimated global significance when only using local tests, we calculate the field significance with the False discovery rate (FDR; Wilks et al., 2016) in the Northern Hemisphere between 20°N and 80°N. We choose a FDR p-value of $a_{FDR}$ = 20% to achieve a global test level at 10%, assuming a spatial decorrelation of ~1.54 $10^3$ km, which is consistent with the previous estimations using the 500-hPa geopotential height (Polyak, 1996).

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# Results

We first analyze the effect of the sea ice loss in winter by comparing PD with PI (PD-PI) and FUT with PI (FUT-PI) in the coupled simulations, using the general linear model. We then investigate the impacts of the PDO and finally how they are modulated by the sea ice loss, using a warm (i.e. positive) PDO phase for illustration.

The air temperature at 2m (Fig. 3) shows as expected a significant warming over the polar cap of about 4°C when comparing PD and PI (top-left) and about 10°C when comparing FUT and PI (top-middle). In its warm phase, the PDO induces warming over the northwest America of about 2°C and a cooling over the North Pacific, over Siberia and south of the North America continent of about 1°C (top-right). The interaction term between the sea ice loss and the PDO is non-negligible (bottom), showing a cooling over North America and warming over northeast Siberia, which thus contributes to slight regional damping of the PDO teleconnections. However, this interaction term is larger for FUT than for PD, and is barely significant for PD sea ice loss. A warm PDO thus modulates the sea ice impact by minimizing the warming in North America and enhancing the warming in northeast Asia. As the analysis is linear, a cold PDO phase will lead to the opposite effect to a warm PDO phase, but the interaction between sea ice loss and the cold PDO still results in a damping of the PDO teleconnections.

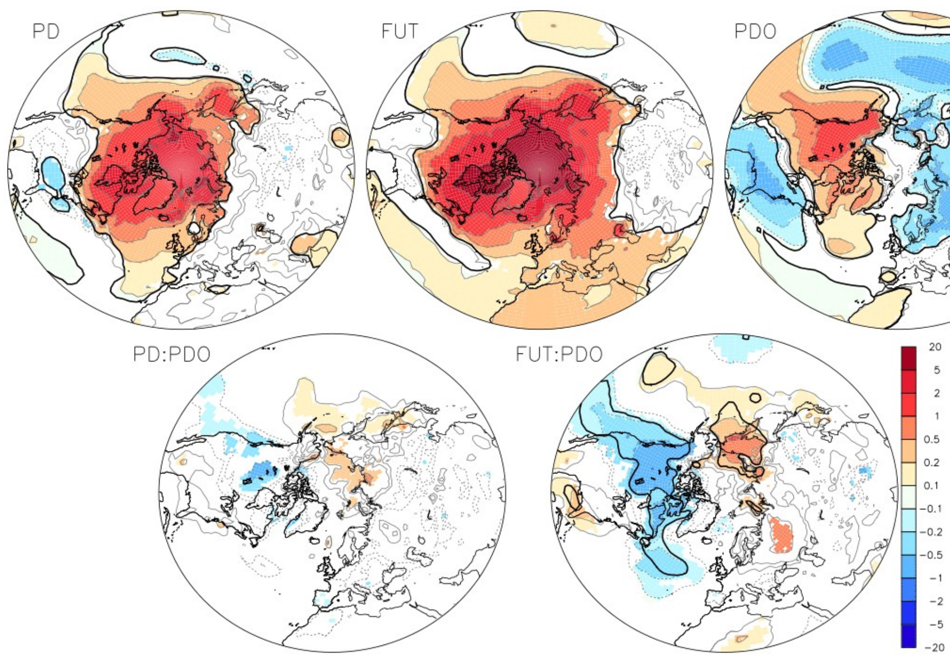


Figure 3: Surface air temperature at 2m (in °C) in response to the sea ice loss and PDO in the coupled simulations when using an analysis of the covariance: (top-left panel) effect of the PD

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sea ice loss ($\beta_{PD}$in Eq.(2)); (top-middle) effect of the FU sea ice loss ($\beta_{FUT}$in Eq. (2)) (top-right) effect of a warm PDO ($\beta_{PDO}$in Eq. (2)); (bottom-left) effect of the interaction between PD sea ice loss and the PDO ($\beta_{PD:PDO}$ in Eq. (2), and (bottom-right) effect of the interaction between the FUT sea ice loss and the PDO ($\beta_{FUT:PDO}$ in Eq. (2)). The color shades a p-value below 10%. The black line indicates field significance, as given by the false discovery rate.

The Arctic sea ice loss additionally induces a significant deepening of the Aleutian Low and a negative NAO-like response. This is shown by the negative sea level pressure anomalies over the Northern Pacific and central Atlantic, together with positive sea level pressure anomalies from Greenland to Norway (Fig. 4, top-left and top-center), with larger and broader anomalies in FUT than in PD. The geopotential height at 500-hPa (Fig. 5, top-left and top-center) also shows a strong increase over the polar cap in response to sea ice loss. It increases above Greenland by as much as 20 m in PD, and 40 m in FUT, which is consistent with the surface warming and the associated increase of the lower tropospheric thickness. A negative AO pattern is also found: the geopotential height at 500-hPa decreases by approximately 15 m over a band from western North America to the Iberian Peninsula. Melting Arctic sea ice also induces a small but significant deepening of the Aleutian low at 500 hPa. In the stratosphere, the geopotential at 50-hPa increases over the polar cap in both FUT and PD cases and slightly decreases over southern Europe for PD and over northern Europe for FUT (Fig. 6, top-left and top-right). Figure 7 (top-left and top-right) further shows the zonal mean zonal wind changes, with a significant weakening of the poleward flank of the eddy-driven jet and of the polar vortex between 50°N and 70°N due to sea ice loss. Between 30°N and 40°N the zonal wind is intensified from the surface to 70 hPa, at the core of the subtropical jet. The zonal wind also decreases south of 20°N, in line with a shrinking of the subtropical jet.

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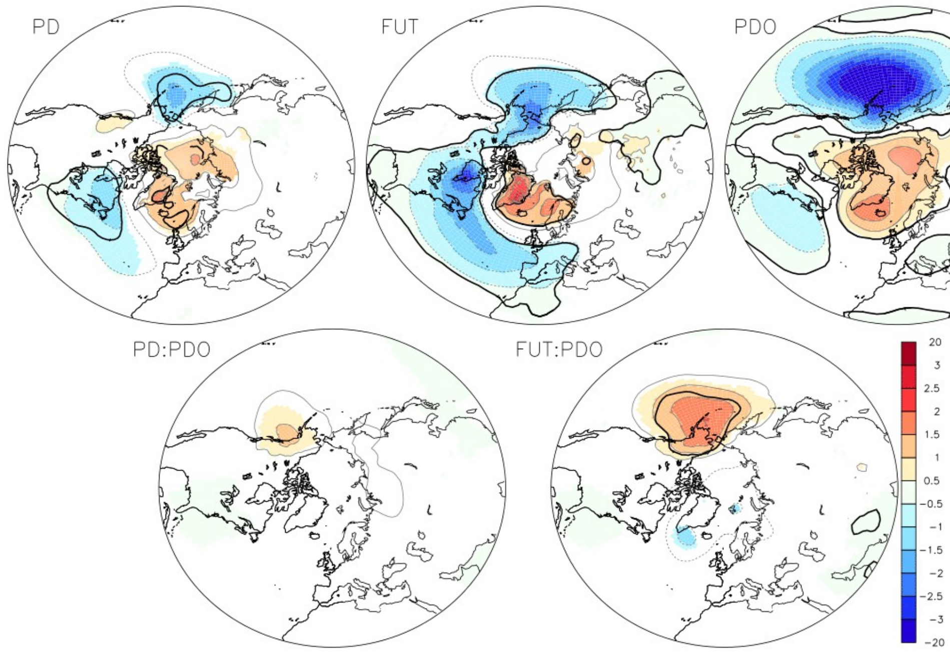


Fig 4: Same as Fig. 3 but for sea level pressure, in hPa.

The experiments can also be used to investigate the influence of a positive PDO on the atmosphere. A warm PDO induces a significant positive Pacific-North American-like (PNA) pattern, with a strong strengthening of the Aleutian Low, a ridge over Northwest America/polar cap, and a small geopotential height increase over southeastern North America (Figs. 4 and 5, top-right). Such impacts are consistent with the influence of the warm equatorial Pacific SST anomalies associated with the PDO onto the PNA (Trenberth and Hurrell, 1994; Newman et al., 2016). In the stratosphere, the geopotential height at 50-hPa shows a tripole pattern with a high over the Arctic and two lows over the eastern North Pacific and Europe, resembling the negative phase of the Arctic Oscillation (Fig. 6, top-right). The warm PDO induces a significant weakening of the poleward flank of the eddy-driven jet from 50°N to 70°N, as well as a weakening of the stratospheric polar vortex between 50°N and 80°N (Fig. 7, top-right). The zonal winds also show a large increase between 20°N and 40°N at the core of the subtropical jet. Such PDO impacts are consistent with findings linking the PDO to the stratosphere based on observations (Woo et al., 2015) and models (Hurwitz et al. 2012; Kren et al. 2016). Nevertheless, it remains unclear whether the stratospheric impacts of the PDO are linked to the extratropical part of the PDO pattern or to the associated equatorial SST anomalies. Indeed, warm equatorial SST anomalies associated with an El Niño have been previously shown to drive a weakening of the Aleutian low, which leads to decreased momentum flux from upward propagating planetary waves that weaken the stratospheric polar vortex (Manzini et al., 2006; Hurwitz et al., 2012; Woo et al., 2015; Kren et al., 2016; Domeisen et al., 2019), a response that is consistent with our regression result for the PDO.

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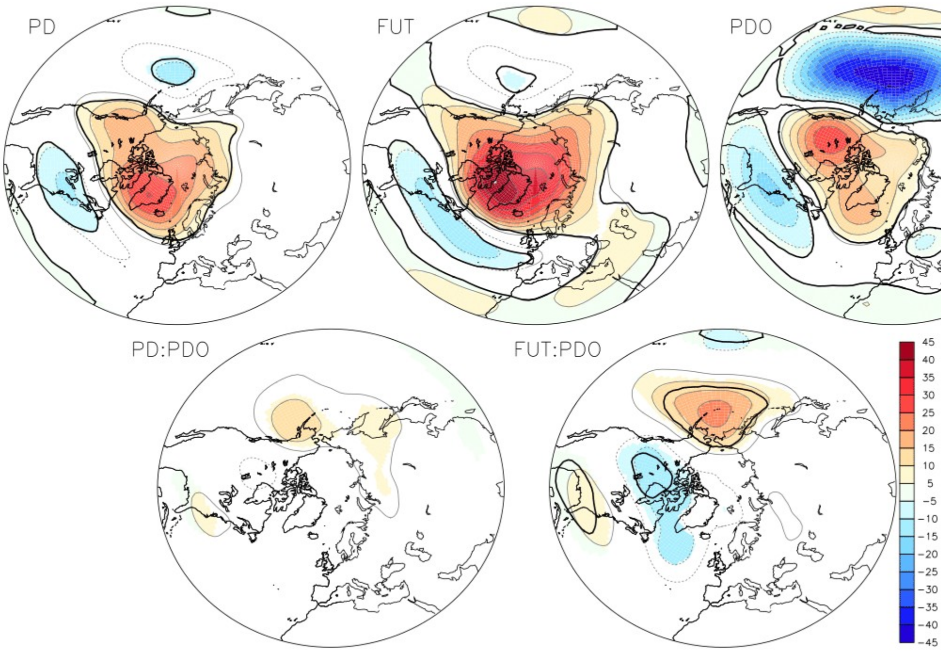


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353 Figure 5: Same as Figure 3 but for geopotential height at 500 hPa, in m.

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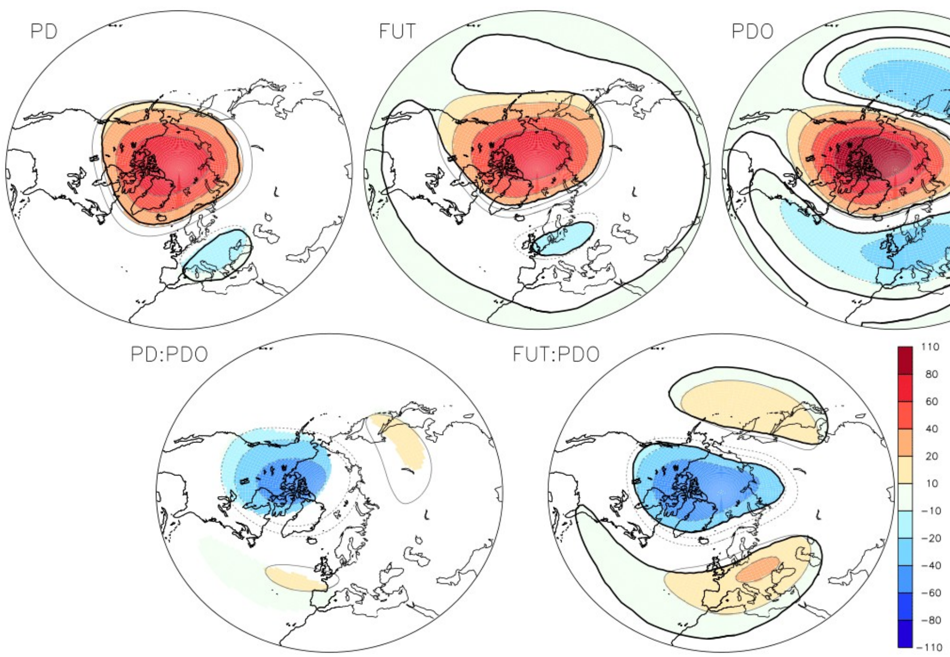


Figure 6: Same as Figure 3 but for geopotential height at 50 hPa, in m.

The interaction between sea ice loss and the PDO leads to a weakening of the Aleutian Low (Fig. 4, bottom) and a pattern reminiscent of a wave train at 500 hPa, resembling a negative PNA phase (Fig. 5, bottom). The results of the interaction between sea ice loss and PDO are qualitatively robust regardless of the magnitude of the sea ice loss (e.g. FUT or PD), but the amplitude of the interaction is small and it is only significant in FUT. In PD, the interaction shows local p-values below 10% but is not field significant. Also, the effect of interaction is stronger and more significant in the stratosphere. At 50 hPa, a significant strengthening of the polar vortex is found, with negative anomalies above the polar cap and positive anomalies over the northwest Pacific and Europe (Fig. 6, bottom). Again, the stratospheric polar vortex increase is stronger and more significant for FUT than for PD. The interaction between PDO and sea ice loss also shows zonal wind changes consistent with a strengthening of the polar vortex (Fig. 7, bottom). Hence, the PDO teleconnections in both troposphere and stratosphere are damped under sea ice loss conditions, in particular for the stratosphere.

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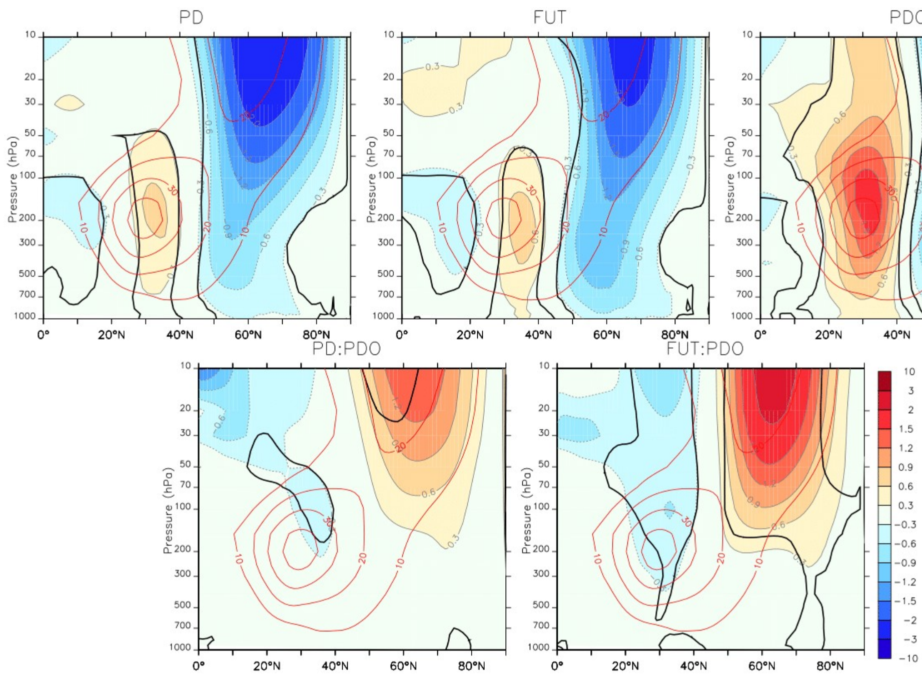


Figure 7: Same as Figure 3 but for zonal mean zonal winds, in m s$^{-1}$. The black line indicates a p-value below 10%.

To understand the causes of the zonal mean wind changes, the zonal-mean diagnostics of transformed eulerian mean quantities are derived following Andrews et al. (1987). In response to FUT sea ice melting, the warming located north of 40 °N is amplified toward the surface in the lower troposphere but extends throughout the troposphere (Fig. 8, top-left). There is also an important warming in the stratosphere from 100 hPa to 10 hPa over the polar cap, north of 60°N. The troposphere also warms between 20°N and 30°N, which can be linked to the shrinking of the subtropical jet (see Fig. 7). A warm PDO phase also leads to a stratospheric warming (Fig. 8, top-right) and a polar vortex weakening (Fig. 7, top-right). However, it is associated with a warming of the tropical troposphere that is intensified in the upper troposphere. The warming over the Arctic associated with a positive PDO is rather uniform and is not intensified at the surface. A quasi-barotropic cooling is also located at 40°N.

Both sea ice loss and PDO lead to a reduced eddy momentum flux at the poleward flank of the subtropical jet peaking around 300 hPa and extending into the stratosphere (Fig. 8, second row). The eddy heat flux (third row) weakens at the lower-troposphere in response to sea ice loss. In addition, both sea ice loss and warm PDO decrease the eddy heat flux between 50°N and 80°N in the lower-stratosphere at 200-hPa, while increasing it above 100-hPa. The anomalous Eliassen–Palm (EP) flux is shown in Fig. 8, (bottom row; vectors), as well as the zonal wind acceleration implied by the EP flux divergence (bottom row; shading). In normal conditions, the EP flux is directed upward and equatorward (not shown) and it converges into the upper troposphere, with two local maximums (Fig. 8, bottom row; contours). One maximum

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is located at 25°N 200-hPa, while the other maximum is between 55°N and 75°N at 400-hPa. This convergence acts to decelerate the zonal wind. The FUT sea ice loss reinforces the convergence between 55°N and 75°N at 400-hPa, with an anomalous upward EP flux in the lower troposphere below (Fig. 8, bottom; color shade). We verified that the convergence is due to the vertical component of the EP flux which is proportional to the ratio between the eddy heat flux and the stratification. As the meridional eddy heat flux shows negative anomalies in this region, the intensification of the upward heat flux in 55°N-75°N mainly results from the weaker atmospheric stratification, leading to a more unstable atmosphere. Between 30°N and 40°N, the EP flux is instead oriented downward in the troposphere, which leads to anomalous divergence between 500-hPa and 200-hPa. It corresponds to the intensification of the core of the subtropical jet in Fig. 7 (top-center). This change is again dominated by the vertical component of the EP flux (not shown) and might reflect the weakening of the meridional eddy heat flux. The same analysis for the PDO influence shows EP flux anomalies somehow similar to those associated with sea ice loss. However, the intensification of the EP flux convergence is located between 40°N and 60°, and the EP flux upper-tropospheric divergence at 30°N is more intense. These changes are again associated with the vertical component of the EP flux (not shown) associated with an intensification of the tropospheric meridional eddy heat flux between 30°N and 40°N. In both sea ice loss and PDO cases; the changes of the eddy momentum flux can be described as a positive feedback reinforcing the changes of the eddy heat flux, as in Smith et al. (2022). In the stratosphere, a clear intensification of the EP flux is simulated poleward and upward in response to sea ice loss and PDO, consistent with the weakening of the polar vortex.

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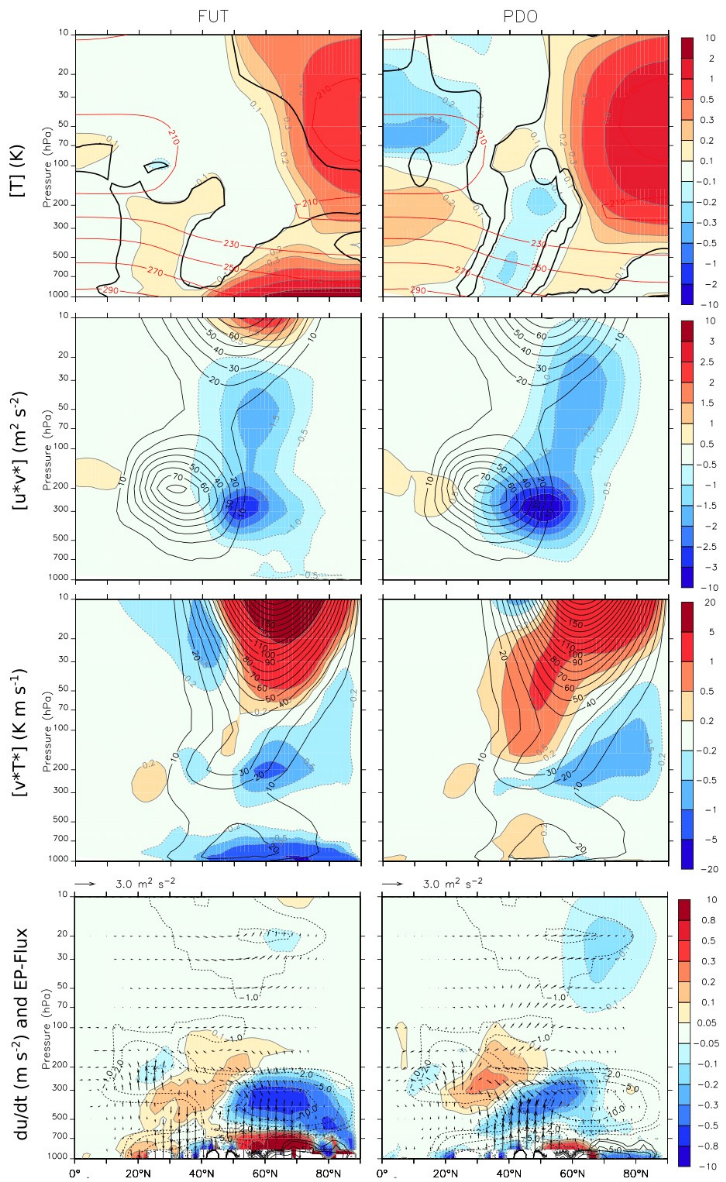


Figure 8: Zonal mean temperature and atmospheric circulation changes related to (left panels) sea ice loss in FUT and (right panels) PDO. Temperature (in K; 1st row), eddy momentum flux (u*v* in $m^2.s^{-2}$; 2nd row), eddy heat flux (v*T* in $K.m.s^{-1}$; 3rd row), zonal wind tendency implied by the Eliassen-Palm flux divergence (in $10^2$ $m.s^{-1}.day^{-1}$ ; bottom row, color shade) and Eliassen-Palm flux ($m^2.s^{-2}$; bottom row, vectors). In the bottom row, the black contours show the zonal wind tendency implied by the Eliassen-Palm flux divergence in the PI ensemble, chosen as a reference. The regressions with a p-value below 10% are indicated by a thick black line in the top panel.

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# SUMMARY & DISCUSSION

We performed sensitivity experiments with the IPSL-CM6A climate model to study the short term response (within 14 months) to the Arctic sea ice loss. We focussed on the winter (DJF) atmospheric circulation changes and how the PDO interacts with the sea ice impacts. The simulations show a robust negative NAO-like pattern in response to sea ice melting, in line with most studies (Deser et al., 2015; Screen et al., 2018; Simon et al., 2021). A positive PNA with a strong deepening of the Aleutian Low is simulated in response to warm PDO, which is a well-established teleconnection (Trenberth et al., 1998; Mantua et al., 2002; Li et al., 2007). The response to Arctic sea ice loss also includes a small deepening of the Aleutian low, as in Blackport and Screen (2019). The discrepancy with other studies in sign (Cjivanovic et al., 2017; Simon et al., 2021) or in amplitude (Screen et al., 2018) can be explained by the timescale investigated. Both Blackport and Screen (2019) and our study are focused on short response time scales less than 5 years, which might be too short to affect the trade winds and to generate SST anomalies in the tropics. The sea ice melting and the PDO were found to generate similar atmospheric circulation changes. Both lead to a weakening of the eddy-driven jet on its poleward flank, an intensification of the subtropical jet and a weakening of the polar vortex. However, for the sea ice loss, these changes are governed by the lower-tropospheric warming north of 50°N and the weaker lower-tropospheric meridional temperature gradient. The weakening of the eddy-driven jet on its poleward flank is induced by weaker surface stratification leading to increased upward Eliassen-Palm flux and acting to reduce the mean zonal flow. Conversely, a warm PDO phase mainly intensifies the transient eddy heat flux at 30°N-40°N, probably through the intensification of the stationary wave pattern. The combined response of the mid-latitude atmospheric circulation to a warm PDO and sea ice melting is not additive, with the interaction between both signals being partly destructive. When sea ice melts during a warm PDO phase, the impacts are smaller than the ones expected by the addition of the two effects. This applies to the anomalies simulated in both the troposphere and stratosphere.

The general linear model presented here can be applied to the analysis of other modes of climate variability or ensembles of sensitivity experiments, such as the idealized experiments of the DCPP panel of CMIP6. The model uses all the ensemble members when estimating the different influences, which are thus based on a larger sample than in traditional methods, and it does not involve the choice of an arbitrary threshold, as when building composites. However, the method does not account for non-linearities, and the impacts of warm and cold PDO could be asymmetric. However, we verified that the changes of the Aleutian low and the polar vortex are symmetric in composites based on the PDO index (Figure 9; top). The AO pattern is only slightly asymmetric in the present-day sea ice conditions, when the neutral and cold PDO states have a similar AO impact (Fig. 9). The AO pattern is symmetric in the preindustrial and future sea ice conditions. Hence, the linear analysis seems applicable to a good approximation.

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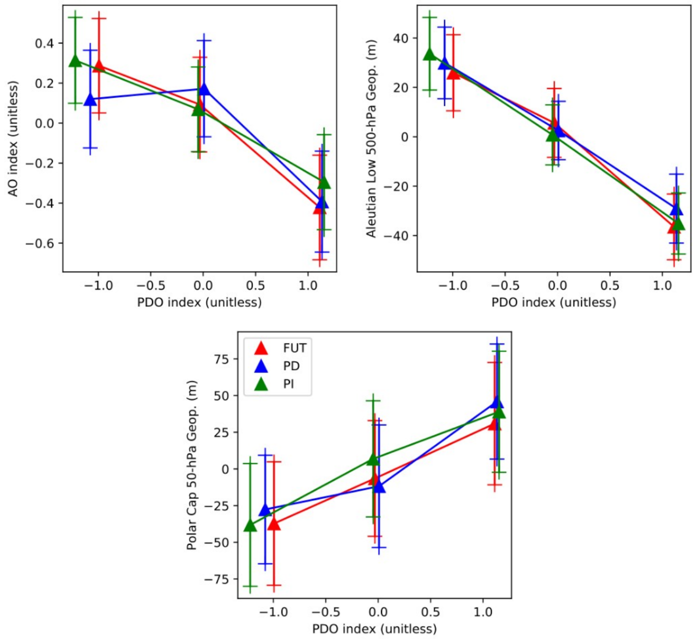


Figure 9: Mean AO index (top-left; unitless), Aleutian low (top-right; in m) and polar cap 50 hPa geopotential height (bottom, in m), for members sorted following the PDO index for the PI (green lines), PD (blue lines) and FUT (red lines) ensembles. The triangles indicate the value for PDO<$Q_{1/3}$, $Q_{1/3}$<PDO<$Q_{2/3}$ and PDO>$Q_{2/3}$, where $Q_{1/3}$ and $Q_{2/3}$ indicate the first and second tercile of a gaussian distribution. The error bar provides the 95% confidence interval. The AO index is calculated as the first principal component of the 500-hPa geopotential height using all the members. The Aleutian low index is the anomaly of the 500-hPa geopotential height in 150°E-180°E 40°N-50°N. The polar cap 50-hPa anomalies is calculated with the mean value of the 50-hPa geopotential north of 60°N.

Concerning the amplitude of the response to sea ice loss, observational study estimates that winter Arctic sea ice loss could have led to a much larger increase, perhaps as much as 200 m over Iceland at 500 hPa over the last four decades if linearity and perpetual winter conditions could be assumed, although it should be less in more realistic conditions (Simon et al., 2020). Nonetheless, the Arctic sea ice loss impact on the NAO is smaller in our sensitivity experiments (30 to 50 m). The reasons for this discrepancy are under active debate (Cohen et al, 2020). This might be explained by too weak eddy feedback represented in models (Smith et al., 2022) but also by the difficulty to cleanly attribute a response to Arctic sea ice decline in observations. Here, we show that the PDO is an important confounding factor that has an impact on the Arctic similar to that of Arctic sea ice loss, especially in the stratosphere. Much care is therefore needed to separate these two effects when using observations. Moreover, depending on the protocol used to constrain the sea ice in coupled model sensitivity experiments, the amplitude of the atmospheric response to sea ice loss can vary by a factor of two (Simon et al, 2021). The

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simulated changes might therefore depend on the methodology used to constrain the sea ice cover. Moreover, the sea ice thickness was not constrained in the sensitivity experiments but might play an important role in the atmospheric circulation response (Lang et al., 2017).

The bulk of our analysis was based on simulations with an ocean-atmosphere general circulation model. However, a different response to sea ice loss might be obtained with atmospheric-only configurations where the two-way air-sea coupling is not allowed. Studies have primarily investigated the ocean feedback on timescales from decadal to centennial. Deser et al. (2015) found that full ocean coupling amplifies the Arctic sea ice loss impact in 100-year simulations, while no feedback was found in an atmospheric model coupled to a slab ocean at decadal timescale in Cvijanovic et al. (2017). However, few studies have investigated short simulations of 14 months, where only fast feedbacks can operate. To determine the role of the coupling, we have performed the same sensitivity experiments but using the atmosphere-only configuration of the IPSL-CM6A model (hereafter ATM). We show that the tropospheric circulation response to sea ice loss is very similar to that in the coupled experiments, although the increase of the 500-hPa geopotential height over the Arctic is weaker in the ATM model (Fig 10, top). Besides, the coupled simulations present a stronger weakening of the stratospheric polar vortex than the atmospheric-only simulations (Fig. 10, middle rows). The lower troposphere warming is more intense in the coupled model, which reflects the presence of sea ice-atmosphere feedbacks, such as those involving thinner sea ice. This leads to a more intense decrease of the tropospheric stratification and intensifies the upward planetary wave propagation into the stratosphere. Since the tropospheric response to a weakened polar vortex resembles the negative AO (Baldwin and Dunkerton, 1999; Kidston, 2015, Cohen et al, 2017; Hoshi et al. 2019), the stronger stratospheric polar vortex weakening might explain the larger AO anomaly in the coupled experiments.

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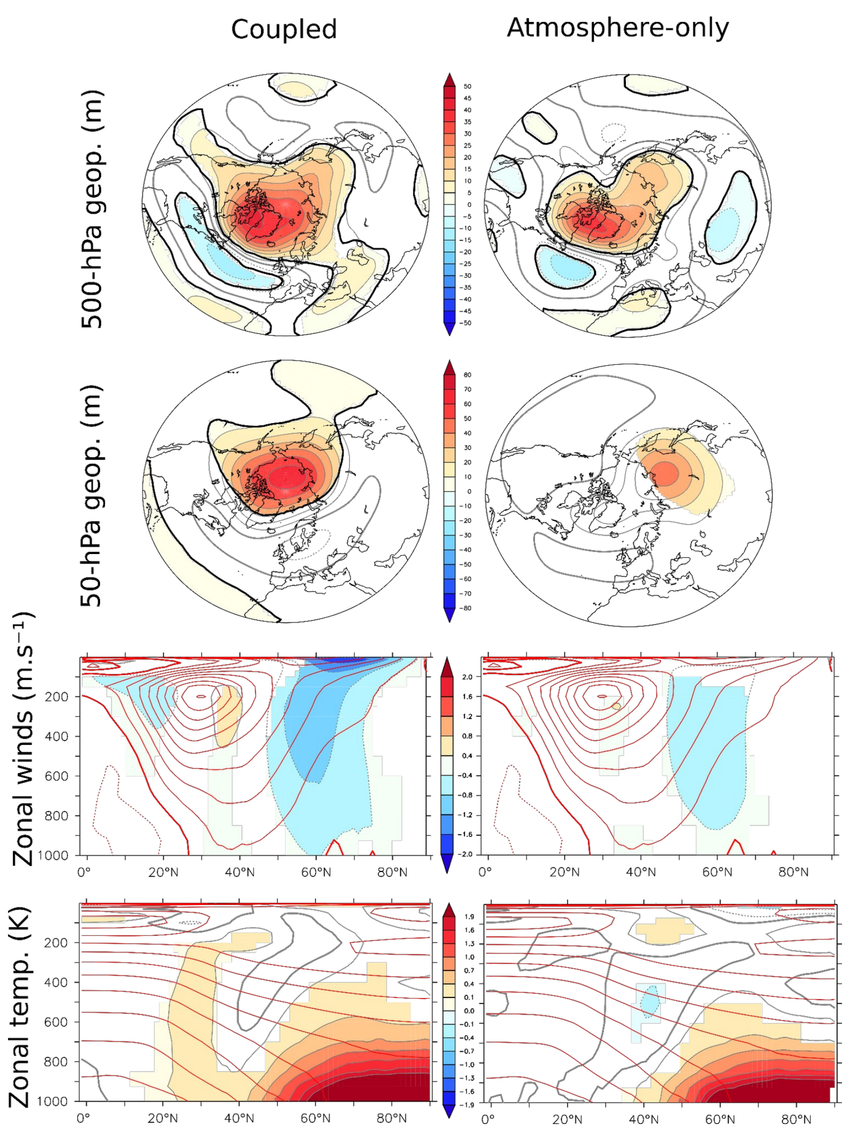


Figure 10: Difference between 200-members ensemble of FU and PI (color and gray outline) in DJF for the geopotential height at 500-hPa (m; top), the geopotential height at 50-hPa (m; middle-top), the zonally averaged zonal wind (m/s; middle-bottom) and the zonally averaged temperature (K; bottom) in the coupled (left) and atmosphere-only (right) configurations of the IPSL-CM6A-LR. Colors are masked if the confidence level of the Student's t-test is less than 90%. The 90% confidence level based on the False Discovery Rate (FDR) is given in black contours for the two top rows. On the middle-bottom and bottom panels, the zonal mean of the wind zonal of the PI simulation in DJF is indicated by the red contours with an interval of 5 m s$^{-1}$, the thick red line indicates zero, solid line positive values and dashed line negative values.

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We applied the same analysis as for the PDO to investigate the AMV, defined as the SST over 0ºNº60N-0ºW80W, influence and its modulation by sea ice loss in the sensitivity experiments with the coupled model, similarly using its distribution among members resulting from the different initial North Atlantic conditions. It was further applied to the quasi-biennial oscillation (QBO) defined as the equatorial zonal wind at 30-hPa. In both QBO and AMV cases, their identified impacts onto the atmospheric circulation were barely significant, and there was no significant interaction with sea ice loss (not shown).

The ocean changes were not investigated in these short simulations, as they are likely to be small and confined to the surface mixed layer. However, the sea ice loss impacts onto the ocean could be very different in longer simulations. Indeed, the atmospheric response to sea ice loss can be different in transient (a few decades) or equilibrium conditions (more than five decades) (Simon et al., 2021; Blackport and Kushner, 2016; Liu and Fedorov, 2019). In particular, the changes in the Beaufort Gyre (Lique et al., 2018), the North Atlantic inflow (Simon et al., 2021) and the Atlantic Meridional Oceanic circulation (Sévellec et al. 2017) would play an important role.

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## Code and data availability

Supporting information that may be useful in reproducing the authors' work is available from the authors upon request (ajsimon@fc.ul.pt or guillaume.gastineau@locean.ipsl.fr).

## Author contributions

AS, GG and CF contributed to the conceptualization of the study and the scientific interpretation of the results. VL and PO developed and coded the nudging method. AS and GG performed the simulations and analysis. GG carried out the Eliassen-Palm flux calculation. AS prepared the first version of the manuscript and, GG, CF and PO have reviewed and edited the manuscript.

## Competing interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this manuscript.

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