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# Transient climate response to Arctic sea-ice loss with two ice-constraining methods

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19 **Abstract**

20

21 The impact of Arctic sea-ice loss on the ocean and atmosphere is investigated focusing on a gradual  
22 reduction of Arctic sea-ice by 20% on annual mean, occurring within 30 years, starting from  
23 present-day conditions. Two ice-constraining methods are explored to melt Arctic sea-ice in a  
24 coupled climate model, while keeping present-day conditions for external forcing. The first method  
25 uses a reduction of sea-ice albedo, which modifies the incoming surface shortwave radiation. The  
26 second method uses a reduction of thermal conductivity, which changes the heat conduction flux  
27 inside ice. Reduced thermal conductivity inhibits oceanic cooling in winter and sea-ice basal  
28 growth, reducing seasonality of sea-ice thickness. For similar Arctic sea-ice area loss, decreasing  
29 the albedo induces larger Arctic warming than reducing the conductivity, especially in spring. Both  
30 ice-constraining methods produce similar climate impacts, but with smaller anomalies when  
31 reducing the conductivity. In the Arctic, the sea-ice loss leads to an increase of the North Atlantic  
32 water inflow in the Barents Sea and Eastern Arctic, while the salinity decreases and the gyre  
33 intensifies in the Beaufort Sea. In the North Atlantic, the subtropical gyre shifts southward and the  
34 Atlantic meridional overturning circulation weakens. A dipole of sea-level pressure anomalies sets  
35 up in winter over Northern Siberia and the North Atlantic, which resembles the negative phase of  
36 the North Atlantic Oscillation. In the tropics, the Atlantic Intertropical Convergence Zone shifts  
37 southward as the South Atlantic Ocean warms. In addition, Walker circulation reorganizes and the  
38 Southeastern Pacific Ocean cools.

## 39 **1. Introduction**

40 The Arctic is a region of pronounced climate change. Since the mid 20<sup>th</sup> century, the Arctic has  
41 warmed more than twice as fast as the rest of the planet (e.g., Bluenden and Arndt, 2012), a  
42 phenomenon referred to as Arctic Amplification. The Intergovernmental Panel on Climate Change  
43 (IPCC) Special Report on the Ocean and Cryosphere in a Changing Climate (Pörtner et al., 2019)  
44 concluded that over the 1979-2018 period the Arctic sea-ice extent has shrunk in all months of the  
45 year with a maximum decrease in September, with a reduction of about 13 % per decade. Also, the  
46 Arctic sea-ice has thinned and the area of multi-year ice has declined by about 90 %. These trends  
47 are expected to increase in the future. The multi-model mean of the representative concentration  
48 pathway (RCP) 8.5 scenario projected a summer ice-free Arctic in the coupled model  
49 intercomparison project version 5 (CMIP5; Stocker et al., 2013) by 2060 and in the version 6  
50 (CMIP6; SIMIP community, 2020) by 2050.

51 The influence of Arctic sea-ice decline on global climate remains under debate, in particular its  
52 influence on mid-latitudes (Overland et al., 2013; Cohen et al., 2014). Observational studies have  
53 linked Arctic sea-ice loss in late autumn to a negative North Atlantic Oscillation in winter (NAO;  
54 King et al., 2016; Garcia-Serrano et al., 2015; Simon et al., 2020), but there is still discussion on the  
55 robustness and pathway of this sea-ice influence. As the observational records are short, climate  
56 models have been extensively used. Among atmospheric general circulation model (AGCM)  
57 studies, there is no consensus on the atmospheric response to sea-ice loss. Some studies (Singayayer  
58 et al., 2006; Strey et al., 2010) found no NAO-like pattern as a response to Arctic sea-ice loss, while  
59 others found a positive NAO response in winter (Screen et al., 2014) or both autumn and winter  
60 (Cassano et al., 2014). Besides, others studies show a negative NAO response to Arctic sea-ice  
61 decline, but with different seasonality: larger in early spring (Seierstad et al., 2009; Sun et al. 2015),

62 in winter (Magnusdottir et al., 2004; Peings and Magnusdottir, 2014) or only in February (Deser et  
63 al., 2010b). Among atmosphere-ocean general circulation model (AOGCM) studies, there is a  
64 broader consensus on a negative NAO-like response in winter (Deser et al., 2015; Blackport and  
65 Kushner, 2016; Blackport and Kushner, 2017; McCusker et al., 2017; Oudar et al., 2017; Smith et  
66 al., 2017; Suo et al., 2017; Screen et al., 2018). In the ocean, observational and modelling studies  
67 found a strengthened North Atlantic inflow and weaker stratification in the Barents Sea and the  
68 Eastern Arctic, a phenomenon called “Atlantification”, reinforcing the sea-ice loss (Årthun et al.,  
69 2012; Polyakov et al., 2017; Barton et al., 2018). However, it is still unclear how and at which rate  
70 the Arctic ocean salinity, temperature and stratification will be modified (Wassmann et al., 2015;  
71 Lind et al., 2018). In AOGCMs, the Arctic sea-ice decline also weakens the Atlantic Meridional  
72 Overturning Circulation (AMOC; Oudar et al., 2017; Sévellec et al., 2017; Suo et al., 2017; Sun et  
73 al., 2018; Wang et al., 2018; Liu et al., 2019) because of freshwater release and modified surface  
74 heat fluxes in the Arctic-North Atlantic region. However, the relative importance of each process  
75 remains unclear.

76 The impacts of Arctic sea-ice loss are not confined to the North Atlantic in coupled models where  
77 ocean-atmosphere coupled feedbacks are accounted for. Deser et al. (2015) showed that the impact  
78 of Arctic sea-ice loss then becomes global. However, the large-scale response is not unanimous.  
79 Most previous studies (Deser et al., 2010b; Deser et al., 2015; Blackport and Kushner, 2016;  
80 Sévellec et al., 2017; Oudar et al., 2017; Suo et al., 2017; Monerie et al., 2019; Screen et al., 2018,  
81 Sun et al., 2018, Liu and Fedorov, 2019; England et al., 2020; Sun et al., 2020) found a tropical  
82 warming with the largest warming in the Central Pacific, similar to the greenhouse gas-driven  
83 warming. This warming called "mini-global warming" in Deser et al. (2015), is associated with an  
84 intensified Aleutian Low in winter. However, the fast transient response to sea-ice loss was found to

85 be less robust when the ocean circulation is not fully adjusted, typically after one to five decades  
86 following a sea-ice perturbation. Blackport and Screen, (2019) found no change in the Aleutian  
87 Low with 5-years-long simulations. In Wang et al. (2018), the Equatorial Pacific and the Southern  
88 Ocean are hardly modified in the first decades of their AOGCM simulation or when using a slab-  
89 ocean instead of a full ocean model. Cvijanovic et al. (2017) rather found a cooling of the  
90 Southeastern Pacific, with their climate model simulations based on both slab-ocean and full-ocean  
91 configurations. Blackport and Kushner (2016), as well as Liu and Fedorov (2019), also found  
92 different oceanic and atmospheric responses in the early (first decades) and late period (after one  
93 century) of their simulations, while Sun et al. (2018) found generally similar responses. The reason  
94 for the discrepancy regarding the transient Pacific response is still under debate.

95

96 Some of these studies used a relatively large sea-ice perturbation yielding an ice-free Arctic during  
97 two to four months (Deser et al., 2015; Blackport et al., 2017; Oudar et al., 2017; Suo et al. 2017;  
98 Sun et al., 2020). However, there is also a need to estimate the impact of smaller Arctic sea-ice loss,  
99 for which the Arctic Ocean in September is not ice-free in September, as in (Blackport et al., 2016;  
100 Blackport et al., 2017; Cvijanovic et al., 2017; Blackport and Screen, 2019). As the climate of the  
101 next decades is important for a wide range of climate impacts (Masson-Delmotte et al., 2018), we  
102 investigate next a moderate Arctic sea-ice loss, corresponding to a loss of 20% on annual mean and  
103 50% reduction in September compared to present-day conditions. As we will detail later, this  
104 corresponds to a reduction expected to occur in approximately 2040.

105

106 Another open question concerns abrupt versus gradual sea-ice reduction. One can argue that the  
107 transient climate response to an abrupt Arctic sea-ice retreat occurring within a few years would  
108 change if the climate system had more time to adjust. As in Sun et al. (2018), we will impose a  
109 gradual sea-ice loss, comparable to that found in scenario simulations.

110

111 Many different methods have been used to constrain sea-ice in AOGCMs: nudging (McCusker et al.  
112 2017; Smith et al., 2017; Suo et al., 2017), flux adjustment (Oudar et al. 2017; Monerie et al.,  
113 2019), ghost forcing/ice-nudging (Deser et al., 2015; Deser et al., 2016; Tomas et al., 2016; Sun et  
114 al., 2020), sea-ice/snow albedos or emissivity modifications (Deser et al., 2015; Blackport et al.,  
115 2016; Blackport et al., 2017; Sévellec et al., 2017; Blackport and Screen, 2019; Liu and Fedorov,  
116 2019), Arctic Ocean albedo modification (Cvijanovic et al., 2015) or changing sea-ice physics  
117 parameters with large uncertainties (Cvijanovic et al., 2017). However, sea-ice and snow thermal  
118 conductivity is also a key parameter for ice melting and growth, and we evaluate next the ability of  
119 thermal conductivity modification to constrain sea-ice. Also, the sensitivity of the climate response  
120 to the methodology remains poorly evaluated, as most previous studies use a single model and only  
121 use one method. Recently Sun et al. (2020) compared the albedo method with ice-flux nudging and  
122 found an identical equilibrium global climate response. Blackport and Screen (2019) impose two  
123 different albedos parameters (albedo of cold deep snow on top of sea-ice or albedo of snow- free  
124 ice) which leads to different seasonal cycles of Arctic sea-ice extent. We will similarly investigate  
125 two different methods but focusing on the fast climate response. We show that both methods induce  
126 qualitatively similar local and remote transient climate responses, but with different magnitude of  
127 Arctic warming. The remote responses to sea-ice reduction simulate in both cases a relative cooling  
128 of the Southeastern Pacific Ocean.

129 In section 2, the methodology and experimental protocol are detailed. Two ice-constraining  
130 methods are presented, and their similarities and differences are discussed. The Arctic and North  
131 Atlantic responses to the Arctic sea-ice retreat are discussed in section 3, while section 4 focuses on  
132 global changes. Conclusion and discussion follow in section 5.

133

## 134 **2. Methodology**

### 135 2.1. Model description

136 We perform simulations with the coupled atmosphere-ocean general circulation model IPSL-  
137 CM5A2 (called here CM5A2; Sepulchre et al., 2019), a modified version of IPSL-CM5A-LR  
138 (called here CM5A; Dufresne et al., 2013) which was used for CMIP5. CM5A2 uses the same  
139 resolution and physical package as CM5A, but it includes an optimized hybrid parallelization to  
140 obtain better computing performances, and a modified tuning to reduce the cold bias of CM5A in  
141 global surface air temperature.

142 The atmospheric component is the LMDZ5A model (Hourdin et al., 2012), with a resolution of  
143  $\sim 3.7^\circ$  in longitude and  $\sim 1.9^\circ$  in latitude and 39 vertical levels up to 4 Pa. The land surface module is  
144 ORCHIDEE (Krinner et al., 2005). The ocean and sea-ice are simulated by the NEMOV3.6 model  
145 (Nucleus for European Modelling of the Ocean; Madec et. al, 2008), using the ORCA2 grid with  
146  $182 \times 149$  cells, corresponding to a nominal resolution of  $2^\circ$ , and 31 levels. The ocean biochemistry  
147 is modelled by PISCES (Aumont and Bopp, 2006). Sea-ice dynamics and thermodynamics are  
148 represented by the LIM2 model (Fichefet and Maqueda, 1997; Fichefet and Maqueda, 1999), a  
149 single ice-category model with three layers (one for snow and two for sea-ice) for heat storage and  
150 vertical heat conduction.



151 As shown in Sepulchre et al. (2019), CM5A2 is realistic in many aspects but, as in many low-  
152 resolution coupled models, the Gulf Stream and the North Atlantic current are too zonal, generating  
153 a cold bias in the North Atlantic sea-surface temperature (SST) of about  $-2^{\circ}\text{C}$  and up to  $-5^{\circ}\text{C}$ . The  
154 AMOC is underestimated with a mean value of  $12 \pm 1.1$  Sv (from  $30^{\circ}\text{S}$  to  $60^{\circ}\text{N}$ ) in pre-industrial  
155 conditions, compared to observational estimates around 19 Sv (Cunningham et al., 2007). This  
156 weak AMOC has been linked to a lack of convection in the Labrador Sea. The main deep-water  
157 formation sites are instead located in the Greenland Sea and south of Iceland.

158 The rainfall in CM5A2 is largely similar to CM5A (Sepulchre et al. 2019). In both versions, there is  
159 an overestimation of rainfall in the Southern Tropics, leading to the “double ITCZ” (Intertropical  
160 Convergence Zone), and an underestimation in the mid-latitudes ( $20^{\circ}\text{N}$ - $40^{\circ}\text{N}$ ).

161 In the version CM5A, September (minimum) and March (maximum) Arctic sea-ice area and  
162 thickness are overestimated, albeit this remains within the range of CMIP5 models (Maslowski et  
163 al., 2012; Stroeve et al., 2012; Kirchmeier et al., 2017). In the updated version CM5A2, sea-ice  
164 extent has been improved in the North Atlantic sector. During the 1979-2005 period of a historical  
165 run, the mean Arctic sea-ice extent in September is about  $5.8 \cdot 10^6$  km<sup>2</sup> (Fig. 1), in annual is about  
166  $10.7 \cdot 10^6$  km<sup>2</sup> (not shown) and the annual thickness is 2.5 m (not shown). The sea-ice extent is  
167 calculated as the total area of all grid cells with at least 15% sea-ice concentration. This compares  
168 well with the respective observed value for the same period:  $5.5 \cdot 10^6$  km<sup>2</sup> in September (Fig. 1;  
169 Cavalieri et al., 1996),  $10.2 \cdot 10^6$  km<sup>2</sup> for the annual mean (not shown; Cavalieri et al., 1996) and  
170 about 2 m (Schweiger et al., 2011).

## 171 2.2 Experimental protocol

172 We first run a control ensemble of 10 members with CM5A2 for 30 years, called CTRL. The  
173 greenhouse gases, aerosols, ozone and land-use are kept constant at the level of the year 2000. Ten  
174 initial atmospheric conditions are chosen randomly from a stabilized present-day control run  
175 starting from a 500-yr spin-up simulation in pre-industrial conditions. Oceanic initial conditions are  
176 identical and correspond to year 90 of this present-day control run. In a 2500-years pre-industrial  
177 control of the CM5A model, the standard indices of Pacific Decadal Oscillation (PDO), Atlantic  
178 Multidecadal Oscillation (AMO) and AMOC have autocorrelation with e-folding time smaller than  
179 10-yr (see Fig. S1 for details). Therefore, we speculate that oceanic initial conditions would not  
180 affect the 10-30 year response investigated here. Hereafter, we discard the first 10 years unless  
181 stated otherwise.

182

183 Figure 1 displays the observed sea-ice extent calculated from the monthly sea-ice concentration  
184 (SIC) based on passive microwave measurements from the National Snow & Ice Data Center  
185 (NSIDC; Cavalieri et al., 1996). It shows that over the last 40 years, the September sea-ice extent  
186 has reduced by about 50%. The CTRL simulation and the existing historical experiments both show  
187 similar mean September sea-ice extent, with values corresponding to that occurring in the 1980-  
188 2005 period. As no scenario runs were available with the CM5A2 version, we use the ones done  
189 with CM5A for CMIP5. To meet a 50 % September reduction, we use as a target the sea-ice extent  
190 simulated in the ensemble mean of the four CM5A RCP8.5 members averaged over the period  
191 2035-2055, called TARGET. This corresponds to an annual reduction of 20% (not shown). As  
192 CM5A shows a large cold bias in the Arctic when compared to the CM5A2 version, the sea-ice loss  
193 is only slightly larger to the reduction occurred in the last 40 years.

194 Two reduced Arctic sea-ice ensembles are then constructed by modifying Arctic sea-ice properties,  
195 while the Southern Hemisphere sea-ice remains unconstrained. To induce sea-ice melt while  
196 ensuring energy and water conservation, we either modify the sea-ice and snow above the sea-ice  
197 albedos or their thermal conductivity. The continental snow properties are unconstrained. Reducing  
198 the ice and snow albedos increases sea-ice melt in spring and summer, while reducing the thermal  
199 conductivity mainly reduces the sea-ice growth in winter. Indeed, when thermal conductivity is  
200 reduced, the sea-ice and snow more effectively insulate the ocean from the atmosphere, so the  
201 ocean (which is at the freezing point) loses less heat in winter and sea-ice basal growth reduces.

202 To determine the sea-ice and snow albedos needed to reproduce the targeted Arctic sea-ice loss  
203 without changing the external forcing, we first use linear regressions, as described in Deser et al.  
204 (2015). Starting from the same initial conditions than the CTRL, we run eight 30-year simulations  
205 with sea-ice and snow albedo reductions ranging from 0% to 70%. After excluding the first 10  
206 years, linear regressions between August-September-October (ASO) Arctic sea-ice area (SIA) and  
207 albedo reduction (blue dots in Fig. 2) provide a first estimate of the albedo reduction needed to  
208 reproduce the ASO Arctic SIA value of TARGET (Fig. 2 top-left). We then repeat simulations with  
209 albedos closer to this initial estimate. To reduce the uncertainty associated with internal climate  
210 variability, we use 5-member (green squares in Fig. 2) or 10-member ensembles (yellow stars in  
211 Fig. 2) with albedo values close to the first estimated value. A reduction of -22.6% for the albedo of  
212 Arctic sea-ice and snow best reproduces the targeted SIA. The same process is followed for the  
213 reduced thermal conductivity (Fig. 2, top-right), and a reduction of 33% is then needed from the  
214 thermal conductivity of Arctic sea-ice and snow.

215 In the following, we will therefore focus on two experiments based on the previous results. The first  
216 ensemble is identical to CTRL except for the Arctic sea-ice and snow albedos reduced by 22.6%

217 and is called ALB. The second one is identical to CTRL, but with Arctic sea-ice and snow thermal  
218 conductivity reduced by 33% and is called THCD. Both ensembles consist of 10 members of 30-yr.  
219 The first 10-yr is discarded. In the following, the impact of the Arctic sea-ice reduction is assessed  
220 by comparing the ensemble-means between ALB (or THCD) and CTRL. Statistical significance is  
221 estimated using Student  $t$ -tests for the difference of means, assuming all members are independent.

222

223 Lastly, we note that the sensitivity of winter January-February-March (JFM) SIA to albedo and  
224 thermal conductivity is different (Fig. 2, bottom). As the albedo modification is acting mostly in  
225 summer, a larger albedo reduction of about -45% is needed to reproduce the target winter sea-ice  
226 area. Interestingly, a reduction of thermal conductivity of about -30% is needed to simulate the JFM  
227 sea-ice area, a value similar to that found when using ASO as a target (-33%), suggesting that the  
228 seasonal cycle of sea-ice loss is best reproduced with the thermal conductivity method.

### 229 2.3 Evaluation of ice-constraining methods

230 The time evolution of the annual Arctic SIA in ALB, THCD, TARGET and CTRL are displayed in  
231 Fig. 3 (left). In CTRL, there is a weak drift with increasing Arctic sea-ice extension, but it remains  
232 small compared to internal variability. The annual sea-ice areas of ALB and THCD are declining  
233 gradually at a rate comparable to that of the TARGET simulation. The results are similar when  
234 focusing on summer or winter (Fig. S2). This contrasts with previous studies that found an abrupt  
235 sea-ice decline after sea-ice albedo modification, as in Blackport and Kushner (2016) or Liu and  
236 Fedorov (2019). In ALB and THCD, the decline is gradual possibly because the perturbation is  
237 small. Indeed, reducing the albedo with a stronger value (70%) simulates an abrupt decline, with  
238 September sea-ice vanishing within 5 years (Fig. S3).

239 Figure 3 (right) compares the seasonal cycles of the Arctic SIA for the different ensembles. The  
240 values for both ALB and THCD remain close to that of TARGET for all months. When  
241 investigating the differences with TARGET, only May and June in THCD are different from  
242 TARGET at the 90% confidence level. We also note that the sea-ice loss is slightly underestimated  
243 in February-March for ALB (and THCD), which is consistent with the smaller sensitivity of the  
244 JFM SIA illustrated in Fig. 2. This good fit in winter for ALB is not in accordance with previous  
245 studies (Blackport et al., 2016; Blackport et al., 2017). This may be due to internal variability and  
246 the fact that TARGET and ALB come from different versions of the model. The annual  
247 (September) sea-ice extent is  $9.8 \cdot 10^6 \text{ km}^2$  ( $5.4 \cdot 10^6 \text{ km}^2$ ) in CTRL and  $7.5 \cdot 10^6 \text{ km}^2$  ( $2.5 \cdot 10^6 \text{ km}^2$ )  
248 in the two reduced-ice ensembles which correspond to a 23% (53%) reduction. Also, the sea-ice  
249 loss equal to  $0.9 \cdot 10^6 \text{ km}^2$  ( $\sim 7\%$ ) during December-January-February (DJF). Figure 4 compares the  
250 spatial patterns of the reduction of winter (DJF) and summer (JJA, from June to August) Arctic sea-  
251 ice concentration (SIC) in TARGET, ALB, and THCD. The spatial distributions of the sea-ice  
252 retreat are relatively similar, especially between ALB and THCD in winter (and spring; not shown):  
253 sea-ice melts mostly in the Barents, Labrador and Chukchi Seas. Compared to TARGET, more ice  
254 melts in the Barents Sea in ALB and THCD, and less in the Labrador Sea. In summer, the Arctic  
255 sea-ice melts almost everywhere with a minimum around the Queen Elizabeth Islands, with only  
256 subtle differences between the ensembles. These small differences may be explained by the use of  
257 two versions of the model: TARGET (CM5A) and ALB or THCD (CM5A2).

258 The similar seasonal sea-ice areas in the two reduced-ice ensembles hide larger differences in sea-ice  
259 thickness (Fig. 5, left). While ALB weakly underestimates the ice thickness compared to TARGET,  
260 THCD has a significantly reduced seasonality. During winter and spring, THCD has thinner ice than  
261 ALB or TARGET, while in summer it has slightly more. ALB strongly melts the surface sea-ice in

262 summer when the incoming shortwave surface radiation is largest, and THCD decreases the basal  
263 sea-ice growth in winter (as less ocean heat is transferred to sea-ice). In summer, changing thermal  
264 conductivity has limited consequences. Indeed, the heat flux in the sea-ice is small, as the sea-ice is  
265 isothermal. In both cases, the sea-ice reduction persists and reemerges in the next season, owing to  
266 coupled interactions among sea-ice thickness, sea-ice concentration and ocean temperature  
267 (Blanchard-Wrigglesworth et al., 2011a). For both ALB and THCD, the greatest thinning occurs  
268 where sea-ice is thickest (Central Arctic; not shown), following the growth-thickness feedback (Bitz  
269 et al., 2004). In ALB and THCD, a thinner snow layer above ice is simulated throughout the year  
270 when compared to CTRL (Fig. 5, right). This is explained by a larger snow melting rate, as the  
271 snow-to-ice conversion is similar to CTRL (not shown). In ALB, the Arctic snow thickness  
272 resembles that of TARGET, except in spring when more incoming solar radiation rapidly melts the  
273 snow. THCD underestimates snow thickness throughout the year, possibly because more heat is  
274 available to melt the snow, as the thermal conductivity is reduced. To illustrate how the  
275 atmosphere/ocean exchanges are modified, Fig. 6 presents the total surface heat flux (short-wave,  
276 long-wave, sensible and latent fluxes; positive downward) over the Arctic. In the Central and  
277 Western Arctic, all three ensembles show anomalous downward heat flux into the ocean, but with  
278 different amplitudes: ALB overestimates while THCD underestimates the heat flux compared to  
279 TARGET (see also Fig. S4, top-left). This is mostly explained by the differences in surface albedo  
280 resulting in different short-wave absorption. Note that the surface albedo in THCD also decreases  
281 over sea-ice (much less than in ALB; not shown) due to the reduction of sea-ice and snow  
282 thicknesses. Most total surface heat flux differences between ALB and THCD are found off Queen  
283 Elisabeth Islands (Fig. 4, bottom). This coincides with the location of multiyear ice, where summer  
284 sea-ice albedo is important. As the sea-ice retreats in the Barents and Chukchi Seas, the winter

285 oceanic heat loss strengthens near the sea-ice edges (Fig. 6, top), mostly due to sensible and latent  
286 heat fluxes (not shown). Over the sea-ice edges, the ALB and THCD ensembles exhibit similar heat  
287 flux changes as they have similar sea-ice losses.

288

289 The area-weighted surface heat flux without short-wave (i.e. turbulent and long-wave fluxes; Fig. 6,  
290 bottom-left) and near-surface air temperature (Fig. 6, bottom-right), averaged north of 70°N display  
291 negative (upward) anomalous heat flux and warmer temperatures for TARGET, ALB and THCD  
292 throughout the year. The annual-mean warming is significantly more pronounced in ALB (1.63°C  
293 north of 70°N) than in THCD (1.16°C north of 70°N; Fig S4, top-right). The radiative budget at the  
294 top of the atmosphere (TOA) shows that the incoming shortwave radiation flux north of 70°N is  
295 doubled in ALB when compared to THCD (not shown). Consistently, the net downwelling  
296 shortwave radiative fluxes increase at the surface, so that the total downward heat flux in the  
297 Canadian Archipelago in ALB is larger than in THCD (Fig. S4, top-left).

298 The Arctic warming is seasonally dependent. Even though the sea-ice cover shows its largest  
299 reduction in summer, the warming over the polar cap in TARGET is maximum in autumn. This  
300 lagged seasonal response, which is reproduced in ALB and THCD, is caused by the upward  
301 turbulent fluxes over the newly opened water, as found previously (Serreze et al., 2007; Screen et  
302 al., 2010; Deser et al. 2010b; Screen et al., 2013; Blackport and Kushner, 2016; Yoshimori et al.,  
303 2018; England et al., 2018). In spring, ALB produces another warm peak, which is significantly  
304 distinct from TARGET. April sea-ice concentration is very similar among ALB, THCD and  
305 TARGET (not shown). The warming mainly occurs near Queen Elisabeth Islands where multiyear  
306 sea-ice is located (not shown), which is likely related to the albedo reduction. The THCD ensemble

307 is colder in winter, as the reduced conductivity leads to a decreased heat conduction through the ice.  
308 As a consequence, ALB is warmer than THCD, which lead to larger outgoing longwave radiation.

### 309 **3. Arctic and North Atlantic responses**

#### 310 3.1. Winter atmospheric changes

311 Atmospheric changes occurring over the North Atlantic are shown for winter, one-to-three months  
312 after the maximum heat flux anomalies. Figure 7 (top) displays the DJF sea level pressure (SLP)  
313 changes in ALB and THCD. In ALB, there are broad anticyclonic anomalies over Northern  
314 Siberia/Eastern Arctic and Greenland, and a low-pressure anomaly over the North Atlantic. This  
315 pattern projects on the negative phase of the NAO. We also see an anticyclonic anomaly in the  
316 Northern Pacific, near the west coast of North America, which is discussed in section 4. In THCD, a  
317 similar pattern is simulated, but with much weaker amplitude, so that the anomalies are not  
318 significant except off west America and above Greenland. As sea-ice loss produces enhanced  
319 warming in the lower troposphere (Deser et al., 2010b; Cattiaux and Cassou, 2013), we also show  
320 the geopotential height at 500 hPa (Z500; Fig. 7, bottom) to illustrate the mid-tropospheric changes.  
321 A broad anticyclonic anomaly appears over the North Pole, consistent with low-tropospheric  
322 warming in both ALB and THCD. The negative Z500 anomaly over the North Atlantic is only  
323 slightly significant in ALB, while the positive anomaly is significant off the northwestern coast of  
324 North America in ALB and THCD. An investigation of the difference between ALB and THCD in  
325 winter further indicates a significant barotropic anticyclone in North Siberia (Fig. S4, bottom-  
326 middle), which is consistent with the anomalous downward heat flux in autumn in this region (Fig.  
327 S4, top-middle). A significant depression over Eurasia is also found (Fig. S4, bottom-middle). The  
328 negative NAO-like pattern is significantly stronger in ALB compared to THCD (Fig. S4, bottom-



329 left) due to the stronger Arctic warming which enhances the pole-to-equator gradient temperature.  
330 At 50-hPa, in the lower stratosphere, ALB shows a weaker polar vortex than THCD (Fig. S4,  
331 bottom-right). Weak polar vortex anomalies classically propagate downward and are followed by  
332 negative Arctic Oscillation (AO) events in winter (Hartmann et al., 2000; Baldwin and Dunkerton,  
333 1999; Baldwin et al. 2003; Kidston et al., 2015). Therefore, it is likely that the stratospheric changes  
334 also contribute to the stronger negative NAO-like anomalies in ALB when compared to THCD.  
335 Nevertheless, such influence on the NAO is larger in early spring, as found by Sun et al., (2015).

336

337 Figure 8 displays the DJF zonal-mean temperature and winds over the North Atlantic sector (80°W-  
338 20°E). A clear temperature increase in lower to mid-troposphere is simulated above the Arctic  
339 (60°N-90°N) as heat is released from the ocean. ALB undergoes a higher and larger Arctic  
340 warming, reaching about 400-hPa compared to 600-hPa for THCD. Significant warming in the  
341 troposphere is also seen between 20°N-40°N up to 200-hPa in both simulations. Weak warming is  
342 found in the free tropical troposphere, resembling what would produce a mini-global warming. The  
343 lower stratosphere north of 60°N is only modified in THCD, with a significant cooling. Consistent  
344 with the weaker meridional temperature gradient, the zonal wind weakens north of 55°N from the  
345 surface up to 100-hPa. The North Atlantic subtropical jet core is amplified around 40°N, 200-hPa.  
346 However, the winds are weaker in the equatorward flank of the jet (30°N to 0°N). For both  
347 ensembles, westerlies at 850-hPa are shifted south in the North Atlantic sector, as the meridional  
348 temperature gradient weakens north of 50°N. The global zonal-mean (over all longitude, not only  
349 Atlantic) winds show a northward shift of the subtropical jet and amplified westerlies at 850-hPa  
350 (Fig. S5). In section 4, we see that this difference is consistent with the anomalies simulated over  
351 the Pacific Ocean.

## 352 3.2 Oceanic response

353 In the ocean, changes induced by melting sea-ice are less seasonal and therefore depicted here as  
354 annual-mean changes. Figure 9 presents the ocean temperature changes averaged over the upper  
355 300 m. The Arctic warms near the summer sea-ice edges, as new open waters have a smaller  
356 surface albedo and allow more incoming solar radiation. A small cooling is simulated in the Barents  
357 and Greenland Seas, where the winter oceanic cooling is amplified as sea-ice retreats (Fig. 4, top).

358 Figure 10 displays oceanic properties for ALB only, as THCD results are similar, albeit with  
359 smaller magnitude (Fig. S6). In the Central Arctic, salinity decreases (Fig. 10, left) and the Beaufort  
360 gyre intensifies (Fig. 10, middle). The cause of these two related features might be the decreased  
361 freshwater export toward the North Atlantic, especially through ice transport at Fram Strait, as  
362 shown by Zhang et al. (2016). On the contrary, a positive salt anomaly is seen around the Eastern  
363 Arctic (Barents, Kara, Laptev, East Siberian, Chukchi Seas), possibly caused by a larger inflow of  
364 North Atlantic water into the Arctic. In the Barents Sea, the 0-300 m temperature decreases slightly,  
365 (Fig. 9), but the top 100 m warms. As the salinity increases (Fig 10, left), the overall density  
366 stratification is reduced, which can lead to increased mixing and a release of the Arctic subsurface  
367 heat, consistently with the cooling of the 0-300m layer. These changes are due mostly to anomalous  
368 horizontal advection rather than to surface fluxes (not shown). The barotropic stream function also  
369 shows a negative anomaly north of the Barents Sea (Fig. 10, middle), consistent with a northward  
370 extension of the Norwegian current bringing more salt up to the north of Barents Sea. All these  
371 changes are consistent with the so-called “Atlantification” found in observations (Årthun et al.,  
372 2012; Polyakov et al., 2017; Lind et al. 2018) and suggests that such a process could be linked to  
373 the Arctic sea-ice loss.

374

375 In Central North Atlantic, a cold and fresh anomaly is simulated along the North Atlantic current  
376 (around 45°N) while warm and salty anomalies are found in the subpolar gyre (see Figs. 9 and 10).  
377 These changes are consistent with the southward shift of the surface westerlies found previously  
378 over the Atlantic sector. This shift of the westerlies can impact the ocean through the changes of the  
379 wind speed and its impacts on turbulent heat fluxes (Deser et al., 2010a; Suo et al. 2017) and by  
380 forcing an “inter-gyre gyre” (Marshall et al., 2001) through a shift of the wind stress curl (Fig. 10,  
381 middle). Indeed, we found that an anomalous gyre is found between Newfoundland and the British  
382 Isles (Fig. 10, center), which cools and decreases the salinity in the southern subpolar gyre by  
383 anomalous advection.

384 For both reduced sea-ice ensembles, the seawater density is slightly reduced over the 0-300 m layer  
385 in the north branch of the subpolar gyre. However, while the surface reduction is due to warming,  
386 deeper density changes are due to a fresh anomaly found in the Greenland Sea, downstream of the  
387 outflow of Arctic water from Fram Strait. The mixed layer depth is shallower in the Greenland Sea  
388 and South of Iceland, at the location of the main deep water formation site in this model (see section  
389 2.1.2). This is consistent with a weakening of the AMOC (see Fig. 11 and S7) that is maximum near  
390 55°N. The AMOC at 55°N, computed by the maximum Atlantic meridional stream function  
391 between 500m and 2500 m, indeed exhibits a steady slowdown with a mean weakening of about 0.8  
392 Sv in ALB and THCD (not shown).

393

#### 394 **4. Global-scale response**

395 The sea surface temperature (SST) anomalies induced by sea-ice loss in ALB (Fig. 12, left) indicate  
396 significant warming both in the South Tropical Atlantic (0°S-20°S) and in the subtropical North

397 Atlantic (20°N-40°N). The SST changes in the subtropical southeast Pacific are similar but more  
398 significant than those in Wang et al. (2018) who analyzed the impact of Arctic sea-ice loss in the  
399 first decades using the ghost forcing method. The Atlantic pattern is consistent with a decrease of  
400 the AMOC (Figs. 11 and S7), which brings less heat from the Southern Hemisphere to the North  
401 Atlantic (Latif et al., 2006; Mignot et al., 2007; Keenlyside et al., 2008; Kageyama et al., 2009).  
402 Besides, the decrease of low-cloud cover in the Southern Atlantic amplifies the warming (not  
403 shown). The North Pacific presents broad warming extending to the western American coasts, with  
404 a maximum north of the Kuroshio Extension. In the South Pacific, the SST pattern resembles the  
405 South Pacific Meridional mode (Zhang et al., 2014), with cooling from 10 to 30°S in the Central-  
406 East Pacific and warming from 20°S to 40°S in the Central-West Pacific. A cooling band is also  
407 simulated at 60°S. The THCD ensemble (Fig. 12, right) shows similar SST anomalies, except for a  
408 warming in the Gulf of Mexico and South Atlantic between 40°S and 50°S.

409

410 In ALB and THCD, the Z500 changes (Figs. 13 and S8, top-left) indicate a weakening of the  
411 Aleutian Low, anticyclonic anomalies centred over the South Pacific and a larger Amundsen Low.  
412 To illustrate the changes in the large-scale tropical atmospheric circulation, the 200-hPa velocity  
413 potential was calculated (Figs. 13 and S8, top-right). This shows the regions of large-scale ascents  
414 for negative velocity potential and descents for positive values, smoothing small scale anomalies  
415 apparent in the vertical velocity. In CTRL, ascents are simulated over the Maritime continent and  
416 the Indo-Pacific warm pool (Figs. 13 and S8, top-right, contours) and descents occur from Eastern  
417 Pacific to Africa. With reduced sea-ice extent, the Walker cell is shifted westward with more ascent  
418 from the Indian Ocean to the Gulf of Guinea and more descent in the Central and Eastern Pacific  
419 Ocean. Even though there is no significant SST cooling in the equatorial Pacific, ALB shows a

420 small equatorial east Pacific cooling with an enhanced zonal SST gradient across the equatorial  
421 Pacific. The associated atmospheric circulation anomalies, therefore, resemble those usually  
422 associated with La Niña (e.g. Sterl et al. 2007) or the cold Interdecadal Pacific Oscillation phase  
423 (Henley et al., 2015; Gastineau et al., 2019).

424

425 Previous work argued that, as Arctic sea-ice melts, the TOA incoming shortwave radiation into the  
426 Arctic increases and the inter-hemispheric northward energy transport should decrease when the  
427 climate is at equilibrium. This leads to an anomalous Hadley cell with northward cross-equatorial  
428 surface winds (Kang et al. 2008; Cvijanovic and Chang, 2013; Yoshimori et al., 2018; Wang et al.  
429 2018), shifting the ITCZ northward. However, we found in ALB and THCD that the atmospheric  
430 meridional energy transport increases in both simulations from 40°S to 65°N (not shown), while  
431 south of 65°N the oceanic meridional energy transport decreases, as the AMOC decreases (Figs. 11  
432 and S7). This leads to southward cross-equatorial surface winds in the equatorial Atlantic. In turn, a  
433 southward shift of the ITCZ is simulated in the Atlantic Ocean, as well as in the south Pacific  
434 convergence zone. Nevertheless, we also found intensified South Pacific trades winds (Figs. 13 and  
435 S7, bottom-left) and anomalous northward cross-equatorial winds are simulated in the Central and  
436 Eastern Pacific, as found by Wang et al. (2018) in the first decades of their simulations. This results  
437 in Hadley circulation changes that are small and insignificant (Fig. S9). We conclude that the  
438 atmospheric northward energy transport changes are complex, as the ocean is not in equilibrium.

439 We also note an increase of precipitation in Brazil and Northeast Australia and drier conditions in  
440 much of North America in boreal winter. The precipitation changes are consistent with the cross-  
441 equatorial wind changes and the Walker circulation anomalies, with a significant southward  
442 (northward) ITCZ shift in the Atlantic (Pacific) Ocean (Figs. 13 and S7, bottom-right). Besides, the

443 SST in the Pacific tends to project on the negative phase of the IPO, even though not significant, is  
444 also consistent with the increase of rain in Brazil (Villamayor et al., 2018). The decrease of  
445 precipitation over California is also seen by Cvijanovic et al. (2017) and explained by large-scale  
446 atmospheric reorganization due to Arctic sea-ice loss. Lastly, the annual precipitation response also  
447 shows a southward shift of the South Pacific Convergence Zone.

448

## 449 **5. Conclusion and Discussion**

450 We investigate the influence of Arctic sea-ice loss on both local and global climate using the IPSL-  
451 CM5A2 model. We focus on the fast transient responses, occurring within 20 years following 10  
452 years of adjustment. We study a relatively moderate Arctic sea-ice loss, corresponding to a 20%  
453 (50%) annual (September) sea-ice extent reduction. Two different methods are implemented to melt  
454 the Arctic sea-ice from a control simulation (CTRL) to assess the robustness of the associated  
455 climate impacts: reducing the albedo (in ALB) or thermal conductivity (in THCD). We adjust their  
456 values in order to reproduce a targeted summer Arctic sea-ice area found in the scenario simulation  
457 of IPSL-CM5A. The resulting sea-ice areas and sea-ice concentration patterns are largely similar in  
458 TARGET, ALB and THCD. However, an underestimation of the winter sea-ice loss is  
459 systematically produced when reducing the albedo, while thermal conductivity reduction is more  
460 able to reproduce the target sea-ice area in both winter and summer. Most previous studies also  
461 found that decreasing the albedo leads to overestimated winter sea-ice (Deser et al., 2015; Blackport  
462 and Kushner, 2016; Screen et al., 2018; Sun et al., 2020). The fact that the ensemble ALB only  
463 simulates a small underestimation of sea-ice loss in winter is consistent with the effect of internal

464 variability and with the (small) difference in winter sea-ice simulated in IPSL-CM5A (used as a  
465 target) and IPSL-CM5A2 (not shown).

466

467 The physical mechanisms reducing the ice are different in the two methods. While albedo modifies  
468 the incoming solar radiation, thermal conductivity modulates the transfer of heat from the ocean to  
469 the atmosphere, controlling the winter sea-ice growth. This induces significant local differences  
470 even if the mean Arctic sea-ice areas are similar. For the reduced albedo simulations, there is a  
471 stronger and less confined Arctic warming, especially in spring (as in Blackport and Kushner,  
472 2016), when sea-ice cover is large. The thermal conductivity method simulates a thinner sea-ice and  
473 snow in winter/spring due to reduced air-sea fluxes.

474 The climate responses are mostly similar with the two methods. However, the magnitude of the  
475 anomalies is larger in the Northern Hemisphere with the albedo ensemble (ALB). Nonetheless, the  
476 Tropical and Southern Hemisphere SST and SLP responses in South Atlantic and South Pacific are  
477 of similar magnitude or larger in THCD. The origin of these small differences between the two  
478 methods remains to be understood using larger ensembles to increase to signal to noise ratio.

479

480 The Arctic sea-ice loss creates a positive sea level pressure anomaly over Northern Siberia and a  
481 negative anomaly in the Central North Atlantic in winter, resembling a negative NAO-like pattern.  
482 In winter, the North Atlantic lower-tropospheric jet is shifted southward which is consistent with  
483 the reduced temperature gradient and the simulated negative NAO-like pattern (Screen et al., 2018).  
484 The subtropical jet in the North Atlantic is also (slightly) shifted southward. However, the global  
485 mean zonal-wind shows a northward shift of the subtropical jet (Fig. S5), due to a strong Pacific

486 contribution. At 40°N, the zonal mean changes are dominated by the weakening of the Aleutian  
487 Low in the Pacific. Even though the warming mostly occurs near the surface, the SLP and Z500  
488 over the Arctic have a barotropic structure, suggesting strong eddy-mean flow interactions, as found  
489 in Deser et al. (2016) and Wang et al. (2018).

490

491 In the past few decades, the Arctic Ocean freshwater content has increased, which has been  
492 explained by the accumulation of freshwater from sea-ice melt and river runoff. Zhang et al. (2016)  
493 linked this accumulation to less sea-ice export as the Beaufort gyre has intensified. This is  
494 consistent with our study as the Beaufort gyre intensifies, while its salinity decreases. The reason  
495 for the spin-up of the gyre has been linked to an anomalous anticyclone over the Beaufort gyre  
496 (Giles et al., 2012) or to reduced sea-ice cover resulting in an increased transfer of momentum to  
497 the ocean (Lique et al., 2018). In our study, such anomalous anticyclone is absent (not shown),  
498 therefore further investigation would be needed to quantify the mechanisms for the Beaufort gyre  
499 intensification. In addition, the salinity increases in the Barents Sea due to stronger North Atlantic  
500 inflow. This is consistent with the so-called Atlantification that is usually invoked to explain sea-ice  
501 variability (Årthun et al., 2012; Polyakov et al. 2017; Barton et al., 2018; Lind et al. 2018). Our  
502 results suggest that Atlantification could be amplified by Arctic sea-ice loss within two or three  
503 decades. The freshwater and heat exchanges between the Arctic and North Atlantic are modified.  
504 The subtropical gyre shifts south and an intergyre gyre develops presumably due to wind changes  
505 (Marshall et al., 2001). The AMOC decreases, which is associated with a shallower mixed layer at  
506 the main convection site. According to previous studies, Arctic sea-ice loss might play a dominant  
507 role in AMOC weakening. For instance, Sévellec et al. (2017) suggested that 75% of the observed  
508 AMOC decline is driven by Arctic sea-ice changes and Sun et al. (2018) found that about 50% of



509 AMOC decline produced at the end of the 21<sup>st</sup> century in a scenario simulation is due to Arctic sea-  
510 ice loss. However, the relative importance of surface heat and freshwater flux in weakening the  
511 AMOC in future climate is still an open question. There is a cold and fresh anomaly in the mid-  
512 latitude around 45°N, which resembles the projected warming minimum (or warming hole) in the  
513 subpolar North Atlantic (Collins et al., 2013) and has been linked to AMOC decrease (Drijfhout et  
514 al, 2012; Sévellec et al., 2017; Suo et al., 2017; Sun et al. 2018). However, as ALB and THCD  
515 show different magnitude of ocean surface cooling (respectively 0.3 °C and 0.08°C) with a similar  
516 intensity of weakening of the AMOC, we suggest that most of the changes are associated with the  
517 southward shift of the westerlies. Lastly, the Atlantic is warmer at 0°S-25°S, which is consistent  
518 with the AMOC weakening (Mignot et al., 2007).

519

520 Even if the equatorial Indo-Pacific shows no significant change associated with sea-ice loss,  
521 warming is simulated in the South subtropical Pacific around 30°S, encircled by cooling around  
522 20°S and 60°S. The pattern resembles the South Pacific meridional mode (Zhang et. al., 2014). It  
523 also broadly resembles a cold IPO (Henley et al., 2015; Gastineau et al., 2019) but with no  
524 significant anomalies in the equatorial band. The cooling around 20°S and South Atlantic warming  
525 is associated with a westward shift of the Walker cells, with more ascent over the Atlantic and more  
526 descent over the Pacific. This suggests that the fast decadal response to sea-ice loss is dominated by  
527 the sea-ice-driven AMOC changes in the Atlantic, which are then driving the Pacific changes  
528 through atmospheric bridges, although the causality was not fully determined. It would be  
529 consistent with previous works where that Atlantic warming leads to a cold IPO phase through  
530 modification of the Walker cells (Li et al. 2016; Ruprich-Robert et al. 2017; Martin-Rey et al.,  
531 2018). However, such mechanism found here might be model-dependent. For instance, Wang et al.

532 (2018) found a large influence of the North subtropical ocean, while in our case the South Atlantic  
533 is key.

534

535 Previous studies found that sea-ice loss is typically associated with a “mini-global warming” at  
536 equilibrium, after several decades of oceanic circulation adjustment. However, in this paper, the  
537 transient changes found after a 10-yr adjustment to sea-ice loss show contrasting results, with a  
538 North Pacific warming and a Southeast Pacific cooling, somewhat resembling those found in the  
539 transient studies of Cvijanovic et al. (2017) and Wang et al. (2018). The reason for the different  
540 response in the Pacific is an open question. The ocean dynamics could be an important aspect.  
541 Wang et al. (2018) find indeed as smaller warming in a climate model fully coupled than coupled  
542 with an ocean mixed layer, especially in Northern Hemisphere. Furthermore, the oceanic initial  
543 state was not varied in our simulations, although it could affect the transient response (Sévellec and  
544 Federov, 2017; Germe et al., 2018). Lastly, as suggested by Monerie et al. (2018) and Smith et al.  
545 (2017), the response to Arctic sea-ice loss could also depend on the mean state. As different  
546 components (ice, AMOC, global temperature) have different adjustment scales, the global response  
547 could change over time. We argue that the fast response to the sea-ice loss of the coming decades  
548 could be quite different from the equilibrium response to sea-ice loss (Liu and Federov, 2019). This  
549 could be clarified by coordinated sensitivity experiments with different climate models.

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553

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559

560

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810 **Figures**

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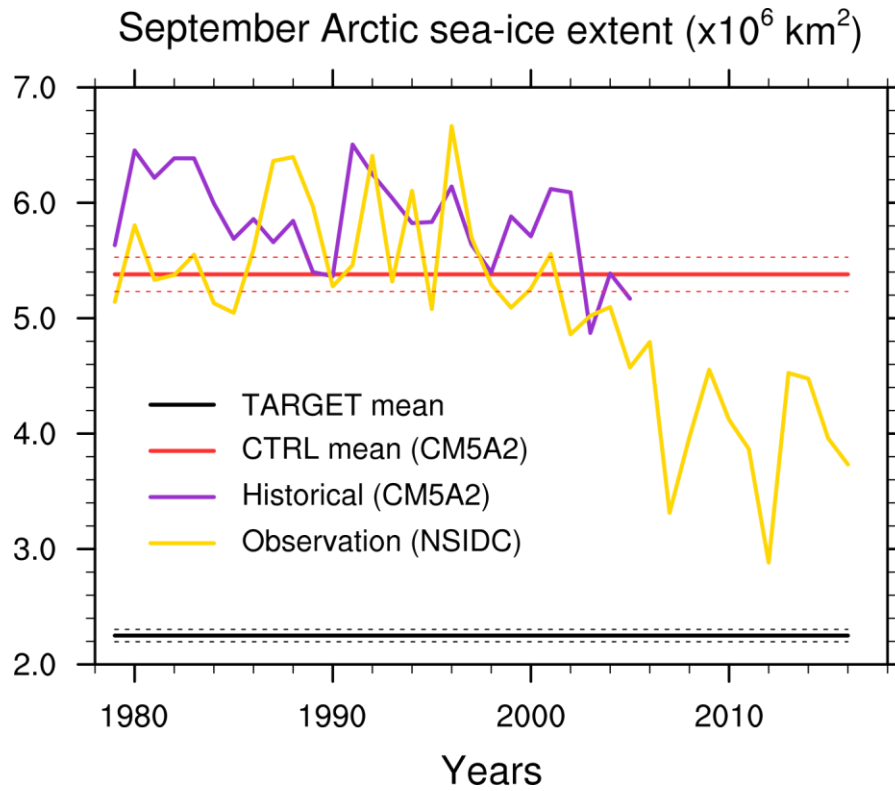
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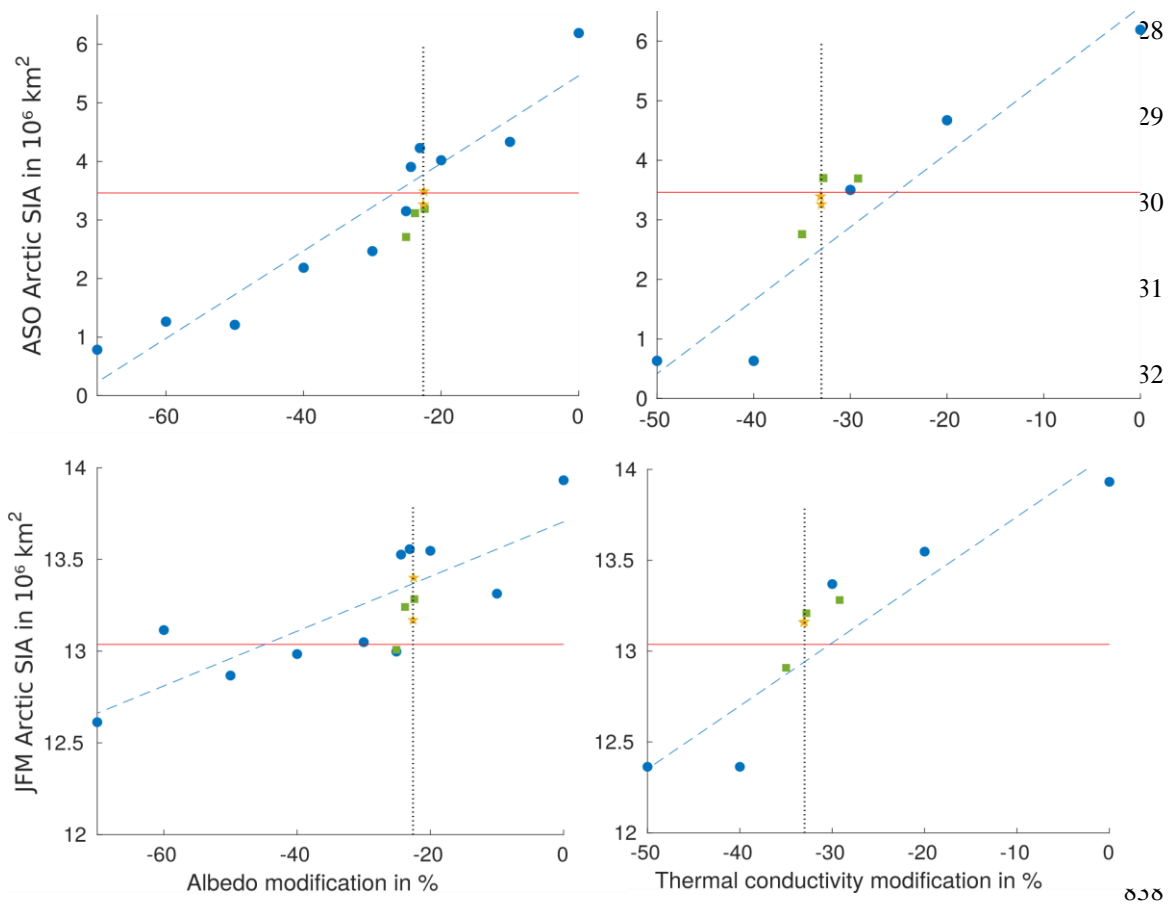
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821 Figure 1: Time evolution of the September Arctic sea-ice extent, in  $10^6$  km<sup>2</sup>, in observation (calculated from  
 822 NSIDC data; Cavalieri et al., 1996; yellow curve), and a historical run with IPSL-CM5A2 (purple line). The  
 823 red (black) thick line shows the mean of the present-day CTRL ensemble (TARGET) and the red (black) thin  
 824 lines display the 90% confidence intervals for the ensemble-means.

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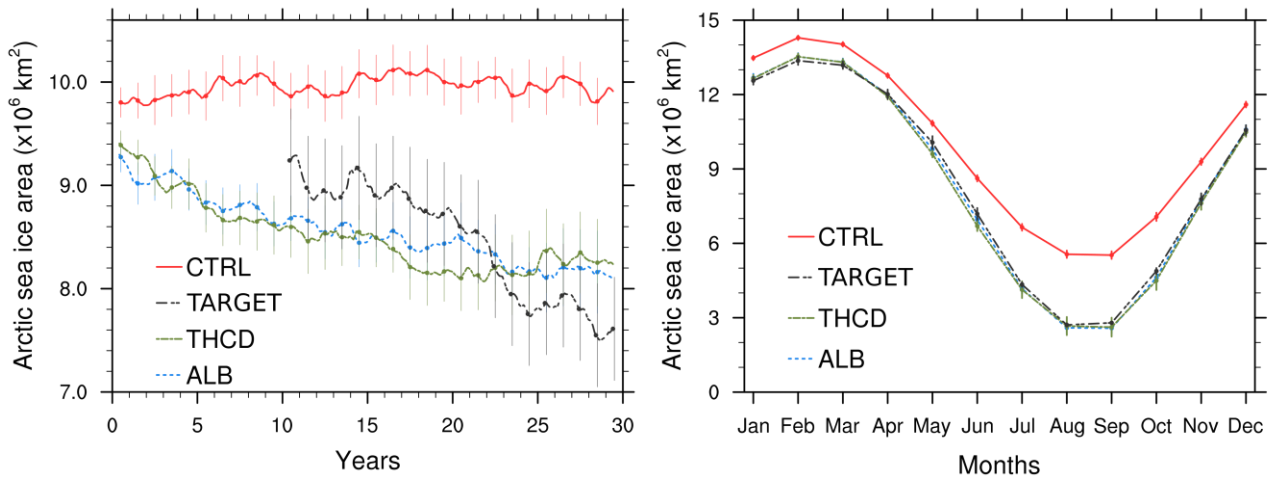


839 Figure 2: Mean Arctic sea-ice area (SIA) averaged over August-September-October (ASO; top) and over  
 840 January-February-March (JFM; bottom), in  $10^6$  km<sup>2</sup>, against the change in albedo (left) and thermal  
 841 conductivity (right), in %. Results from single members are shown by blue dots together with its linear  
 842 regression (dashed blue line). Results from 5-members ensembles and 10-members ensembles are shown by  
 843 green squares and yellow stars respectively. The target (Arctic sea-ice area for the period 2035-2055 with  
 844 CM5A) is indicated by the red line, and the reduction of albedo (22.6%) and thermal conductivity (33%) for  
 845 the two experiments ALB and THCD respectively are indicated by dotted black lines.

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848 Figure 3: Time-series of the annual-mean Arctic sea-ice area (left), and seasonal cycle averaged over the  
 849 years 10-30 (right), for TARGET (black line), CTRL (red line), ALB (blue line) and THCD (green line)  
 850 ensembles, in  $10^6 \text{ km}^2$ . Vertical bars indicate the 90% confidence intervals for the ensemble-means.

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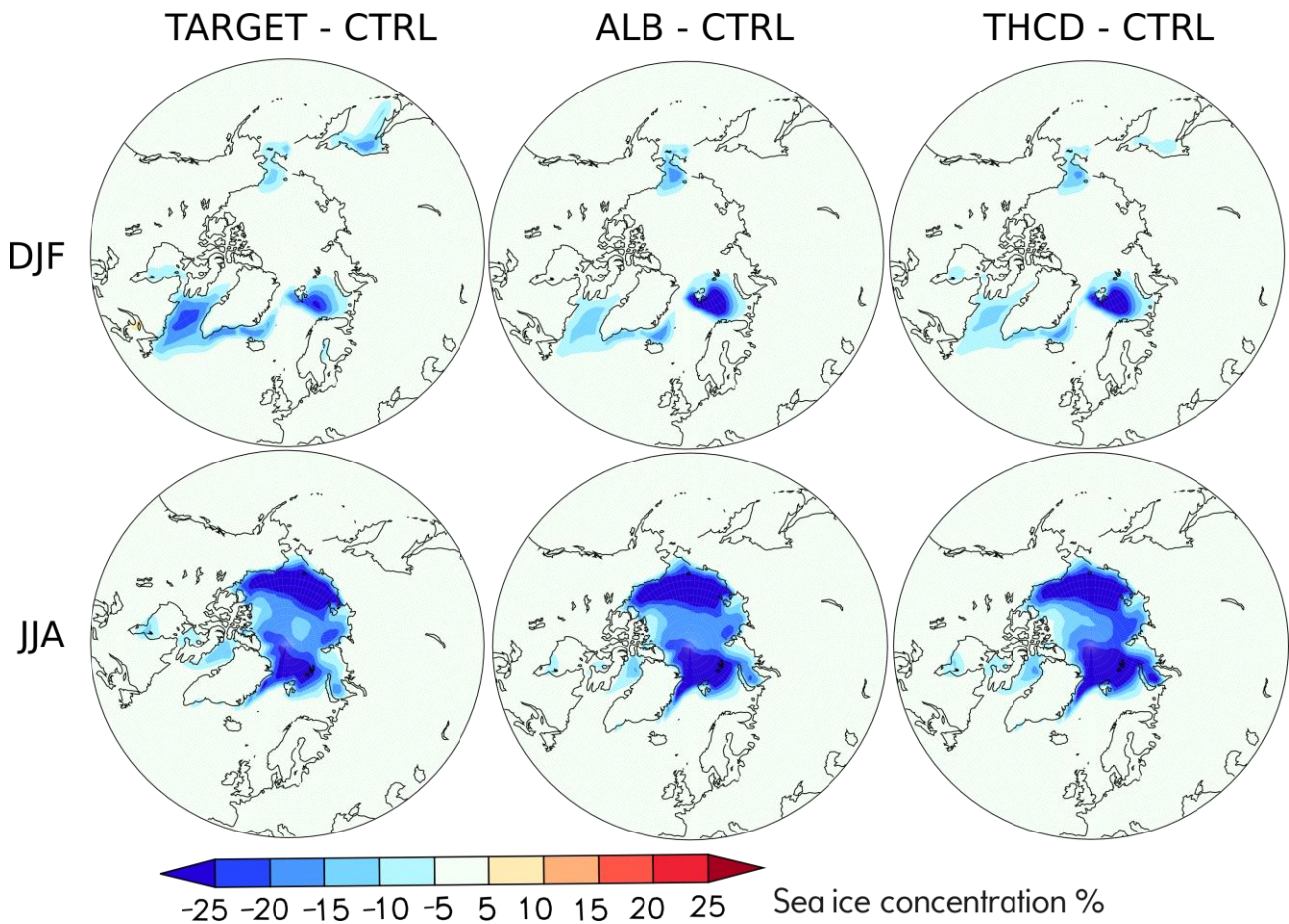
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864 Figure 4: Difference in sea-ice concentration compared to the CTRL simulation, for TARGET (left), ALB  
 865 (middle) and THCD (right), in %, averaged over December-January-February (DJF; top) and June-July-  
 866 August (JJA; bottom).

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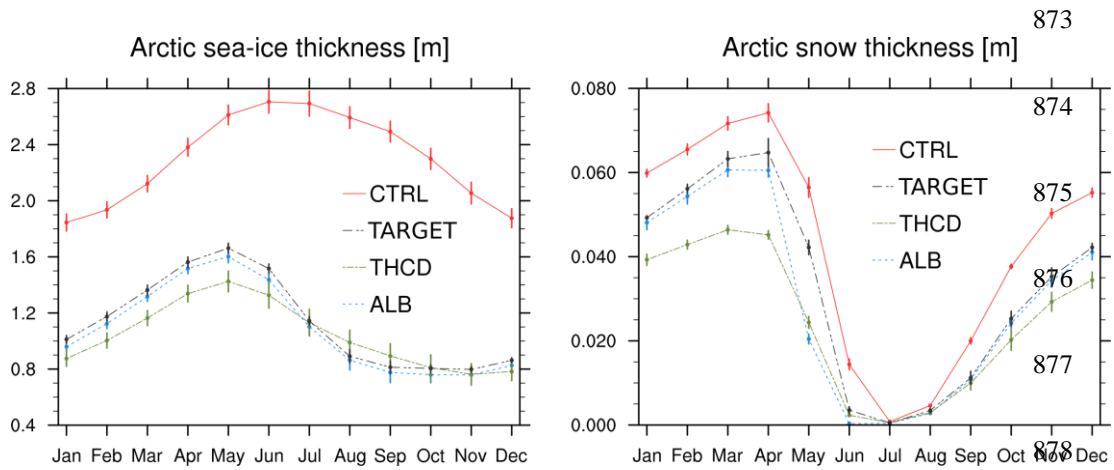
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879 Figure 5: Seasonal cycle of Arctic sea-ice (left) and snow on sea-ice (right) thickness, in m, for CTRL (red),  
 880 TARGET (black), ALB (blue) and THCD (green). Vertical bars indicate the 90% confidence interval for the  
 881 ensemble-mean.

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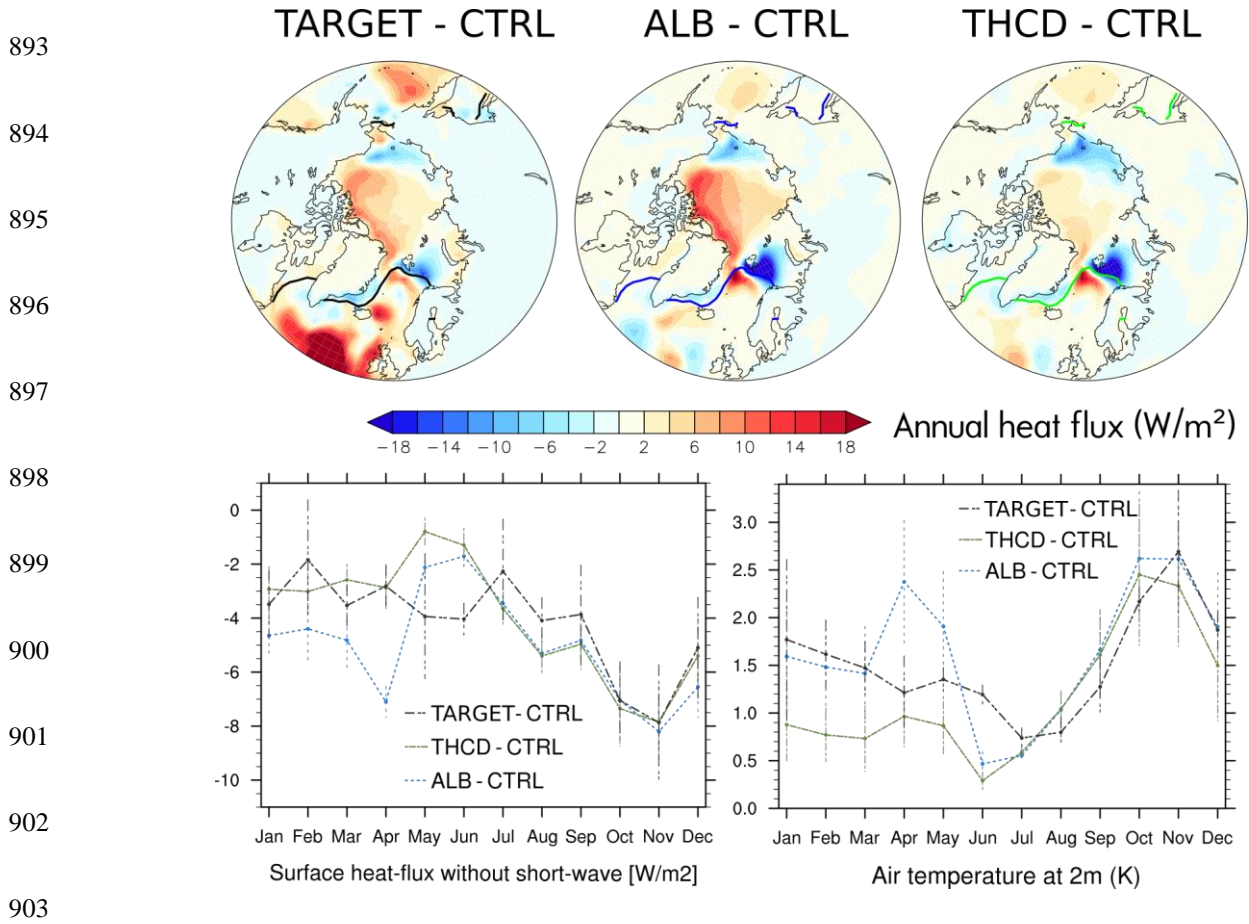
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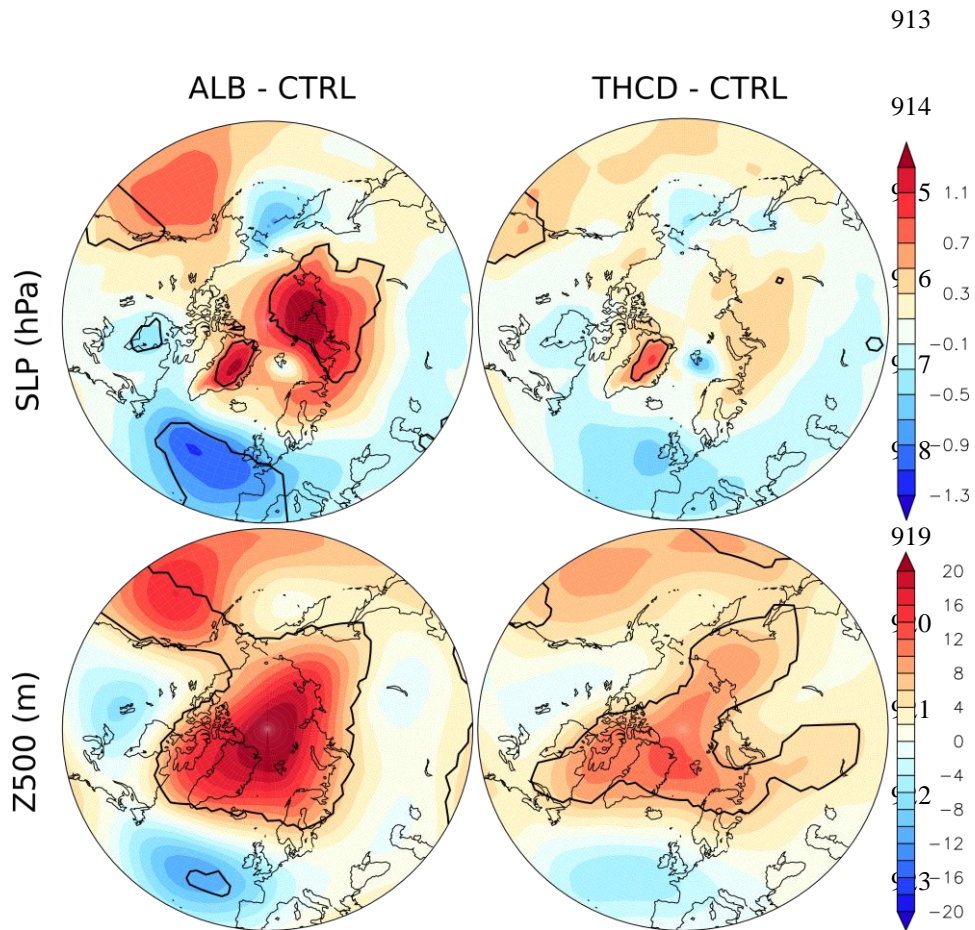


904 Figure 6: Anomalies of the annual-mean total heat flux with respect to CTRL (positive downward), in W/m<sup>2</sup>  
 905 for TARGET (top-left), ALB (top-middle) and THCD (top-right). The lines indicate the sea-ice edge (i.e. 15  
 906 % in concentration threshold) for the corresponding ensemble (black for TARGET, blue for ALB and green  
 907 for THCD). The anomalies north of this line are significant at the 90% confidence level. Mean seasonal cycle  
 908 of the anomalies with respect to CTRL, averaged north of 70°N for the surface heat flux without short-wave  
 909 (i.e, sensible, latent and long-wave heat fluxes; positive downward; bottom-left), in W/m<sup>2</sup> and air  
 910 temperature at 2 m (bottom-right), in K. Bars illustrate the 90% confidence interval for the ensemble-mean.

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925 Figure 7: Anomalies of sea-level pressure (SLP; top), in hPa and geopotential height at 500 hPa (Z500;  
 926 bottom), in m, averaged over December-January-February with respect to CTRL for ALB (left) and THCD  
 927 (right). Black lines indicate the 90 % confidence level.

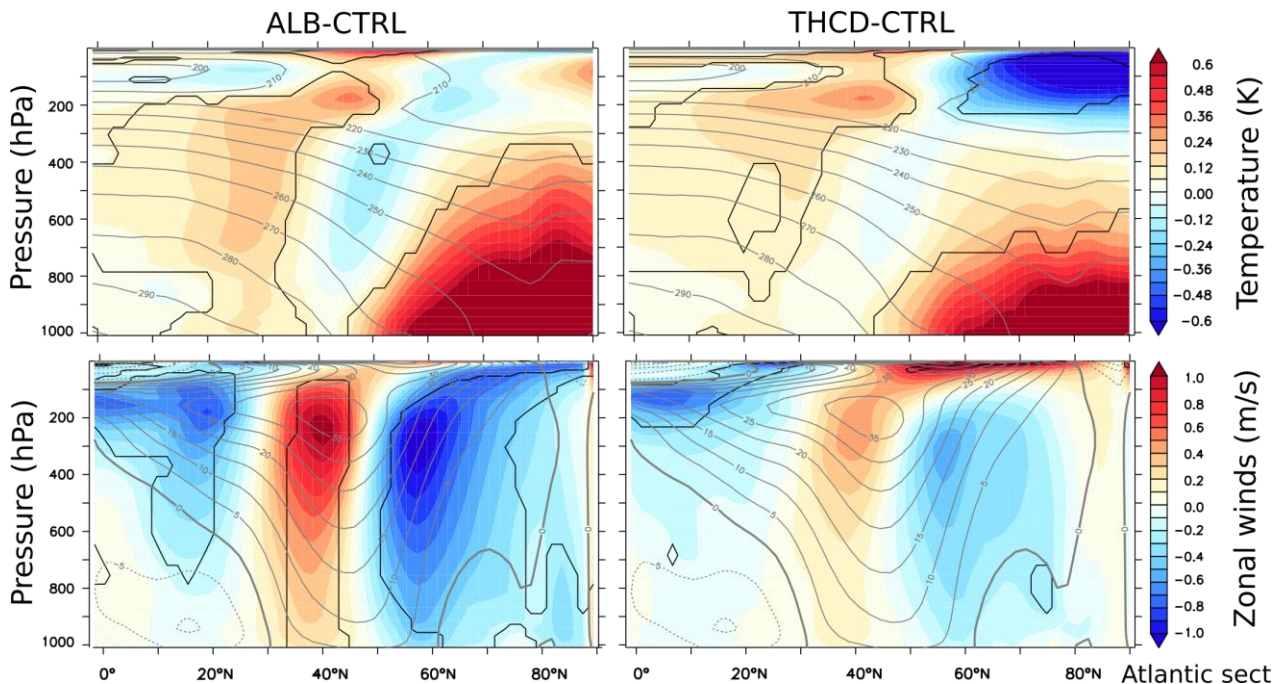
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934 Figure 8: Anomalies of zonal-mean air temperature (top), in K, and zonal-mean zonal wind (bottom), in m/s,  
 935 averaged over December-January-February (DJF) and over the North Atlantic sector (80°W-20°E) with  
 936 respect to CTRL for ALB (left) and THCD (right). Grey contours indicate the climatology and black  
 937 contours show the 90 % confidence level.

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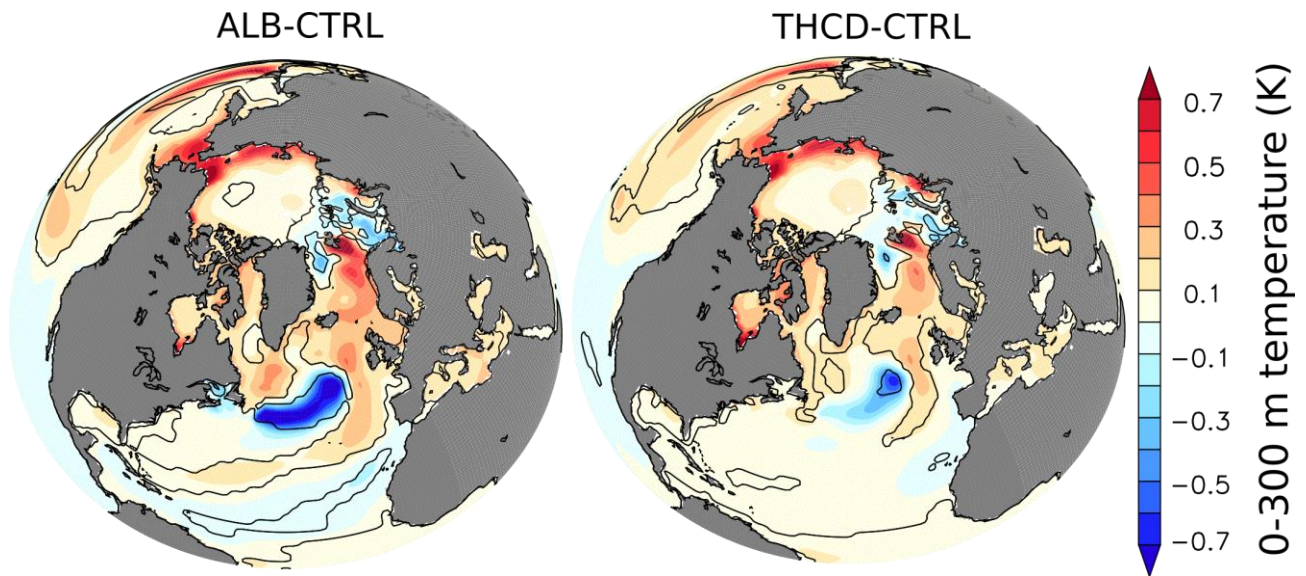
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947 Figure 9: Anomalies of the annual-mean ocean temperature averaged over the upper 300 m, in K, with  
948 respect to CTRL for ALB (left) and THCD (right). The 90% confidence level is depicted by the black  
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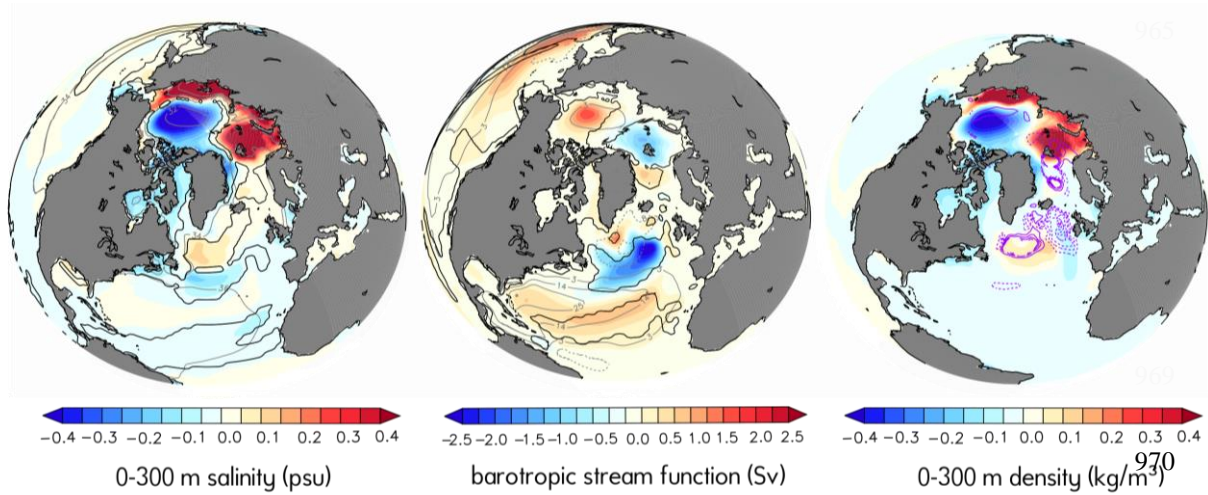
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972 Figure 10: Anomalies of the annual-mean salinity averaged over the top 300 m of the ocean (left), in psu, of  
 973 the barotropic stream function (middle; positive clockwise), in Sv, and of the density averaged over the top  
 974 300 m of the ocean (right) for ALB minus CTRL. Black contour defines the 90% confidence level, the mean  
 975 CTRL value is in gray contour. In the right panel, the mixed layer depth difference for ALB minus CTRL is  
 976 shown in purple line (dashed for negative) with the contour intervals as follow: (-140,-100,-60,-40,-  
 977 20,20,60,100,140), in m.

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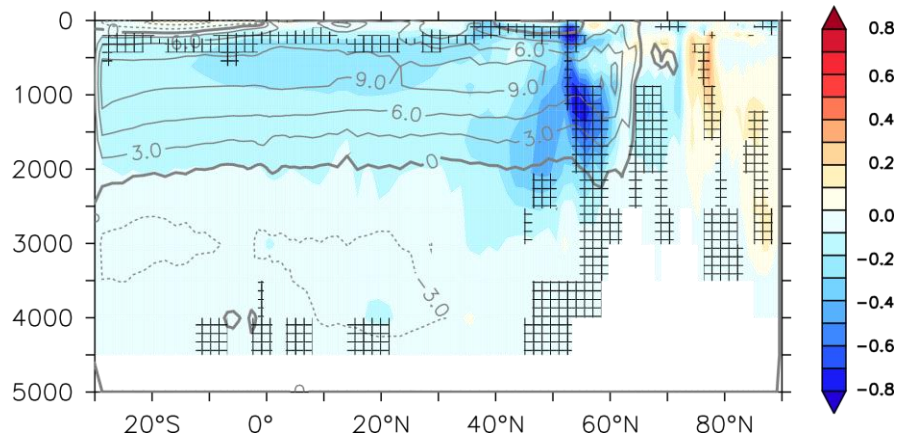
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992 Figure 11: Anomalies of the Atlantic meridional stream function, for ALB minus CTRL, in Sv. The mean  
993 AMOC of the CTRL simulation is superimposed (grey contours; positive clockwise) and hashes illustrate the  
994 anomalies with a confidence level larger than 90%.

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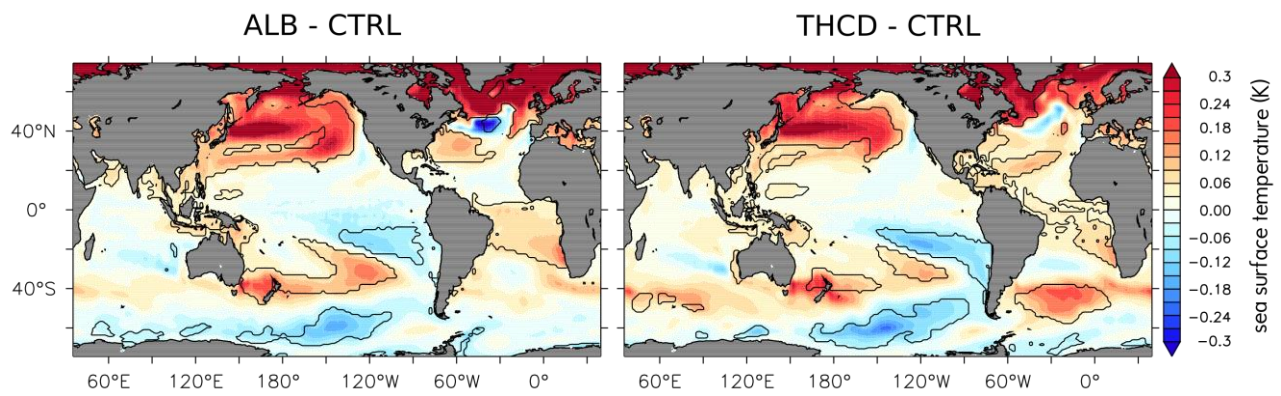
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1008 Figure 12: Anomalies of the annual-mean sea surface temperature, in K, with respect to CTRL for ALB (left)

1009 and THCD (right). Black contour shows the 90 % confidence level.

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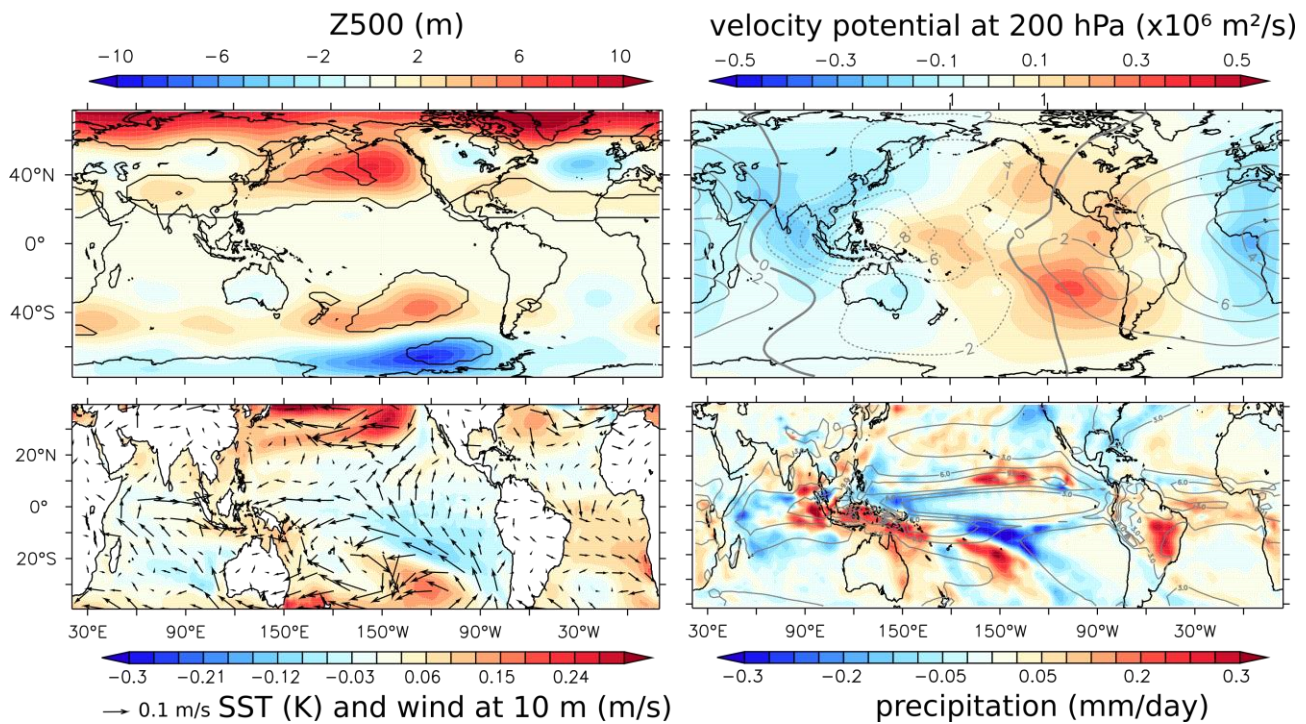
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## ALB - CTRL



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1016 Figure 13: Annual-mean anomalies of the geopotential height at 500 hPa (Z500; top-left), in m, velocity  
 1017 potential at 200 hPa (top-right), in  $10^6 \text{ m}^2/\text{s}$ , sea surface temperature, in K (shading) with the wind at 10 m  
 1018 (arrows; bottom-left), in m/s and precipitation (bottom-right), in mm/day, with respect to CTRL for ALB. In  
 1019 the upper left panel, the 90% confidence level is shown in black contour. The gray contour provides the  
 1020 corresponding value in CTRL in the right panels.