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► **To cite this version:**

Guillaume Gastineau, Javier García-Serrano, Claude Frankignoul. The Influence of Autumnal Eurasian Snow Cover on Climate and Its Link with Arctic Sea Ice Cover. *Journal of Climate*, 2017, 30 (19), pp.7599 - 7619. 10.1175/JCLI-D-16-0623.1 . hal-01669420v1

HAL Id: hal-01669420

<https://hal.sorbonne-universite.fr/hal-01669420v1>

Submitted on 20 Dec 2017 (v1), last revised 17 Jul 2018 (v2)

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2 **The influence of autumnal Eurasian snow cover**
3 **on climate and its link with Arctic sea ice cover**

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5

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16 May 15th, 2017

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21

22 Abstract:

23

24 The relationship between Eurasian snow cover extent (SCE) and Northern
25 Hemisphere atmospheric circulation is studied in reanalysis during 1979-2014 and in
26 CMIP5 preindustrial control runs. In observations, dipolar SCE anomalies in November,
27 with negative anomalies over eastern Europe and positive anomalies over eastern
28 Siberia, are followed by a negative phase of the Arctic Oscillation (AO) one and two
29 months later. In models, this effect is largely underestimated, but four models simulate
30 such relationship. In observations and these models, the SCE influence is primarily due
31 to the eastern Siberian pole, which is itself driven by the Scandinavian pattern (SCA),
32 with a large anticyclonic anomaly over the Urals. The SCA pattern is also responsible for
33 a link between Eurasian SCE anomalies and sea ice concentration (SIC) anomalies in the
34 Barents-Kara Sea.

35 Increasing SCE over Siberia leads to a local cooling of the lower troposphere, and
36 is associated with warm conditions over the eastern Arctic. This leads to a polar vortex
37 weakening in December and January, which has an AO-like signature. In observations,
38 the association between November SCE and the winter AO is amplified by the SIC
39 anomalies in the Barents-Kara Sea, where large diabatic heating of the lower
40 troposphere occurs, but SCE is the main driver of the AO. Conversely, the sea ice
41 anomalies have little influence in most models, which is consistent with the less
42 frequent SCA, the colder mean state, and the underestimation of troposphere-
43 stratosphere coupling in these models.

44

45 **1. Introduction**

46 The role of Arctic conditions in the mid-latitude winter climate is under debate,
47 especially for the North Atlantic sector (Overland et al. 2015). In this region, the
48 atmosphere has a dominant short-timescale chaotic intrinsic variability and is mainly
49 unpredictable. However, several studies suggest that the variability of Arctic sea ice
50 extent (Yamamoto et al. 2006; Francis et al. 2009; Honda et al 2009; Wu and Zhang
51 2010; Frankignoul et al. 2014; Garcia-Serrano et al. 2015, Koenigk et al. 2016, King et al.
52 2016) and Eurasian snow cover extent (SCE, e.g. Cohen and Entekhabi 1999, Cohen et al.
53 2007, Cohen and Jones 2011) have some influence onto the atmosphere during winter.
54 Such influence may account for an improvement in skill of long-range prediction due to
55 continental snow (Jeong et al., 2013, Orsolini et al., 2013) and sea ice (Scaife et al. 2014)
56 initialization and improved continental snow cover physics (Riddle et al. 2013) in
57 current forecast systems.

58 Continental snow cover affects the atmosphere via changes in surface albedo
59 (Cohen 1994). A larger snow cover increases the surface albedo and reflects shortwave
60 radiation away from the surface (Gong et al. 2004, Jeong et al., 2013). A snowpack also
61 insulates the atmosphere from the soil surface. In winter at high latitude, these two
62 effects explain that snow enhances the diabatic cooling at the surface and in the
63 atmospheric boundary layer (Fletcher et al. 2007; Dutra et al. 2011), which locally
64 increases the sea level pressure (SLP). A larger SCE over Eurasia has been reported to
65 intensify and expand the Siberian high (Jeong et al., 2011; Orsolini et al., 2013). This
66 modifies the land/sea contrast and the stationary wave pattern, and may lead to
67 enhanced upward planetary wave propagation, thus weakening and warming the polar
68 vortex in the stratosphere (Saito et al., 2001, Cohen et al. 2007, Orsolini et al., 2016). A
69 weak polar vortex can persist for several weeks and influence the underlying

70 troposphere by downward propagation of circulation anomalies. The influence of the
71 Eurasian snow cover has received most attention in autumn, as it shows a statistically
72 significant relation with the following winter Arctic Oscillation (AO) and North Atlantic
73 Oscillation (NAO), from December to March (Cohen et al., 2007; Déry and Brown, 2007;
74 Allen and Zender, 2010; Cohen et al., 2012).

75 Sea ice concentration (SIC) changes may also influence the atmosphere. The most
76 reported influence concerns SIC in the Barents-Kara Sea, where SIC in autumn has a
77 statistically significant influence on the following winter NAO (Petoukov and Semenov,
78 2010; Kim et al., 2014; Garcia-Serrano et al., 2015; King et al., 2016). Sea ice insulates
79 the ocean from the atmosphere, so that a sea ice loss increases the heat flux from the
80 ocean to the atmosphere. The resulting diabatic heating is large, but localized near the
81 sea ice edge (e.g. Magnusdottir et al. 2004; Deser et al. 2004, 2007). This leads to
82 changes in the tropospheric eddies and the planetary wave pattern, which may alter the
83 polar vortex (e.g. Nakamura et al. 2015, 2016). The modified polar vortex may then
84 influence the troposphere by downward propagation in the following weeks or months,
85 with important impacts during periods of polar vortex breakdown, such as in February
86 (Jaiser et al. 2016).

87 The influence of SIC thus shares a large similarity with that of the Eurasia SCE
88 during fall (October and November), as both may involve a stratospheric pathway.
89 Furthermore, continental SCE and Arctic SIC are linked, as a reduced Arctic sea-ice
90 extent leads to a moistening of the atmospheric boundary layer, which increases the
91 moisture flux into eastern Siberia, increasing snowfall, as suggested by Cohen et al.
92 (2014a) and found by Wegmann et al. (2015) using a Lagrangian analysis. The sea ice
93 and snow cover are also connected by the influence of Ural Blocking, which has been
94 reported to cause warm Arctic–cold Eurasia anomalies in winter (Luo et al., 2016). The

95 two surface influences are, therefore, connected, and their interaction might amplify the
96 atmospheric response found by separately considering snow cover and sea ice (Cohen et
97 al. 2014a). However, only a few studies have investigated the links between the SCE and
98 sea ice. Moreover, the relative effect on the atmosphere of the Arctic sea ice and
99 Eurasian snow cover is largely unknown. In addition, the influence of tropical SST
100 variability needs to be clarified, as the tropical teleconnections may both influence the
101 snow cover over Eurasia and modify the atmospheric circulation (Fasullo, 2004), leading
102 to a possible confusion between cause and effect.

103 As the observational record is mostly limited to the recent decades, climate
104 models can be used to investigate the impact of SIC and SCE variability with a much
105 larger sampling, even if the stratospheric polar vortex is too stable in models, which may
106 inhibit the troposphere-stratosphere coupling (Furtado et al., 2015). The aim of this
107 study is to investigate the influence of autumnal Eurasian snow cover variability in
108 observations and climate models, and the links with that of the sea ice cover. We find
109 that snow cover anomalies in November have a dominant influence on the atmospheric
110 circulation in observations and several models. The SCE anomalies are found to be
111 associated with SIC anomalies over the Barents-Kara Sea, as both are modulated by the
112 Scandinavian pattern, which is the dominant mode of atmospheric variability in
113 November.

114 The next section describes the methodology. The analysis of the snow cover and
115 its links with the atmosphere is discussed in Section 3. The processes linking the snow
116 cover to the atmosphere are investigated in Section 4. Finally, conclusions and a
117 discussion are provided in last section.

118

119

120 **2. Data and methods**

121 *a. Observations*

122 Monthly sea ice cover is downloaded from the NOAA/National Snow and Ice Data
123 Center (Comiso, 2012). Weekly Northern Hemisphere continental snow cover is
124 retrieved from the NOAA/Rutgers University Global Snow Laboratory, and aggregated
125 into monthly data. Both products are based on passive microwave measurements
126 (SSM/I) and extend from 1979 to 2014. The sea-level pressure (SLP), geopotential
127 height, air temperature, and heat flux (accumulated from 24h forecasts) are from the
128 ERA-Interim reanalysis (Dee et al., 2011).

129 A quadratic trend is removed from all variable before the analysis to remove the
130 effect of the global warming. This also removes the multi-decadal variability and lower
131 frequencies, and the large Arctic sea ice decrease from 2005 onward (e.g. Close et al.
132 2015).

133

134 *b. Models*

135 Monthly SLP, snow cover, geopotential, SIC, SST and heat fluxes anomalies are
136 downloaded from the CMIP5 archive for 12 coupled ocean atmosphere models (Table 1)
137 using the preindustrial multi-centennial control simulations with constant external
138 forcing. All model fields are interpolated onto a common 2.5°x2.5° horizontal grid. A
139 quadratic trend was removed from all outputs to remove the possible influence of model
140 drift.

141

142 *c. Maximum covariance analysis*

143 Maximum covariance analysis (MCA) is used to estimate the main modes of area-
144 weighted covariability between the atmosphere and the underlying snow cover. We use

145 snow cover anomalies over northern Eurasia (40°N - 65°N ; 0°E - 180°E). The SLP
146 anomalies in the Northern Hemisphere (20°N - 90°N) are chosen to represent the
147 tropospheric circulation. The MCA decomposes the covariance matrix of the two fields
148 using singular value decomposition (Bretherton et al., 1992). Each mode of covariability
149 is characterized by two times series and associated spatial patterns. Here, the MCA time
150 series are standardized (divided by their standard deviation). The spatial patterns are
151 illustrated by the homogeneous covariance map for the field that leads (regression on
152 the same field time series) and the heterogeneous covariance map for the field that lags
153 (regression on the MCA time series of the other field), which preserves orthogonality
154 (Czaja and Frankignoul, 2002). The MCA modes are characterized by their normalized
155 squared covariance (NSC, i. e. the squared singular value divided by the variance of both
156 fields), the correlation (R) between the MCA time series, and the squared covariance
157 fraction (SCF, i. e. the ratio of covariance explained). In order to evaluate the robustness
158 of the MCA modes, we repeated the MCAs using 100 random permutations of three-
159 years blocks for the SLP field. The number of NSC and R that exceed the observed values
160 gives the levels of significance for NSC and R.

161 The mode of covariability between the snow cover and the atmosphere are
162 expected to reflect the influence of atmospheric perturbations on the SCE when the two
163 fields are in phase or, because of snow cover persistence, when the atmosphere leads.
164 When the snow cover leads the atmosphere by one month or more, a significant MCA
165 mode could indicate an influence of the snow cover (or concomitant boundary forcing)
166 on the atmosphere, as the extratropical atmosphere has an intrinsic persistence of at
167 most 10 days (Vautard, 1990). However, the El Niño Southern Oscillation (ENSO) has
168 persistent remote teleconnections that may give rise to persistent MCA modes not solely
169 linked to local boundary forcing. Hence, we (largely) remove these teleconnections from

170 both snow and atmospheric data by multivariate regression when (and only when) the
171 snow cover field leads the atmosphere, assuming that they lag the tropical Pacific SST by
172 two months in the atmosphere, while they vary with lag for the snow in order to get
173 unbiased estimates (see Frankignoul et al., 2011). The tropical SST variability is
174 represented by the first three empirical orthogonal functions (EOFs) of the monthly
175 tropical Indo-Pacific SST. The regressions are performed separately for each season, to
176 account for the seasonal changes of the ENSO teleconnection, and separately for positive
177 and negative values of the Principal Components (PCs), to account for the asymmetry
178 (see supplemental material text for details). We verified that similar MCA results are
179 obtained by assuming a one-month lag for the ENSO teleconnections, or even without
180 removing the ENSO signal (see Table S1).

181

182 *d. Rotated empirical orthogonal function*

183 The main patterns of Northern Hemisphere (20°N - 90°N) SLP variability are
184 given by rotated empirical orthogonal function (REOFs) analysis, using the first 15 EOFs
185 in the rotation, which accounts for 95% of the variance. To preserve orthogonality of the
186 PCs, we scaled the EOFs by the square root of its eigenvalue before performing the
187 varimax rotation (Kaiser 1958). The rotated PCs are standardized, and the REOF
188 patterns are given by regression on these time series.

189

190 *e. Regression analysis*

191 We used both univariate and multivariate least squares regression. We remove
192 the tropical teleconnections from all data before the regression analysis, following the
193 same methodology as the MCA (see section 2.c). The level of statistical significance is
194 tested with 100 permutations of the atmospheric fields in 3-yr blocks to take serial

195 autocorrelation into account. The number of permuted regression slopes that exceeds
196 the observed value provides the p-value.

197

198 **3. The links between Eurasian snow cover and the atmosphere**

199 *a. Detection of the snow cover influence*

200 The normalized squared covariance (NSC) of the first MCA mode provides an
201 estimate of the dominant covariability between the SCE and SLP anomalies. It is shown
202 as a function of lag and season for the observations in Fig. 1. The largest NSC are mostly
203 obtained when the atmosphere is in phase with the SCE or leads it by one month
204 (negative lag), reflecting that the atmosphere controls the formation of snow cover
205 anomalies. The largest covariability occurs for SLP in March at lag 0 and for SLP in
206 February when it leads by one month. This is consistent with the occurrence of the
207 largest interannual snow anomalies in March, and the largest atmospheric variability in
208 February.

209 At positive lag, the snow cover leads the atmosphere, which may reflect the SCE
210 forcing of atmospheric anomalies. The most significant links are found between
211 November snow cover and SLP in December (lag 1) and January (lag 2), as well as
212 between February snow cover and SLP in March (lag 1), as the NSC and R are both
213 significant at the 5% level (Fig. 1). The covariability is weaker when October SCE leads
214 the atmosphere, whether by 1, 2 or 3 months (p-values are 10%, 28%, 40% for NSC and
215 13%, 38%, 20% for R). Our results thus contrast with the commonly argued impact of
216 October Eurasian snow cover on winter SLP (Saito and Cohen, 2003), as further
217 discussed in Appendix. A significant covariance (p-value<10%) is also found for SLP in
218 August and September, when the snow leads by one month.

219 The influence of November SCE onto the atmosphere in December and January is
220 the main focus of this paper, and it is discussed below. The late winter snow influence
221 found in March has been reported in several studies (Barnett et al., 1989; Saito and
222 Cohen, 2003; Zhang et al., 2004; Peings and Douville, 2010; Peings et al. 2011); it is not
223 investigated here, as the processes are different from the fall influence studied here.
224 Similarly, the covariability in late summer is not discussed here; it shows a reduction of
225 snow cover in south-western Norway preceding anticyclonic conditions over the North
226 Atlantic (not shown), and might be due to concomitant North Atlantic SST forcing
227 (Gastineau and Frankignoul, 2015).

228 The same analysis has been performed with the CMIP5 models, and a significant
229 covariability between SCE and SLP anomalies is found in several cases. The results are
230 summarized in Fig. 2, which shows the level of statistical significance of the NSC and R
231 for the first MCA mode (left panel). The similarity with the observational data is given
232 by the spatial pattern correlation of the homogeneous SCE and heterogeneous SLP
233 covariance maps between each model and the observation (right panel). When using
234 November SCE anomalies and December SLP (black symbols in Fig. 2), there are four
235 models out of 12 (CanESM2, MPI-ESM-LR, GISS-E-R and CESM1) suggesting an impact of
236 the November SCE anomalies that is reasonably similar to that observed (spatial
237 correlation between 0.2 and 0.9). These four models show a first MCA mode that is 10%
238 significant for NSC and R, except for MPI-ESM-LR, which is only 12% significant for R.
239 Among these four models, only CESM1 is a low-top model, while the others are high top
240 models with lid height above 45km (Seviour et al., 2016).

241 The SCE influence seems to be less persistent in models, as the first MCA mode
242 with November SCE is only significant at lag 2 (SLP in January) in CESM1 (red symbols
243 in Fig. 2), as opposed to observations (Fig. 1). When using October SCE and November

244 SLP (blue symbols in Fig. 2), there are only two models out of 12 suggesting an impact of
245 the October snow cover anomalies (CSIRO-Mk3-6 and CCSM4). When using October SCE
246 and December (January) SLP, only one model, FGOALS-g2 (IPSL-CM5A-LR), provides a
247 potential impact. We conclude that consistent with observations, more CMIP5 models
248 suggest an impact of November SCE than October SCE. Next, we will discuss the spatial
249 patterns corresponding to these modes of covariability.

250

251 *b. Spatial pattern of the November snow cover influence*

252 The covariance maps for November SCE and December SLP are shown in Fig. 3. In
253 observations, the first MCA mode shows dipolar snow cover anomalies (Fig. 3a, colors),
254 with a pole over eastern Europe and an opposite polarity over south-eastern Siberia,
255 Northern Mongolia, and Northern China. Both poles are located at the margin of the
256 snow-covered surface in November (see Fig. S1). This SCE dipole precedes SLP
257 anomalies (black contours) broadly projecting on a negative phase of the AO, with a
258 large signature over the North Atlantic. The covariance maps at lag 2 (Fig. 3b, November
259 SCE / January SLP) are almost identical, but the SLP anomalies are weaker, especially
260 over Western Europe. Note that the covariance maps at lag 3 (November SCE / February
261 SLP) are also similar, although the significance level for NSC and R are 1% and 27%,
262 respectively.

263 The MCA patterns in the four CMIP5 models (CanESM2, MPI-ESM-LR, GISS-E-R,
264 CESM1) identified previously are broadly similar to the observed ones (Fig. 3c-f), with a
265 positive snow cover anomaly in southern Siberia and a negative one over eastern
266 Europe preceding a negative AO-like pattern by one month. However, the amplitudes
267 are smaller than in observations (note the different color and contour interval in Fig. 3).
268 Furthermore, the snow cover anomalies are slightly shifted, as the November SCE

269 climatology shows less snow over Eurasia, especially over Europe (Fig. S1). In the
270 following we only consider this subset of four models, as illustrated by the averaged
271 covariance map (Fig. 3g).

272 To take into account the different sampling in models (≥ 500 yr) and
273 observations (36 yr), we performed similar MCA analysis on separate 36-yr segments
274 from each of the four model simulations. These 36-yr segments are selected using a shift
275 of 6 years between two consecutive ones, so that for instance a 1000-yr run results in
276 160 36-yr segments. The mean NSC and R for the first MCA mode in these segments are
277 larger than the ones computed from the entire run (compare Fig. 3h and values on top of
278 Fig. 3c-f), but still smaller than in observations, with the 95% percentile of their
279 distributions lower than the observed value. Therefore, it is very likely that the models
280 do underestimate the snow influence.

281

282 *c. Origin of the snow cover dipolar variability in November*

283 To determine the origin of the dipolar snow cover anomalies, November SLP and
284 2m air temperature anomalies are regressed onto the (standardized) MCA time series of
285 November SCE, referred to as MCA-snow (Fig. 4). For the CMIP5 models, we only
286 consider the four models (CanESM2, MPI-ESM, GISS-E2-R and CESM-BGC) that are
287 consistent with observations and show the multi-model average of the regression
288 patterns, while the number of models with a regression of the same sign documents
289 their robustness, and provides a measure of inter-model spread.

290 The SLP anomalies associated with the snow dipole in both observations (Fig. 4a)
291 and models (Fig. 4b) are characterized by a large anticyclonic anomaly over the Urals
292 and a depression over Europe. The SLP pattern shares some similarity with the Eurasian
293 pattern type 1 (Barnston and Livezey, 1987), the Scandinavian pattern (Bueh and

294 Nakamura 2007), the Russian pattern (Smoliak and Wallace, 2015) or the anomalies in
295 Ural Blocking conditions (Luo et al., 2016). A similar pattern was also reported to result
296 from the October SCE response (Cohen et al., 2014b). We will refer to this atmospheric
297 patterns as the Scandinavian pattern (SCA) in the following. Figure 4 illustrates that
298 warm (cold) air temperature anomalies are associated with negative (positive) SCE
299 anomalies, consistent with the warm (cold) advection by the anomalous atmospheric
300 circulation, as in the Greenland, Barents and Kara Seas that are affected by warm
301 advection from the Norwegian Sea.

302 In observations, a dipolar SCE pattern similar to that in Fig. 3a and a SCA-like SLP
303 pattern is also obtained as first MCA mode of simultaneous SLP and SCE anomalies in
304 November, with 42.1% of squared covariance fraction (SCF), as shown in (Fig. 5a), while
305 an AO influence onto the snow cover is only obtained as mode 3 (SCF = 11.6%). This is
306 consistent with the first REOF of November SLP, which corresponds to the SCA (Fig. 6a).
307 In December, however, the simultaneous covariability between SLP and SCE is
308 dominated by the AO (SCF=55.1%, Fig. 5b), which decreases the advection from the
309 relatively warm ocean toward the cooler Eurasian Continent. It also shifts southward
310 the precipitation associated with the Atlantic stormtrack (Hurrell, 1995), which
311 increases the SCE over Europe. We also see negative SCE anomalies east of the Caspian
312 Sea associated with warm advection from the Mediterranean region.

313 On the other hand, the MCA suggests that, in most of the four models, the AO
314 already has the largest impact on snow cover in November (Fig. 5c), with a much larger
315 impact downstream of Europe, as shown by the positive anomalies over Eastern Siberia.
316 Only CESM1 simulates the SCA pattern and its dipolar snow cover signature as first MCA
317 mode (not shown). In fact, the first REOF of November SLP is also AO-like in all models
318 (Fig. 6b). To establish its robustness, we have used as above distinct 36-yr chunks from

319 each control simulation, to reproduce the observed sampling. The SCA and AO are
320 identified using the largest spatial pattern correlation with the observed SCA (November
321 REOF1) and AO (November REOF3), respectively. The AO variance fraction is
322 systematically larger than observed (Fig. 6c, yellow), while the SCA one is smaller (Fig.
323 6c, red). This is consistent with the larger (smaller) role of the SCA (AO) in the
324 observation, when compared to model simulations, and it can be explained by either
325 natural atmospheric variability or model biases. Indeed, CMIP5 models use relatively
326 coarse horizontal resolutions, and are known to underestimate winter blocking episodes
327 (Dawson et al, 2012), leading to an overestimation of the NAO regimes (Cattiaux et al.,
328 2013).

329

330 **4. Processes of the November snow cover influence**

331 *a. Role of Siberian snow cover*

332 The relative importance of the two poles of the November SCE dipole can be
333 analyzed using two indices: the mean SCE anomalies over eastern Europe (20°E-58°E,
334 48°N-60°N) and over eastern Siberia (70°E-140°E, 43°N-56°N). A bivariate regression of
335 SLP anomalies in December on these two indices shows significant SLP anomalies in the
336 observations (Figs. 7a and 7b), with negative SLP anomalies off Western Europe and
337 positive anomalies over the polar cap. However, the eastern Siberia pole has the largest
338 and most significant influence on SLP, and its impact is more AO-like. In the four models
339 (Figs. 7c and 7d), Siberian SCE anomalies also have a larger, AO-like influence on SLP,
340 while European SCE is linked to a weak SLP dipole between Greenland and Scandinavia.
341 Therefore, the most robust signal seems to be linked to the Siberian SCE influence,
342 which is consistent with the reported influence of October snow cover (Saito and Cohen,
343 2003).

344

345 *b. Associated surface changes*

346 The influence of surface conditions is evaluated using SCE and SIC regressions
347 onto MCA-snow in Fig. 8. The SCE anomalies in November (Fig. 8c,d) are preceded in
348 October (Fig. 8a,b) and followed in December (Fig. 8e, f) by similar, but smaller,
349 anomalies over eastern Siberia, which is consistent with the snow cover persistence
350 over that region (Déry and Brown, 2007), and reflected in the large correlation (around
351 0.5) between October and November SCE (see Fig. S1). European SCE anomalies are also
352 present from October in the models, but not in observations. A significant retreat of the
353 sea ice edge in the Barents Sea is also found for both models and observations in
354 October and November, which is also visible in December in the models.

355 The surface heat flux in lead and lag conditions can be used to discuss the
356 processes leading to the atmospheric circulation response. The heat flux preceding the
357 SCE is dominated by the atmospheric forcing of the snow cover, as for SST anomalies,
358 while the heat flux lagging the SCE should primarily reflect the heat flux directly forced
359 by the SCE (the thermodynamical component), although it could be strongly affected by
360 the surface heat flux intrinsically associated with the atmospheric response (hereafter
361 the dynamical heat flux component); at lag 0, both effects play a role and may even
362 cancel (Frankignoul et al. 1998). Since the surface heat flux responds rapidly to the
363 surface conditions (simultaneously on monthly timescale), one can use in-phase
364 relations to estimate the (thermodynamical) heat flux driven by the SCE anomalies, if
365 the (larger) dynamical component is removed. To do so, we first calculate the heat flux
366 by adding surface radiative and turbulent fluxes. A standardized atmospheric index,
367 referred to as ATM, was computed by projecting the November SLP anomalies over
368 30°N-90°N 80°W-180°E onto the SCA-like patterns shown in Fig. 4. The dynamical heat

369 flux component corresponding to one standard deviation of the MCA-snow index is
370 obtained by regressing the heat flux anomalies onto ATM, multiplied by the correlation
371 between ATM and MCA-snow (shown in Fig. S2). The total heat flux anomaly associated
372 with the SCE pattern in Fig. 3a is given by the regression of the heat flux onto MCA-snow
373 (shown in Fig. S3), while the difference of the two (Figs. 9a and 9b) is an estimate of the
374 thermodynamical effect. Figs. 9c-d show the net heat flux components integrated over
375 three boxes (see purple boxes in Fig. 9a-b) located over Siberia, Europe, and the Barents
376 and Kara Seas. The location of the boxes was adjusted to capture the snow and sea-ice
377 influences in models and observations.

378 In November, the heat flux changes induced by the snow cover are downward
379 over a wide latitudinal band in central Siberia from lake Balkhash to Sakhalin Island in
380 ERA-Interim and models (Fig. 9a-b), although the results are noisy in ERA-Interim. This
381 is consistent with a net cooling effect of positive snow cover anomalies, as the larger
382 surface albedo leads to more reflected shortwave radiation, and as the surface may be
383 more insulated from the warmer soil if the snow depth also increases (Orsolini et al.,
384 2016). The cooler surface temperature results in a dominant reduction of longwave
385 radiation and sensible heat flux. However, the turbulent fluxes have a larger
386 contribution in models, while the longwave and shortwave components dominate in
387 observations (Fig. 9c and 9d). Conversely, the heat flux anomalies are upward in ERA-
388 Interim over eastern Europe and Scandinavia where the SCE decreases, while in models,
389 there is almost no net heating effect. Interestingly, over the Barents-Kara Seas, the heat
390 flux is mainly upward over open-water in the Nordic Seas, which suggests a large
391 heating of the atmosphere where the sea ice has retreated in November. This is
392 consistent with an active influence of SIC anomalies onto the lower troposphere.
393 However, while the total heat flux release over the Barents-Kara Seas is dominant in

394 ERA-Interim, it is smaller and less robust in models. The same analysis applied to the
395 December heat flux provides comparable results over Europe and Siberia (see Fig. S4),
396 but the heating over the Barents-Kara Seas is larger in models, while a net cooling is
397 obtained in observations. This is because the sea-ice anomalies persist in December in
398 models (see Fig. 8f), while they vanish in ERA-Interim (Fig. 8e).

399 In summary, the diabatic forcing of SCE anomalies is consistent in models and
400 ERA-Interim, with cooling when the SCE increases. However, the diabatic heating from
401 the SIC anomalies over the Barents-Kara Seas is larger but it is also less robust than the
402 one associated with SCE. As the surface heat flux anomalies are not assimilated in ERA-
403 Interim and largely depend on the model physics, these results might be model
404 dependent.

405

406 *c. Troposphere-stratosphere coupling*

407 We calculated the regressions of the SLP (Fig. 10), zonal-mean temperature and
408 geopotential height (Fig. 11) onto November MCA-snow, from October to January. In
409 observations, the November SCE anomalies are preceded in October by a small
410 anticyclone centered over the northern coast of Siberia (Fig. 10a), as in Cohen et al.
411 (2002). In November, one month later, the SCA pattern (Fig. 10c) is visible, with cold
412 tropospheric anomalies over Eurasia between 40°N and 60°N, above the positive SCE
413 anomalies, and warm tropospheric anomalies at 78°N, at the location of the Barents-
414 Kara Seas (Fig. 4). The zonal mean anomalies are largely barotropic below 300-hPa,
415 which illustrates the main role of the tropospheric eddies in settling the SCA pattern.
416 The anomalous anticyclone over Eurasia has been interpreted as a response to October
417 Siberian snow cover, the snow-induced cooling acting to reinforce and expand westward
418 the Siberian High (Cohen et al., 2007; Jeong et al., 2011; Orsolini et al., 2013). However, it

419 can also be interpreted as a result of the stationary Rossby wave induced by the
420 anomalous turbulent heat flux from the sea ice retreat in the Barents-Kara Seas (e.g.
421 Honda et al. 2009; Garcia-Serrano et al. 2015), or as internal atmospheric variability
422 since simultaneous relations primarily show the SCE forcing by the SCA. In the lower
423 stratosphere, there is a warming over the polar cap (75°N-90°N, between 300-hPa and
424 100-hPa) and positive geopotential height anomalies above (Fig. 11a) that depicts a
425 weakening of the polar vortex. In December, one month later, a barotropic negative
426 NAO/AO pattern appears in the Euro-Atlantic region (Fig. 10e), while the polar vortex is
427 further weakened, with stratospheric temperature anomalies above 100 hPa that are
428 only significant between 40°N and 65°N (Fig. 11c). The regressions are similar in
429 January, with the SLP anomalies projecting on the AO (Fig. 10g), and stronger zonal-
430 mean geopotential height and temperature anomalies (Fig. 11e).

431 In the CMIP5 models, the atmospheric anomalies in October (Fig. 10b), which
432 precedes by one month the SCE anomalies, show alternating trough and ridges from the
433 North Atlantic to south-eastern Asia, with anticyclonic anomalies over the Urals and a
434 depression over Northern Europe, clearly indicative of a stationary wave and already
435 reminiscent of the SCA pattern. In November, the anomalies are more complex and
436 larger, with a dominant anticyclonic circulation over the Urals extending into the Arctic
437 (Fig. 10d), so that the Siberian High is clearly intensified and shifted westward, while the
438 SLP response is AO-like in December and, to a lesser extent, in January. The temperature
439 anomalies show a large warming in the lower troposphere north of 70°N (Fig. 11b, d)
440 from November to December, and display an important warming in the polar
441 stratosphere that persists into January only in the lowermost stratosphere at 200-hPa.
442 The warm anomalies are rather baroclinic in the polar troposphere, which is consistent
443 the influence of Arctic SIC reduction noted in Cattiaux and Cassou (2013). In November

444 and December, there are also cold temperature anomalies below 400-hPa south of the
445 positive SCE anomalies, likely associated with the cold temperature found over Siberia
446 where the snow cover increases (Fig. 10b,d). In models, both the tropospheric NAO/AO
447 pattern and the anomalies in the stratosphere are smaller during January, but they are
448 still significant (Fig. 10h and 11f).

449 The troposphere-stratosphere coupling is further illustrated by the polar cap
450 temperature (65°N-90°N) regression onto the MCA-snow index in Fig. 12. For
451 observations, the daily air temperature was used, while only monthly data was available
452 for models. The observations show a warming in the lower stratosphere between 200-
453 hPa and 70-hPa from December to February, as found by Cohen et al. (2014b) and
454 Orsolini et al. (2016), but it is only 10% significant for a few days in early December and
455 January. There are also hints of downward propagation in late December and late
456 January. In models the polar cap temperature anomalies are only half the ones observed,
457 the timing is different as the warming starts in November, one month earlier, and the
458 downward propagation is faster in the stratosphere with little penetration into the
459 troposphere.

460 In summary, the diabatic heating from the November SCE and, possibly, SIC
461 anomalies is associated with a stationary wave pattern that weakens the polar vortex.
462 Particularly in observations, the AO changes obtained one and two months later are
463 consistent with the downward propagation of polar vortex weakening. Next, we will
464 establish the relative importance of the SIC and snow cover anomalies.

465

466 *d. Link with sea ice anomalies*

467 In order to compare the role of SIC and SCE, we also perform a MCA using SIC
468 over the Barents-Kara Sea (65°N-85°N; 15°E-100°E) in November and SLP in December.

469 We additionally perform a MCA using both November SIC and SCE concatenated into a
470 single predictor field, with SLP as predictand field. The results are summarized in Table
471 2. When only using November SIC as predictor, the NSC is highly significant, but the
472 correlation R is lower than when using SCE, and not significant at the 10% level, as in
473 Garcia-Serrano et al. (2015; see also Fig. S5). On the other hand, using concatenated SCE
474 and SIC predictors is as significant as with SCE alone, and the MCA patterns (Fig. 13a)
475 show that the snow dipolar anomalies and the sea ice retreat in the Barents-Kara Seas
476 precede a negative AO-like pattern in December, which is consistent with previous
477 results (Fig. 8), but for larger SIC changes. Interestingly, SCE and SIC seem to contribute
478 similarly to the SLP response in Fig. 13. Indeed, projecting SIC anomalies onto the SIC
479 part of the MCA covariance map (referred to as MCAcat_SIC) and SCE anomalies onto the
480 SCE part (referred to as MCAcat_SCE) yields two well correlated time series (0.58,
481 significant at the 5% level) that compare well with the atmospheric December MCA time
482 series (Fig. 13b).

483 In order to evaluate the relative influence of the SCE and SIC pattern, we used the
484 time series associated with the SCE and SIC fields in the SCE/SLP (MCA-snow) and
485 SIC/SLP (referred to as MCA-SIC) individual MCA, respectively, to separate more clearly
486 the SIC and SCE influences. These two times series have a correlation of 0.42, and a
487 bivariate regression of the SLP using these two time series shows little multicollinearity
488 (variance inflation factor of 1.4). The regression slopes (Fig. 14) show that the SCE holds
489 a larger signal in observations, which is consistent with the higher correlation in the
490 MCA analysis (see Table 2). The SIC has a similar influence, but its amplitude is twice
491 smaller, and it is less significant. These results are not modified when using other
492 indices for SCE or SIC.

493 The concatenated MCA yields similar results for the four models, with a SCE
494 dipole and a decrease of SIC in November preceding the December AO (not shown),
495 although the NSC and correlation are much lower, and adding SIC to SCE (or considering
496 SIC alone) strongly degrades the significance (Table 2). Yet, the correlation between the
497 MCAcat_SCE and MCAcat_SIC time series (Table 3) is significant in each model, even if it
498 is lower than in observations, which can be explained by the different sampling, the
499 smaller SCA occurrence, or model biases such as the colder mean state in pre-industrial
500 climate, which allow less Barents-Kara SIC variability. However, these significance tests
501 are biased since the four models were selected based on their response to SCE, not to
502 SIC, and other CMIP5 models are more sensitive to SIC (Garcia-Serrano et al. 2016).

503 The same analysis was conducted using SIC anomalies in early autumn
504 (September or October) together with November SCE (Table S2), which provides
505 significant results only when using October SIC, with patterns as in Fig. 13, but smaller
506 NSC and R. We also repeated the analysis using November SIC/SCE and SLP in January
507 and February (Table S3), as the stratospheric pathway is also important during late
508 winter (Kim et al., 2014; Jaiser et al., 2016), but the MCA results are much less significant
509 in the observations.

510

511 *e. Link with the Scandinavian pattern*

512 The upward influence of tropospheric planetary waves into the stratosphere due
513 to atmospheric dynamics, such as during blocking situations, can also explain that the
514 SCA is followed by an AO-like pattern one month later, without any influence of surface
515 heating (Kuroda and Kodera, 1999; Takaya and Nakamura, 2008; Martius et al., 2009;
516 Woollings et al., 2010). To test the influence of such troposphere-stratosphere coupling,
517 we use an MCA with Eurasian SLP (0E-150E, 45N-85N), Eurasian SCE, and Barents/Kara

518 SIC in November concatenated as the predictor field, and Northern Hemisphere SLP in
519 December as the predictand field. For the sake of simplicity, the ENSO variability was
520 not removed in the analysis. In both observations and models, the results of this MCA
521 are strongly significant (Table 2), and the covariance maps are similar to Fig. 13, with
522 the homogeneous SLP covariance map in November resembling the SCA (not shown).

523 Time series of the three November predictors (SCE dipole, Barents/Kara SIC,
524 SCA) are then studied. The time series associated with the SCE and SIC fields are
525 obtained as before from the SCE/SLP (MCA-snow) and SIC/SLP (MCA-SIC) individual
526 MCAs, while the SCA index is given by the first rotated EOF of the Eurasian SLP (0E-
527 150E, 45N-85N) in November. To distinguish the impacts of each predictor, a
528 multivariate regression of the December SLP on the three predictors is done, noting that,
529 despite the large correlation between predictors, multicollinearity is limited (variance
530 inflation factors < 2.0). The results (Fig. 15a-c) again show that the SCE dipole has the
531 dominant influence onto SLP in December, while the SIC provides weaker, but
532 significant anomalies as in the bivariate regression in Fig. 14. The SCA seems to be also
533 important for the SLP over the British Isles or Alaska, but the anomalies are weaker and
534 not significant. A similar multivariate regression using an AO index, as given by the first
535 EOF of December SLP is shown in Fig. 15d. Again, the SCE appears to be the best
536 predictor of the AO, followed by the SIC, while the SCA has the lowest R^2 . Taking the
537 three indices as predictors with a multivariate regression only slightly improves the
538 variance explained by the SCE alone. In the four models (Fig. 15d, symbols using the
539 right vertical axis), the same analysis also shows that the SCE dipole still plays the
540 dominant role in three models, while the SIC has a dominant influence only in one model
541 (CanESM). In all models, the SCA pattern also appears as good predictor of the AO. This
542 suggests that, in these models selected based on their response to SCE, internal

543 atmospheric dynamical processes may also explain the statistical relationship found
544 among SCE, SIC and the atmosphere one month later, hence that the influence of SCE and
545 SIC is underestimated. These conclusions are not substantially modified when using
546 other indices for the AO, the snow dipole or the Barents-Kara SIC anomalies.

547

548 **5. Discussions and Conclusion**

549 We have investigated the links between Eurasian SCE and the atmosphere in
550 observations during 1979-2014 and CMIP5 models. We found that a dipole of snow
551 cover anomalies in November with positive (negative) snow cover anomalies over
552 eastern Siberia (eastern Europe) leads to a negative AO one month later in December.
553 The largest statistical links are found when considering November SCE, as in Orsolini et
554 al. (2016), but other studies focus more on October snow cover (Cohen and Entekhabi,
555 1999; Cohen et al. 2007; Cohen and Jones 2011; Handorf et al. 2015). Lagged regression
556 actually reveals that the November SCE is related to similar anomalies in October, but
557 statistical significance is too limited with the MCA using October SCE. The choice of the
558 data set, the methodology and the period considered might explain this discrepancy (see
559 Appendix A). The CMIP5 models, in general, fail to simulate this potential effect of snow
560 cover. Nevertheless, a weaker, but similar, relationship between the SCE and the AO is
561 present in four models: CanESM, MPI-ESM-LR, GISS-E-R and CESM-BGC.

562 The models and ERA-Interim indicate that downward (upward) heat flux
563 anomalies are simulated over positive (negative) snow cover anomalies over Siberia
564 (Europe) during November. We verified that eastern Siberia pole of the snow dipole
565 anomalies has the best relationship with the AO one month later both in observations
566 and models, so that the SCE over Siberia seems to have the largest influence. The
567 diabatic cooling of the troposphere over Siberia is consistent with the intensification and

568 westward expansion of the Siberian High. This may lead to a polar vortex weakening
569 from November to January driven by upward planetary wave activity flux, as found
570 previously in observations (Saito et al. 2001; Handorf et al. 2015; Furtado et al. 2016)
571 and in sensitivity experiments using SCE anomalies (Gong et al., 2004; Fletcher et al.,
572 2009; Peings et al., 2012; Orsolini et al. 2013; Orsolini et al. 2016). Here, we show that
573 the same process can be verified qualitatively using multi-centennial control climate
574 model simulations, although the SCE influence is much weaker.

575 The atmospheric pattern responsible for the variability of the snow cover dipole
576 is the Scandinavian pattern (SCA, as in Bueh and Nakamura, 2007), with a large
577 anticyclone over the Urals. Such anticyclone leads to northerly cold advection east of the
578 anticyclone, bringing cold air over Siberia, and southerly warm advection over Central
579 Europe and the Barents and Kara Seas. The SCA forcing explains that the Barents/Kara
580 SIC and Eurasian snow cover are largely correlated (Wegmann et al., 2015; Furtado et
581 al., 2016). We find that the models produce less frequent SCA-like and more frequent
582 AO-like events, possibly linked to blocking processes that are not well simulated in low
583 resolution models (Dawson, 2012), but this could also be due to natural atmospheric
584 variability. Deficiencies in the simulation of the SCA characteristics in models might
585 therefore explain the weaker SCE influence in models. In addition, the upward heat flux
586 driven by a retreat of the sea ice cover in the Barents-Kara Seas is weaker and less
587 robust in the models than in ERA-Interim, perhaps explaining why the SIC influence is
588 also underestimated in the four models that simulate the SCE impacts.

589 A MCA using SLP and combined SCE and SIC suggests that November SCE and SIC
590 forcing provide similar covariability with the December AO in observations. However, a
591 bivariate regression reveals that the SCE dipole is a much better predictor than the
592 Barents-Kara SIC anomaly. As the SCE and the SIC variability are linked, both field can

593 constructively interfere to weaken the polar vortex, as suggested in Cohen et al. (2014a),
594 although the surface forcing from the snow cover anomalies might be dominant. On the
595 other hand, the November SIC in models has an impact on the AO in only one model,
596 perhaps because they were selected based on their representation of the SCE influence.
597 When investigating more systematically the links between Greenland-Barents-Kara SIC
598 and the NAO/AO in most CMIP5 models, Garcia-Serrano et al. (2016) did find a robust
599 SIC influence, but they noted that the timing or the processes for the SIC influence are
600 model dependent. Here, the lack of links between November SIC and December
601 atmosphere may result from our selection of the models based on their representation
602 of the SCE impact (and not SIC impact), and also from the model averaging that may mix
603 different behavior among models. The weaker SCE influence in models and the lack of
604 link between the SCE and SIC is consistent with the underestimated troposphere-
605 stratosphere coupling in models, as found in Furtado et al. (2015). However, it can also
606 be explained by the poor simulation of the SCA, the colder climate in preindustrial
607 control simulation, or natural climate variability.

608 A better understanding of the coupling between land snow cover, Arctic sea ice,
609 and the atmosphere using dedicated climate model experiments would be necessary to
610 properly assess the causality links and better discriminate between their influence on
611 the winter AO. Nonetheless, the methodology used here could be applied to climate
612 projection of the 21st century in order to investigate how the polar amplification of
613 global warming will modify the links between the atmosphere and Arctic surface
614 conditions.

615

616

617

618 **Appendix : October snow cover influence**

619

620 The influence of October SCE on the atmosphere is discussed by using the MCA
621 results, when SLP lags by one month, although statistical significance is limited (see Fig.
622 1). The covariance maps (Fig. A1a) show that increasing October SCE over northern
623 Eurasia precedes a SLP pattern in November that has some resemblance with the SCA,
624 plus a deeper Aleutian low. This differs from the negative AO found later, from
625 December to February. It might be due the snow data used, as many previous studies
626 used a more integrated snow index, such as the Eurasian snow cover areal extent (e.g.
627 Cohen et al. 2007; Cohen and Fletcher 2007). It could be due to differences in
628 methodology, as Furtado et al. (2016) used multivariate EOF. It could also be due to non-
629 stationarity (Peings et al., 2013). For instance, Cohen et al. (2007) considered the 1948-
630 2004 period, Cohen and Fletcher (2007) the 1972-2005 one, while we focus on 1979-
631 2014.

632 To investigate the possible influence of non-stationarity, we performed the MCA
633 in different sub-periods (Table A1). The most significant influence of October snow
634 cover on SLP is found for November in the 1979-2005 period, as used in Cohen and
635 Fletcher (2007); the MCA mode is also significant for December SLP, with a MCA pattern
636 (Fig. A1d) sharing a large similarity with previous studies (i.e. Handorf et al., 2015).
637 However, the levels of significance are limited when the DJF atmosphere is considered. If
638 1979-2011 or 1979-2014 is used, significance is lost. Hence, the detected influence of
639 the October snow cover is sensitive to the period.

640

641

642 *Acknowledgments.*

643 The research leading to these results has received funding from the European Union 7th
644 Framework Programme under grant agreement #308299 (NAACLIM). Javier García-
645 Serrano also received funding from H2020-funded DPETNA grant (MSCA-IF-EF
646 655339). We thank the European Center for Medium Range Weather Forecast for
647 providing the ERA-Interim reanalysis, the Rutgers University for providing the observed
648 snow cover fields, the National Snow and Ice Data Center for providing the sea ice
649 extent. For their role in producing, coordinating, and making available CMIP5 model
650 outputs, we acknowledge the climate modeling groups, the World Climate Research
651 Programme's Working Group on Coupled Modelling and the Global Organization for
652 Earth System Science Portals. This study benefited from the IPSL mesocenter facility
653 which is supported by CNRS, UPMC, Labex L-IPSL, which is funded by the ANR (Grant
654 #ANR-10-LABX-0018), and by the European FP7 IS-ENES2 project (Grant #312979).
655

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854 Tibetan Plateau: the associated circulation and influence on the east Asian summer
855 monsoon. *J. Clim.*, **17**, 2780–2793.

856

857 **Tables**

858

859 TABLE 1. CMIP5 models and control simulations used.

860

	Group	Model	AGCM Resolution	length (year)
1	CCCma	CanESM2	2.8°x2.8° L35	995
2	CNRM-CERFACS	CNRM-CM5	1.4°x1.4° L31	850
3	CSIRO-QCCCE	CSIRO-Mk3-6-0	1.9°x1.9° L18	500
4	LASG-CESS	FGOALS-g2	2.8°x2.8° L26	700
5	MIROC	MIROC-ESM	1.4°x1.4° L40	630
6	MPI-M	MPI-ESM-LR	1.9°x1.9° L47	1000
7	MRI	MRI-CGCM3	1.1°x1.1° L48	500
8	NASA-GISS	GISS-E2-R	2.5°x2° L40	550
9	NCAR	CCSM4	1.25°x0.9° L26	600
10	NCC	NorESM1-ME	2.5°x1.9° L26	250
11	NSF-DOE-NCAR	CESM1-BGC	1.25°x0.9° L26	500
12	IPSL	IPSL-CM5A-LR	1.9°x3.75° L39	1000

861

862

863

864 TABLE 2. Statistics of different MCAs using December SLP as the left field, and November
 865 snow cover (SCE), SIC, concatenated SCE and SIC (SCE+SIC) or concatenated SCE, SIC
 866 and Eurasian SLP (SCE+SIC+SLP_{Eur}) as the right field. For the models, the mean over the
 867 four selected models is given. The level of significance is given in parenthesis for
 868 observation (see section 2c for details). For climate models, the number in parenthesis
 869 indicates the number of models, out of four, where the level of significance is equal or
 870 below 10%.

871

	OBS		Models	
	NSC	R	NSC	R
SCE	2.5 (0%)	0.82 (1%)	0.10 (4/4)	0.23 (4/4)
SIC	2.9 (3%)	0.61 (18%)	0.14 (1/4)	0.14 (1/4)
SCE+SIC	2.4 (0%)	0.75 (2%)	0.10 (2/4)	0.16 (0/4)
SCE+SIC+SLP _{Eur}	2.1 (0%)	0.78 (0%)	0.14 (4/4)	0.24 (4/4)

872

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876

877 TABLE 3. Correlation between MCAcat-SCE and MCAcat-SIC time series. The bold
 878 numbers indicate 1% significance.

879

Data	Correlation
Observations	0.58
CanESM2	0.26
GISS-E2-R	0.24
MPI-ESM-LR	0.40
CESM1-BGC	0.27

880

881

882 TABLE A1. Statistics of the MCA using October snow cover and SLP in following months,
 883 using different time periods (79-05 : from 1979 to 2005 ; 79-11 : from 1979 to 2011 and
 884 79-14 : from 1979 to 2014), and atmospheric months (NOV : November ; DEC :
 885 December ; DJF : December-January-February). The level of statistical significance is
 886 given in parenthesis.

887

Period	SLP season	NSC	R
79-14	NOV	1.3 (10%)	0.70 (13%)
79-14	DJF	1.1 (29%)	0.63 (32%)
79-05	NOV	1.9 (3%)	0.83 (5%)
79-05	DEC	1.9 (6%)	0.80 (6%)
79-05	DJF	2.4 (6%)	0.71 (25%)
79-11	NOV	1.1 (27%)	0.77 (21%)
79-11	DEC	1.6 (9%)	0.71 (27%)
79-11	DJF	1.5 (11%)	0.66 (44%)

888

889

890

891 **Figures Caption**

892 **Figure 1 :**

893 Normalized squared covariance (NSC, contours, in %) for the first MCA mode between
894 observed SLP and Eurasia snow cover, for each month in the atmosphere. The lag is
895 positive when the snow cover leads SLP. The gray shade provides the level of statistical
896 significance for NSC. The plus symbols indicate the atmospheric month and time lag
897 where the level of significance for the correlation (R) is below 5%.

898

899 **Figure 2 :**

900 (a) Scatter plot of the confidence level, in %, of the normalized squared covariance, NSC,
901 versus that of the correlation, R, for the first MCA mode between SLP and Eurasian snow
902 cover. (b) Scatter plot of the spatial correlation between the SLP covariance map found
903 in models and that of ERA-Interim, versus the spatial correlation between the snow
904 cover covariance map found in models and that of ERA-Interim. The black indicates the
905 results for SLP in December and SCE in November (one month lag). The blue indicates
906 the results for SLP in November and SCE in October (one month lag). The red indicates
907 the results for the SLP in January and SCE in November (two month lag). In (b), the bold
908 symbols indicate levels of significance lower than 15% for both NSC and R.

909

910 **Figure 3 :**

911 (a) Homogeneous snow cover fraction (in %) and heterogeneous SLP (in hPa)
912 covariance maps for the first MCA mode, for December SLP and November snow cover,
913 when the snow cover leads by one month the atmosphere, in ERA-Interim. (b) Same as

914 (a), but using January SLP with a 2 month lag. (c), (d), (e), (f) and (g) same as (a) but for
915 CanESM2, MPI-ESM, GISS-E2-R, CESM1-BGC and the mean of the four models,
916 respectively. Note that the color scale is different for observation and models. (h) Box
917 plots of the NSC and R statistics from the MCA using 36-yr periods extracted from the
918 control runs of each models (1: CanESM2, 2:MPI-ESM, 3: GISS-E2-T and 4: CESM1-BGC),
919 error bars show the 5% and 95% percentiles. The dashed horizontal lines show the NSC
920 and R values in observations.

921

922 **Figure 4 :**

923 Regression of SLP (contours, in hPa) and 2m air temperature, (color, in K) on the MCA-
924 snow index, in November, for (a) ERA-Interim and (b) the subset of four models. In (a),
925 colors are masked if the level of significance is above 10% for observation. In (b), colors
926 indicate anomalies of the same sign among the four models.

927

928 **Figure 5 :**

929 Homogeneous SLP (in hPa) and heterogeneous snow cover (in %) covariance maps for
930 the first MCA mode, when the SLP and snow cover are simultaneous (no lag), for (a)
931 November fields in ERA-Interim; (b) December fields in ERA-Interim and (c) November
932 fields in the mean of the four models subset.

933

934 **Figure 6 :**

935 (a) REOF1 of November SLP (in hPa) in ERA-Interim. (b) Same as (a) for the model mean
936 REOF1 using the four models. In (a), the variance fraction is given in parenthesis. In (b),
937 the minimum and maximum variance fraction among the four models is indicated in
938 parenthesis. (c) Box plots of the November variance (in %) explained by the SCA and the

939 NAO/AO in 36-yr chunks from the control runs of each models (1: CanESM2, 2:MPI-ESM,
940 3: GISS-E2-R and 4: CESM1-BGC); the error bars give the 5% and 95% percentiles, and
941 the dashed horizontal lines the AO and SCA variance fraction in observations.

942

943 **Figure 7 :**

944 Regression of the December SLP in hPa onto (Left) European and (Right) Siberian snow
945 anomalies, given by multivariate regression; for (upper) ERA-Interim and (lower) the
946 subset of four models. In (a) and (b), colors are masked if the level of statistical
947 significance is above 10%. In (c) and (d), colors indicate anomalies of the same sign
948 among the four models.

949

950 **Figure 8 :**

951 Regression of the snow cover fraction (color shades over continent, in %) and sea ice
952 concentration (blue contours and shades over the ocean, in %), onto the November
953 MCA-snow index, for (a) ERA-Interim in October; (b) the four models in October; (c) and
954 (d) Same as (a) and (b) for November; (e) and (f) same as (a) and (b) for December. The
955 sea-ice concentration contour interval is 5% in observations, and 1% for models, the
956 zero contour is removed. The thick gray contour provides the 50% contour for
957 climatological SIC.

958

959 **Figure 9 :**

960 Thermodynamical component of the heat flux, positive upward, in $W m^{-2}$, associated in
961 November with the November MCA-snow index in (a) ERA-Interim and (b) the four
962 models. The color scale is different over land and ocean to emphasize the changes over
963 continental surfaces. Note the different contour intervals for ERA-Interim and models.

964 (c,d) Regressions of the shortwave (SW), longwave (LW), sensible (SH), latent (LH) and
965 total (Tot) heat flux over the Siberia (SIB), Europe (EUR) and Barents-Kara Sea (B/K)
966 integrated over boxes shown in (a) and (b) with histograms for (c) ERA-Interim and (d)
967 the four models mean. In (d) the error bars indicate the minimum and maximum values
968 among models.

969

970 **Figure 10 :**

971 Regression of the SLP, in hPa (contour interval 0.5 hPa), onto the MCA-snow index, (left
972 column) ERA-Interim and (right column) models, in (a), (b) October; (c), (d) November;
973 (e), (f) December and (g), (h) January . The thick black line indicates 5% significance for
974 observations or anomalies of the same sign among the four models. The contour interval
975 at -0.2 and 0.2 hPa was added for models.

976

977 **Figure 11 :**

978 Regression of the zonal-mean temperature (gray contours and color shades, in K) and
979 geopotential height (blue contours, in m) onto the MCA-snow normalized index, for (left
980 column) ERA-Interim and (right column) models, in (a), (b) November; (c), (d)
981 December and (e), (f) January. Colors indicate zonal mean temperature (left) level of
982 significance below 10% or (right) anomalies of the same sign among the four models.

983

984 **Figure 12 :**

985 Regression of the temperature over the polar cap (65°N-90°N) onto the MCA-snow
986 normalized index, for (a) ERA-Interim and (b) models. The thick black lines indicate (a)
987 level of significance below 10% or (b) anomalies of the same sign among the four
988 models. Note the different contour intervals.

989

990 **Figure 13 :**

991 (a) Snow cover (color over land, in %) and SIC (color over ocean, in %) homogeneous
992 covariance map and SLP (in hPa) heterogeneous map for the first MCA mode using
993 combined snow/sea-ice in November and SLP in December for ERA-Interim. (b) (black)
994 MCAcat_SCE, (red) MCAcat_SIC and (green) atmospheric SLP yearly time series from the
995 MCA (normalized).

996

997 **Figure 14 :**

998 Regression slopes of a bivariate regression of the SLP (in hPa) for the (a) MCA-snow, and
999 (b) MCA-SIC indices. Colors indicate level of significance below 10%.

1000

1001 **Figure 15 :**

1002 Regression slopes of a multivariate regression of the SLP (in hPa) onto the (a) snow
1003 dipole, (b) Barents-Kara Sea SIC and (c) SCA indices. In (a-c) colors indicate level of
1004 significance below 10%. (d) R^2 value of univariate regressions using the AO index as
1005 predictand and snow dipole, Barents-Kara Sea SIC or SCA as predictor. ALL indicates the
1006 R^2 when using the three indices in a multivariate regression. Note that the y-axis is
1007 different for observation (bars, left axis) and models (symbols, right axis).
1008 The black symbols (bars) provide the results for models (observations), thick symbols
1009 (bars) indicating level of significance of R^2 below 10%.

1010

1011 **Figure A1 :**

1012 (a) Homogeneous October snow cover fraction (in %) and November heterogeneous SLP
1013 (in hPa) covariance maps for the first MCA mode, when the snow cover leads by one

1014 month the atmosphere, for ERA-Interim during 1979-2014. (b) Same as (a) but for the
1015 1979-2005 period. (c) Same as (a) but using the December SLP. (d) Same as (c) but for
1016 the 1979-2005 period.

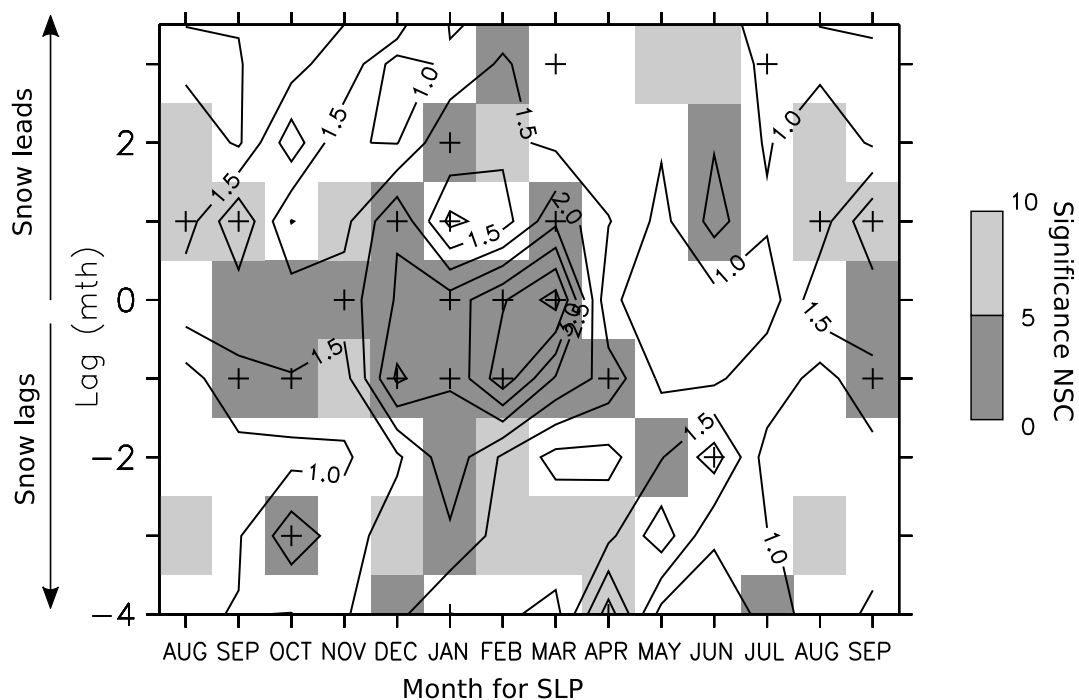


Fig. 1 : Normalized squared covariance (NSC, contours, in %) for the first MCA mode between observed SLP and Eurasia snow cover, for each month in the atmosphere. The lag is positive when the snow cover leads SLP. The gray shade provides the level of statistical significance for NSC. The plus symbols indicate the atmospheric month and time lag where the level of significance for the correlation (R) is below 5%.

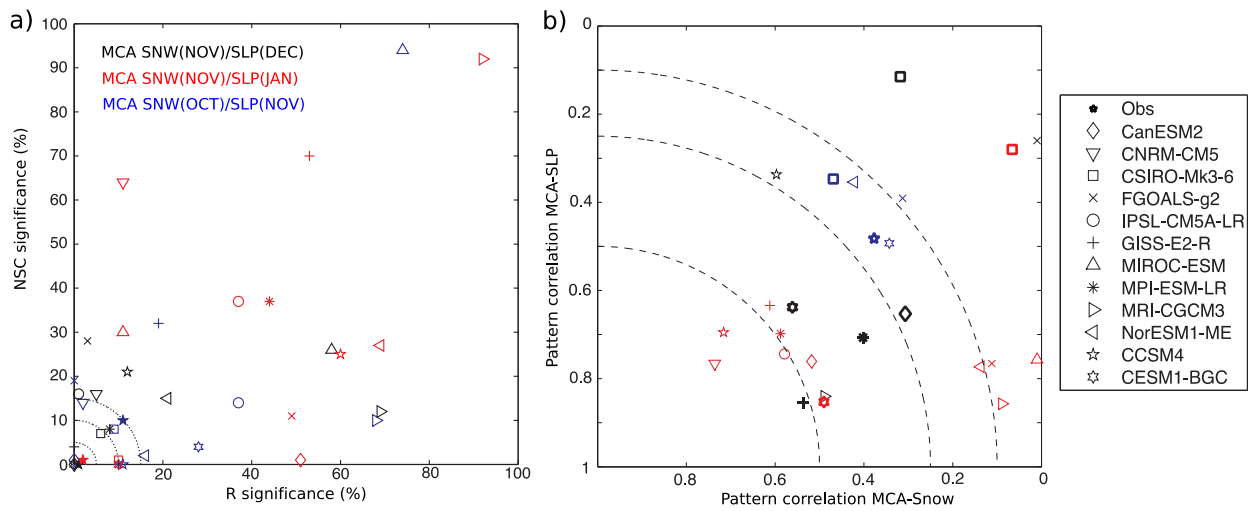


Fig. 2 : (a) Scatter plot of the confidence level, in %, of the normalized squared covariance, NSC, versus that of the correlation, R, for the first MCA mode between SLP and Eurasian snow cover. (b) Scatter plot of the spatial correlation between the SLP covariance map found in models and that of ERA-Interim, versus the spatial correlation between the snow cover covariance map found in models and that of ERA-Interim. The black indicates the results for SLP in December and SCE in November (one month lag). The blue indicates the results for SLP in November and SCE in October (one month lag). The red indicates the results for the SLP in January and SCE in November (two month lag). In (b), the bold symbols indicate levels of significance lower than 15% for both NSC and R.

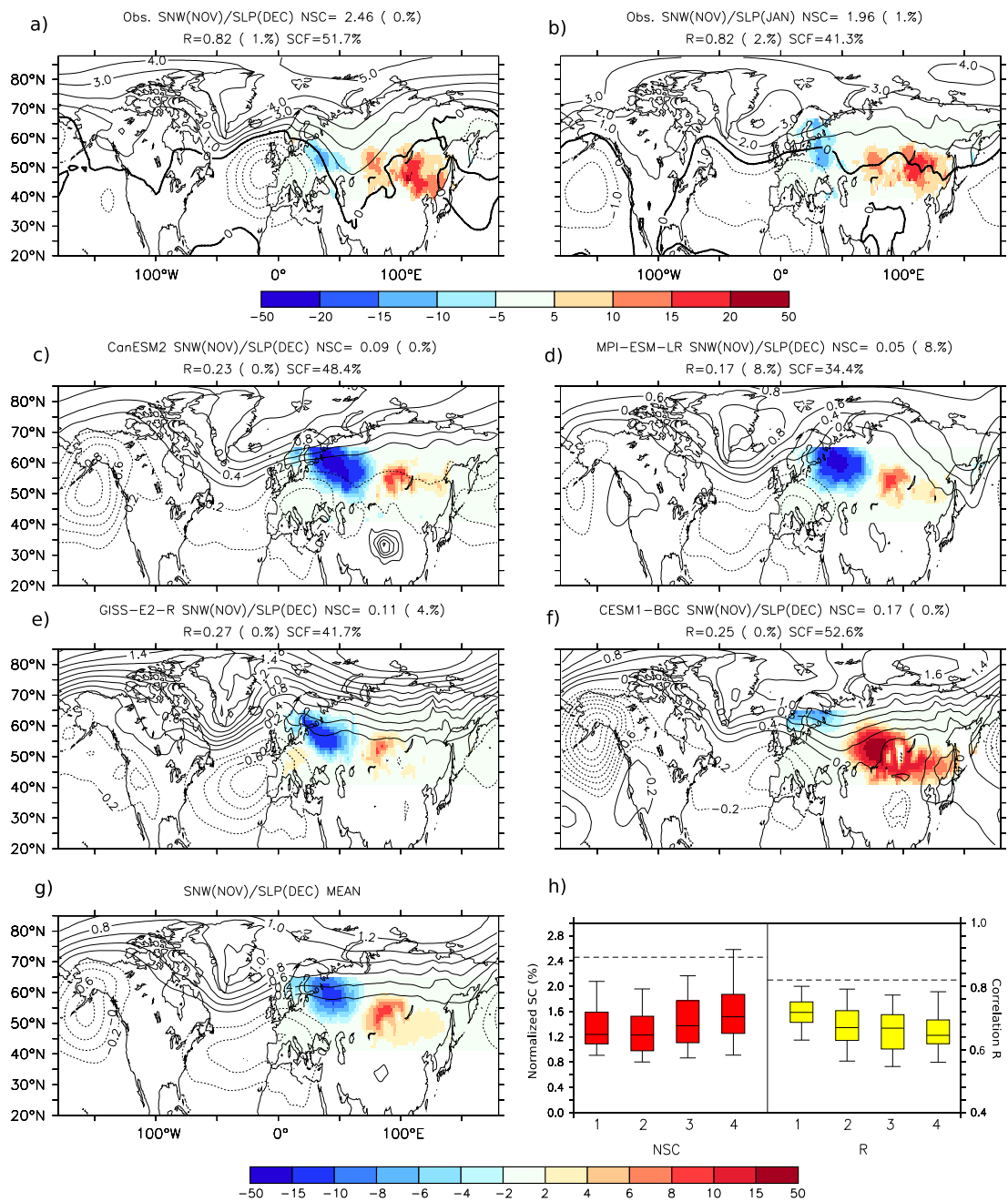


Fig. 3 : (a) Homogeneous snow cover fraction (in %) and heterogeneous SLP (in hPa) covariance maps for the first MCA mode, for December SLP and November snow cover, when the snow cover leads by one month the atmosphere, in ERA-Interim. (b) Same as (a), but using January SLP with a 2 month lag. (c), (d), (e), (f) and (g) same as (a) but for CanESM2, MPI-ESM, GISS-E2-R, CESM1-BGC and the mean of the four models,

respectively. Note that the color scale is different for observation and models. (h) Box plots of the NSC and R statistics from the MCA using 36-yr periods extracted from the control runs of each models (1: CanESM2, 2:MPI-ESM, 3: GISS-E2-T and 4: CESM1-BGC), error bars show the 5% and 95% percentiles. The dashed horizontal lines show the NSC and R values in observations.

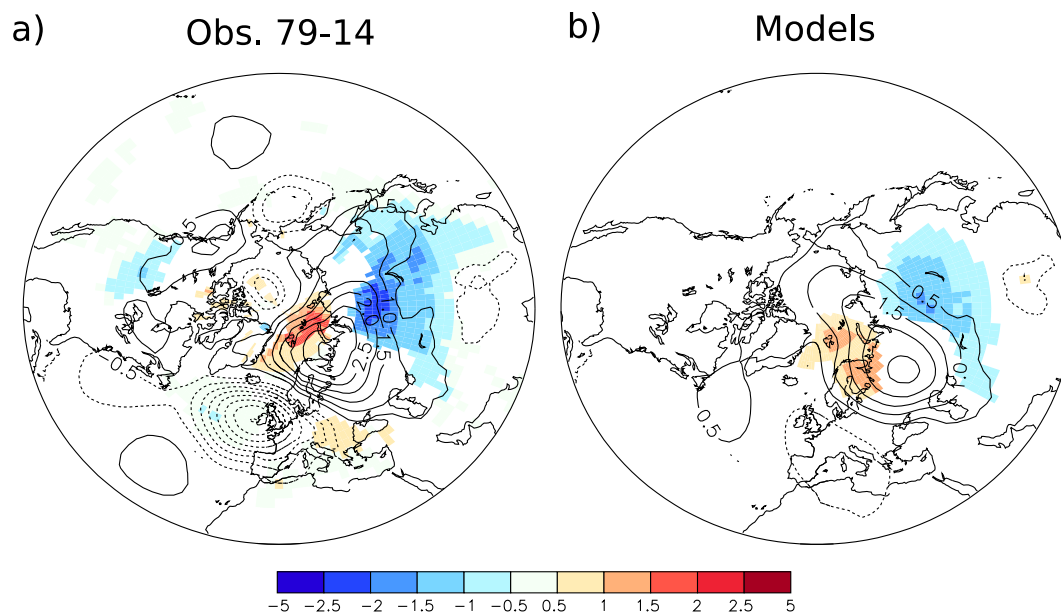


Fig. 4 : Regression of SLP (contours, in hPa) and 2m air temperature, (color, in K) on the MCA-snow index, in November, for (a) ERA-Interim and (b) the subset of four models. In (a), colors are masked if the level of significance is above 10% for observation. In (b), colors indicate anomalies of the same sign among the four models.

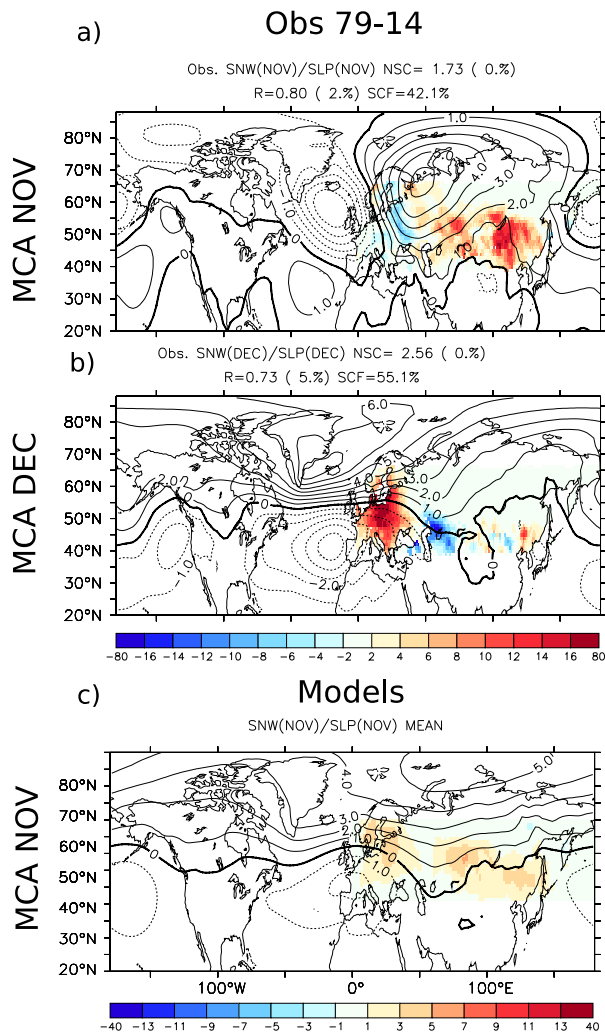


Fig. 5 : Homogeneous SLP (in hPa) and heterogeneous snow cover (in %) covariance maps for the first MCA mode, when the SLP and snow cover are simultaneous (no lag), for (a) November fields in ERA-Interim; (b) December fields in ERA-Interim and (c) November fields in the mean of the four models subset.

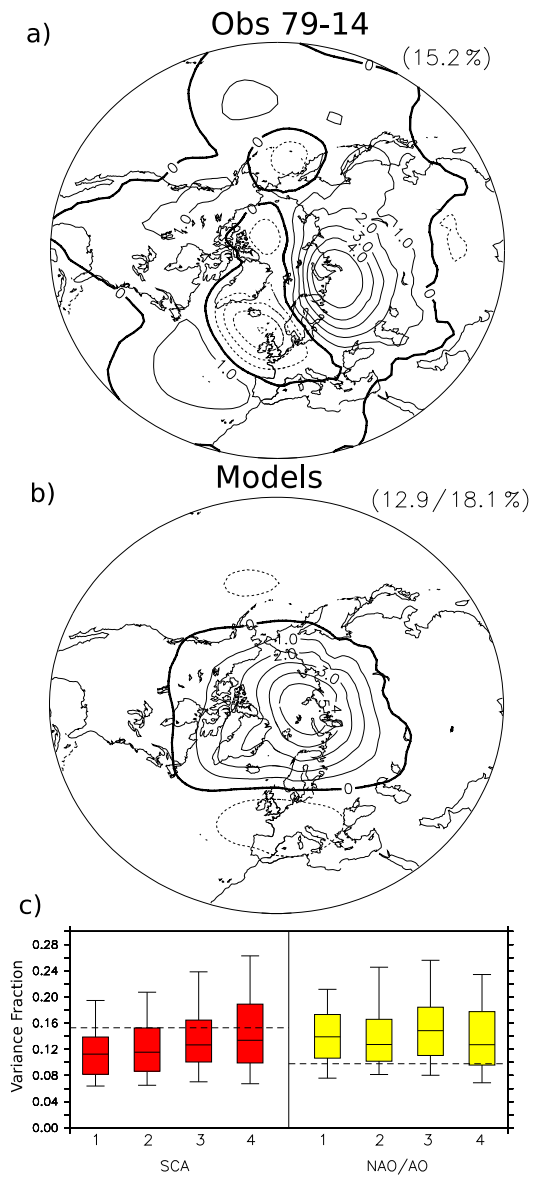


Fig. 6 : (a) REOF1 of November SLP (in hPa) in ERA-Interim. (b) Same as (a) for the model mean REOF1 using the four models. In (a), the variance fraction is given in parenthesis. In (b), the minimum and maximum variance fraction among the four models is indicated in parenthesis. (c) Box plots of the November variance (in %) explained by the SCA and the NAO/AO in 36-yr chunks from the control runs of each models (1: CanESM2, 2:MPI-ESM, 3: GISS-E2-R and 4: CESM1-BGC); the error bars give the 5% and 95% percentiles, and the dashed horizontal lines the AO and SCA variance fraction in observations.

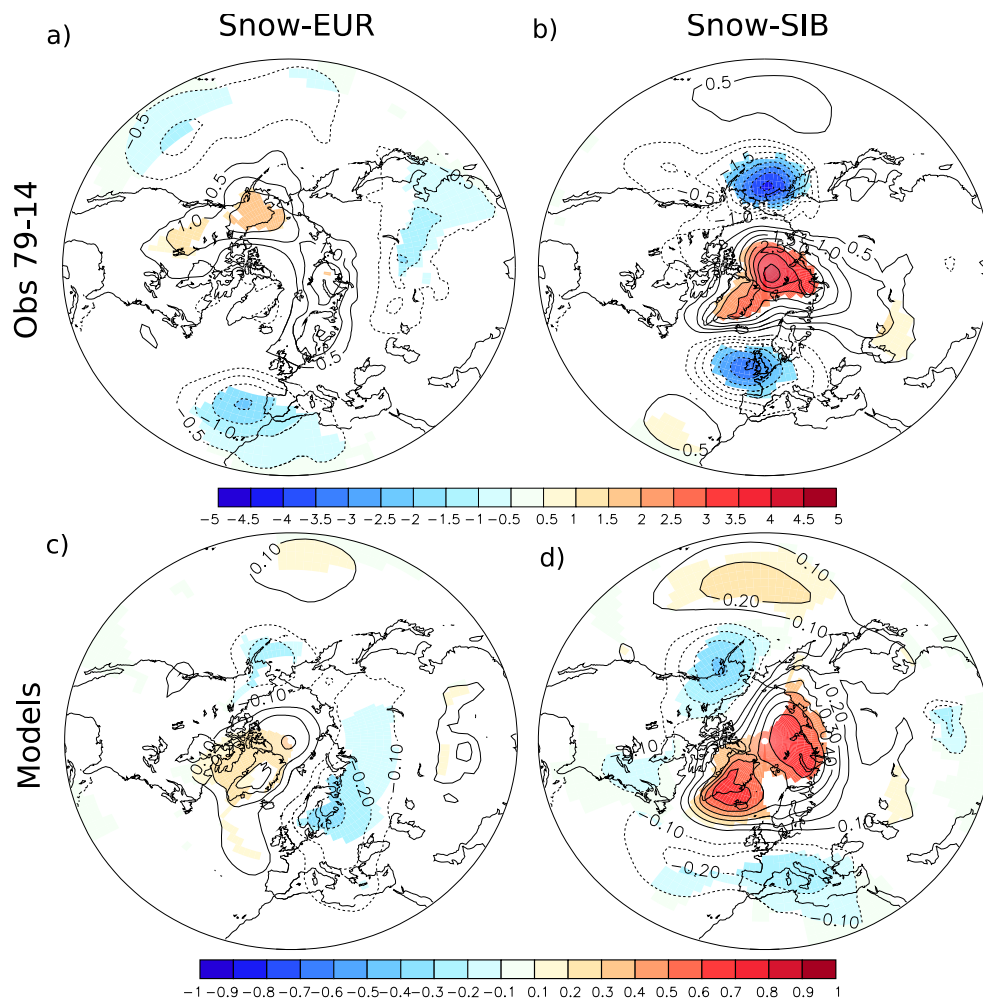


Fig. 7: Regression of the December SLP in hPa onto (Left) European and (Right) Siberian snow anomalies, given by multivariate regression; for (upper) ERA-Interim and (lower) the subset of four models. In (a) and (b), colors are masked if the level of statistical significance is above 10%. In (c) and (d), colors indicate anomalies of the same sign among the four models.

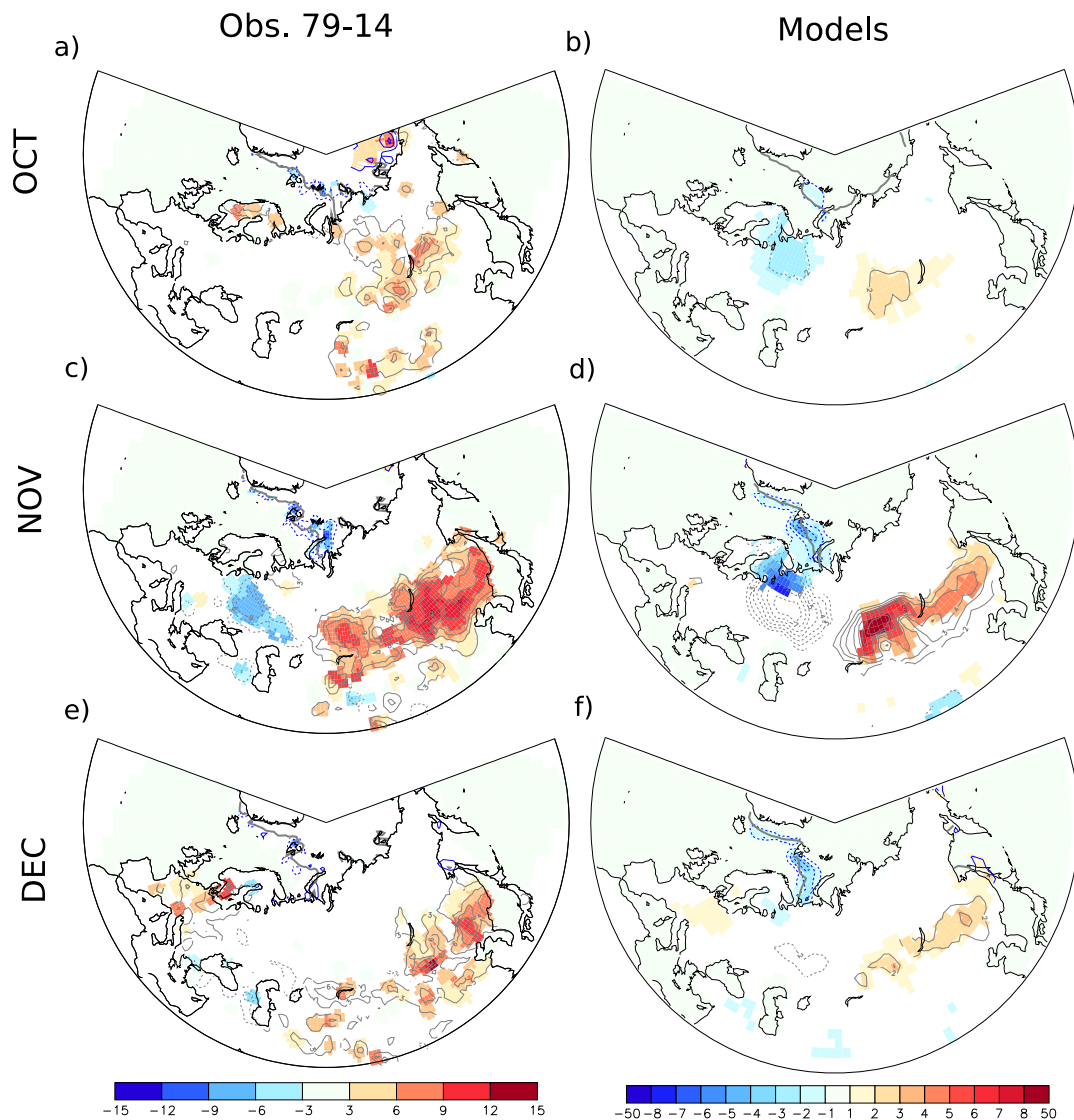


Fig. 8 : Regression of the snow cover fraction (color shades over continent, in %) and sea ice concentration (blue contours and shades over the ocean, in %), onto the November MCA-snow index, for (a) ERA-Interim in October; (b) the four models in October; (c) and (d) Same as (a) and (b) for November; (e) and (f) same as (a) and (b) for December. The sea-ice concentration contour interval is 5% in observations, and 1% for models, the zero contour is removed. The thick gray contour provides the 50% contour for climatological SIC.

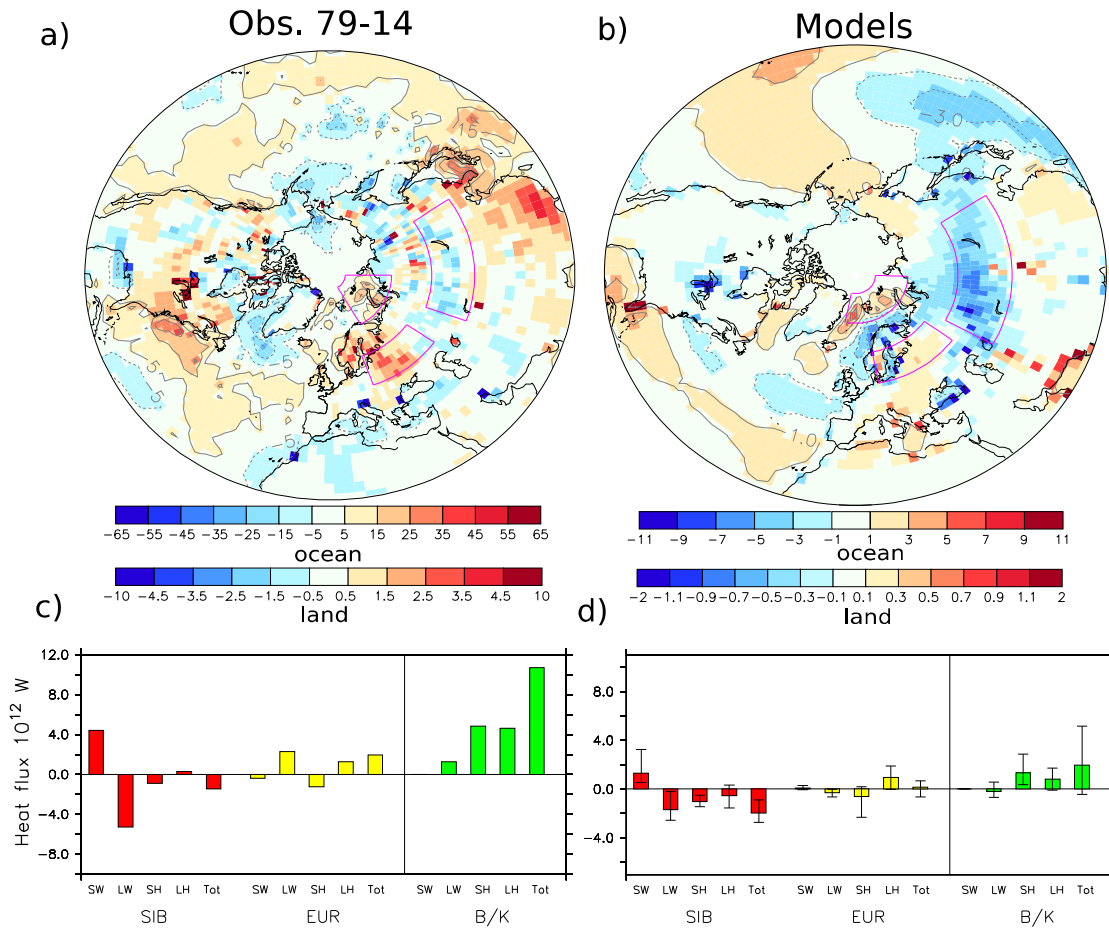


Fig. 9 : Thermodynamical component of the heat flux, positive upward, in $W m^{-2}$, associated in November with the November MCA-snow index in (a) ERA-Interim and (b) the four models. The color scale is different over land and ocean to emphasize the changes over continental surfaces. Note the different contour intervals for ERA-Interim and models. (c,d) Regressions of the shortwave (SW), longwave (LW), sensible (SH), latent (LH) and total (Tot) heat flux over the Siberia (SIB), Europe (EUR) and Barents-Kara Sea (B/K) integrated over boxes shown in (a) and (b) with histograms for (c) ERA-Interim and (d) the four models mean. In (d) the error bars indicate the minimum and maximum values among models.

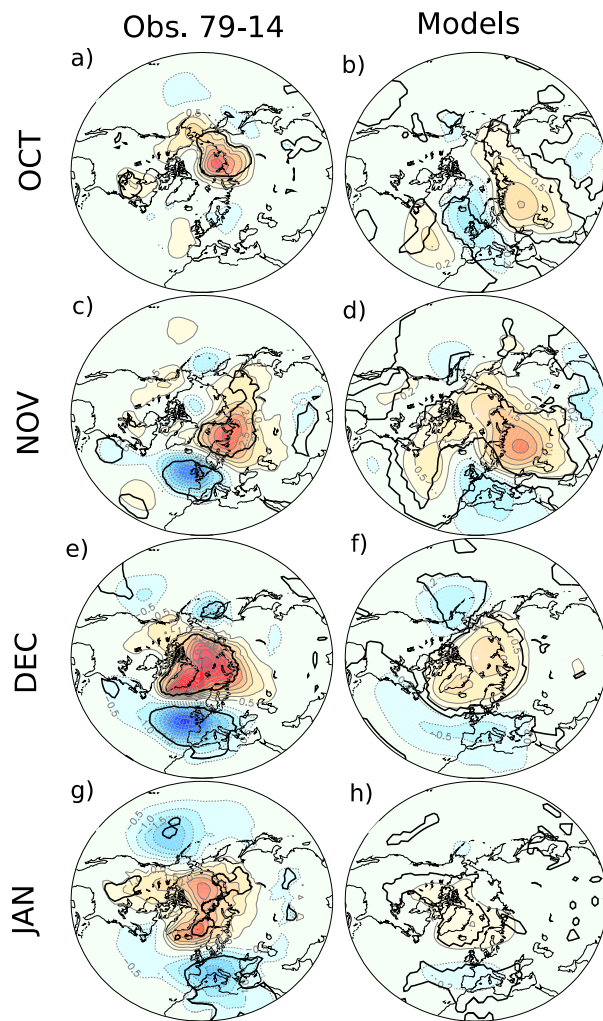


Fig. 10 : Regression of the SLP, in hPa (contour interval 0.5 hPa), onto the MCA-snow index, (left column) ERA-Interim and (right column) models, in (a), (b) October; (c), (d) November; (e), (f) December and (g), (h) January . The thick black line indicates 5% significance for observations or anomalies of the same sign among the four models. The contour interval at -0.2 and 0.2 hPa was added for models.

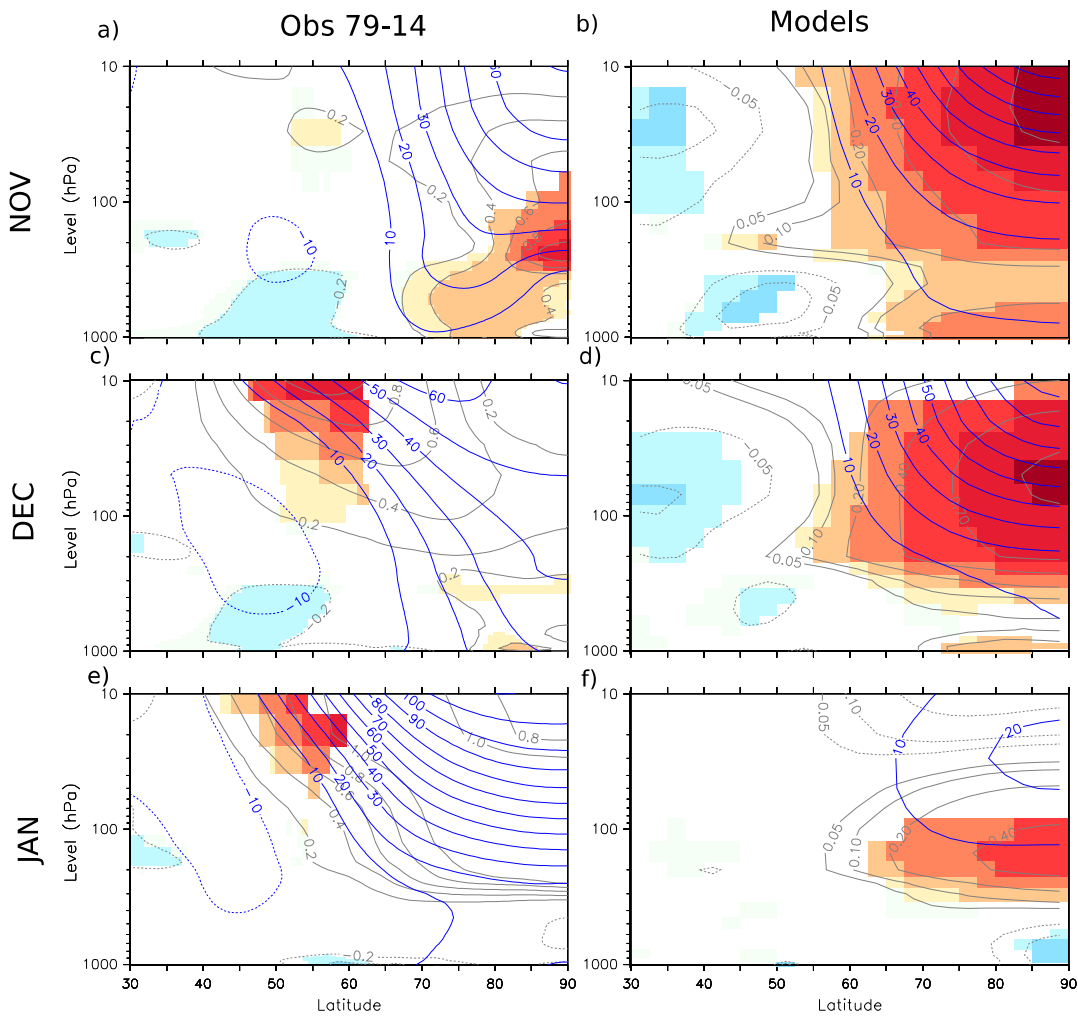


Fig. 11 : Regression of the zonal-mean temperature (gray contours and color shades, in K) and geopotential height (blue contours, in m) onto the MCA-snow normalized index, for (left column) ERA-Interim and (right column) models, in (a), (b) November; (c), (d) December and (e), (f) January. Colors indicate zonal mean temperature (left) level of significance below 10% or (right) anomalies of the same sign among the four models.

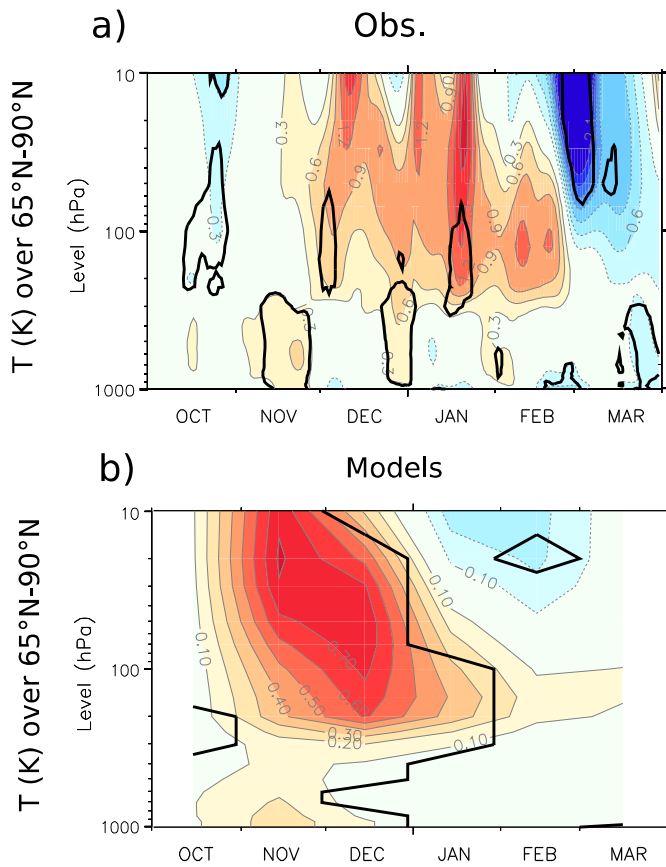


Fig. 12 : Regression of the temperature over the polar cap (65°N-90°N) onto the MCA-snow normalized index, for (a) ERA-Interim and (b) models. The thick black lines indicate (a) level of significance below 10% or (b) anomalies of the same sign among the four models. Note the different contour intervals.

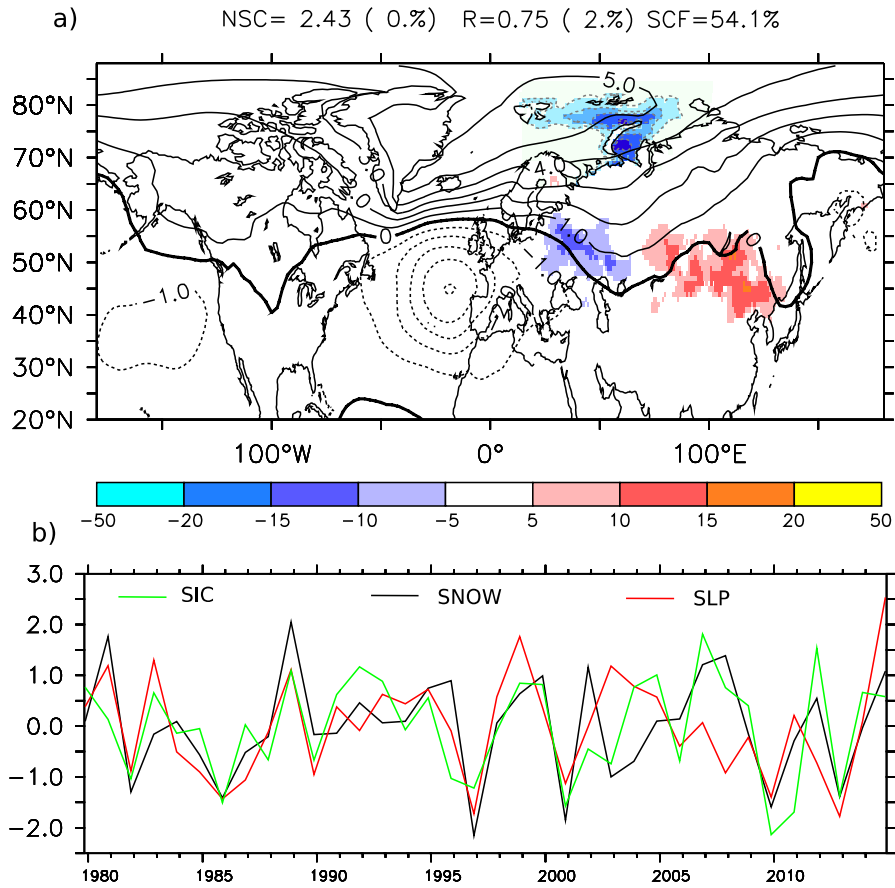


Fig. 13 : (a) Snow cover (color over land, in %) and SIC (color over ocean, in %) homogeneous covariance map and SLP (in hPa) heterogeneous map for the first MCA mode using combined snow/sea-ice in November and SLP in December for ERA-Interim. (b) (black) MCAcat_SCE, (red) MCAcat_SIC and (green) atmospheric SLP yearly time series from the MCA (normalized).

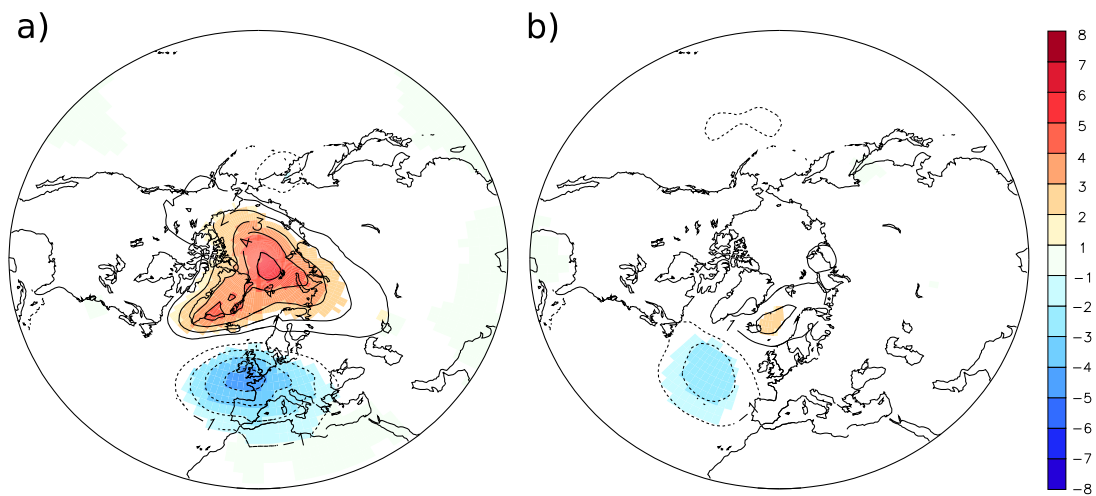


Fig. 14: Regression slopes of a bivariate regression of the SLP (in hPa) for the (a) MCA-snow, and (b) MCA-SIC indices. Colors indicate level of significance below 10%.

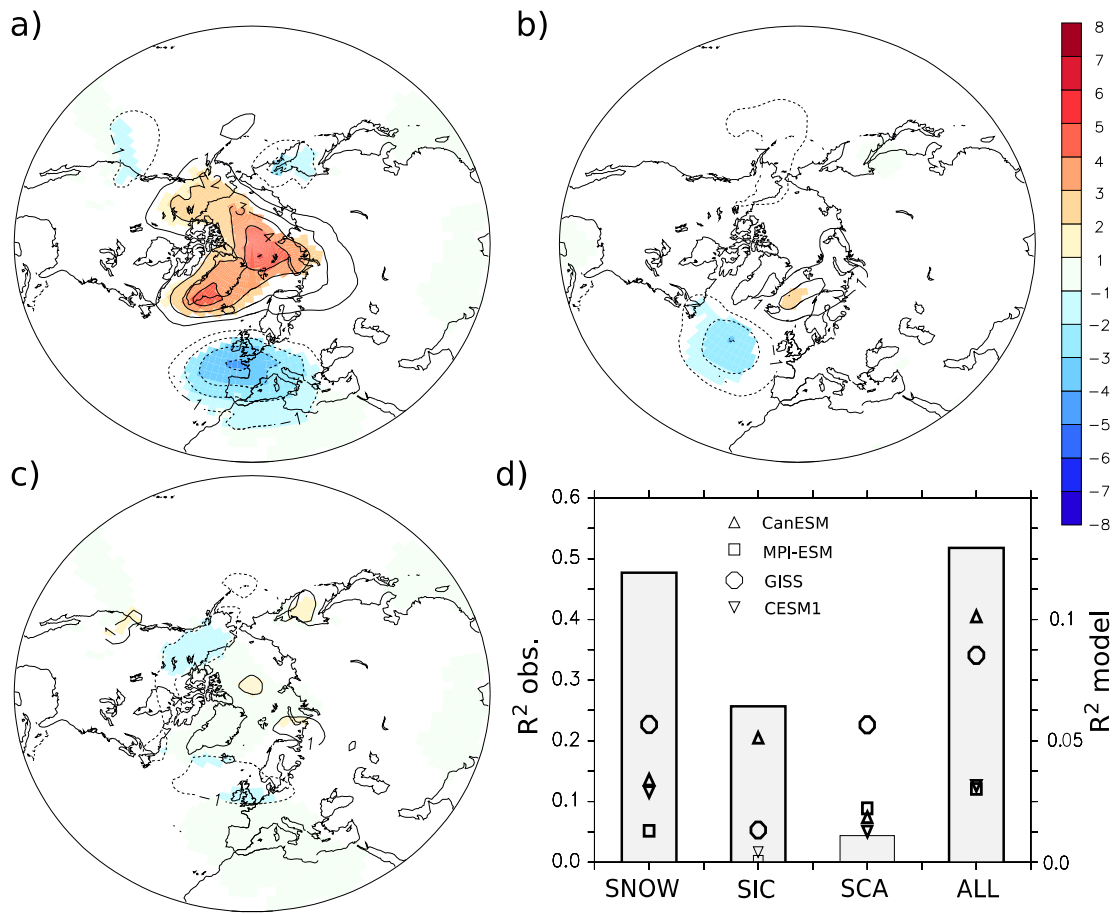


Fig. 15 : Regression slopes of a multivariate regression of the SLP (in hPa) onto the (a) snow dipole, (b) Barents-Kara Sea SIC and (c) SCA indices. In (a-c) colors indicate level of significance below 10%. (d) R² value of univariate regressions using the AO index as predictand and snow dipole, Barents-Kara Sea SIC or SCA as predictor. ALL indicates the R² when using the three indices in a multivariate regression. Note that the y-axis is different for observation (bars, left axis) and models (symbols, right axis). The black symbols (bars) provide the results for models (observations), thick symbols (bars) indicating level of significance of R² below 10%.

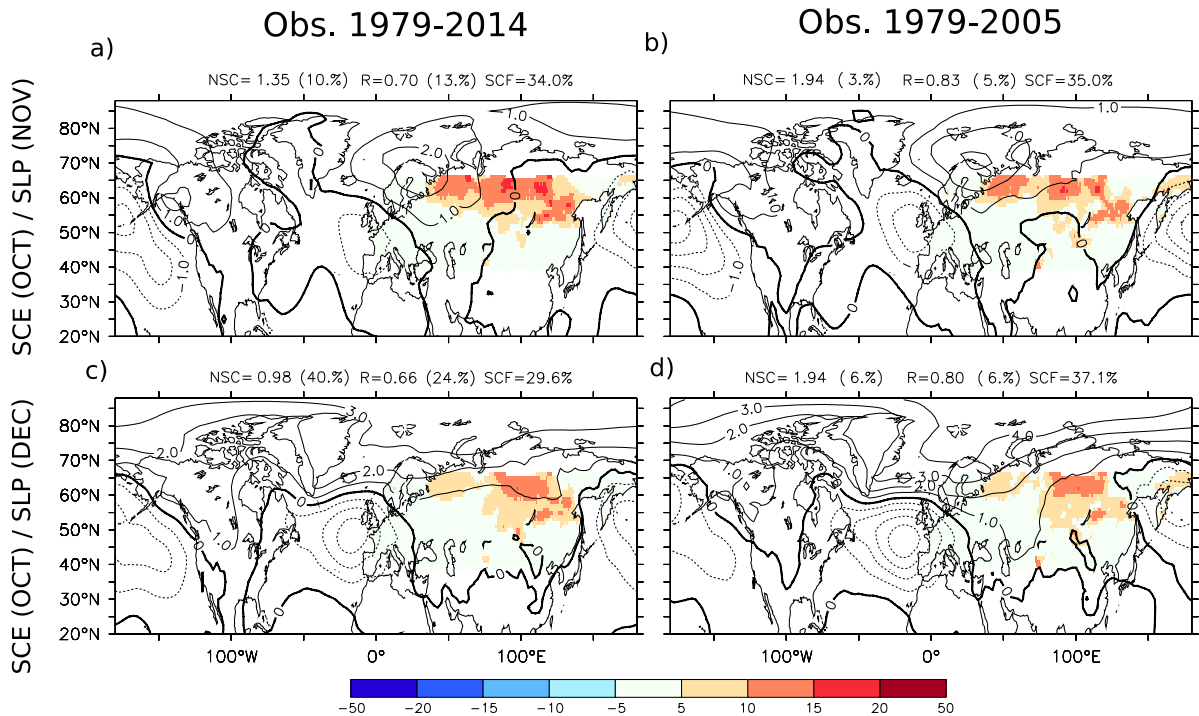


Fig. A1 : (a) Homogeneous October snow cover fraction (in %) and November heterogeneous SLP (in hPa) covariance maps for the first MCA mode, when the snow cover leads by one month the atmosphere, for ERA-Interim during 1979-2014. (b) Same as (a) but for the 1979-2005 period. (c) Same as (a) but using the December SLP. (d) Same as (c) but for the 1979-2005 period.