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1 **On the role of Eurasian autumn snow cover in**
2 **dynamical seasonal predictions**

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8 **Abstract** Seasonal predictions leverage on predictable or persistent components
9 of the Earth system that can modify the state of the atmosphere. The land surface
10 provides predictability through various mechanisms, including snow cover, with
11 particular reference to Autumn snow cover over the Eurasian continent. The snow
12 cover alters the energy exchange between surface and atmosphere and induces a di-
13 abatic cooling that in turn can affect the atmosphere locally and remotely. Lagged
14 relationships between snow cover in Eurasia and atmospheric modes of variability
15 in the Northern Hemisphere have been documented but are deemed to be non-
16 stationary and climate models typically do not reproduce observed relationships
17 with consensus. The role of the snow in recent dynamical seasonal forecasts is
18 therefore unclear. Here we assess the role of Autumn Eurasian snow cover in a set
19 of 5 operational seasonal forecasts with large ensemble size and high resolution
20 and with the help of targeted idealised simulations. Forecast systems reproduce
21 realistically regional changes of the surface energy balance. Retrospective forecasts
22 and idealised sensitivity experiments identify a coherent change of the circulation
23 in the Northern Hemisphere. The main features of the atmospheric response are
24 a wave-train downstream over the Pacific and North America and a signal in the
25 Arctic. The latter does not emerge in reanalysis data but is compatible with a
26 lagged but weak and fast feedback from the snow to the Arctic Oscillation.

27 **Keywords** Seasonal climate predictions · Eurasian Snow Cover

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28 1 Introduction

29 The role of autumn Eurasian snow cover for the interannual variability of the
30 Northern Hemisphere climate was examined by many studies. Observational evi-
31 dence promoted the concept of a causal relationship between snow cover and the
32 Northern Hemisphere annular mode. This linkage is also exemplified by the rela-
33 tionship between snow cover and the near-surface Arctic Oscillation (AO) or the
34 North Atlantic Oscillation [NAO, Saito and Cohen, 2003, Garfinkel et al., 2020].
35 The AO/NAO are dominant modes of variability for the circulation of, respectively,
36 the Northern Hemisphere and the north Atlantic. Emphasis has been posed on a
37 stratospheric pathway [Cohen et al., 2014a, 2007] that can be explained in terms of
38 a coupled troposphere-stratosphere adjustment to surface forcing [Reichler et al.,
39 2005, Fletcher et al., 2007]. This mechanism has implications for predictability on
40 subseasonal and seasonal time scales, as it implies that autumn snow cover induces
41 circulation changes in late autumn and in winter.

42 Years of development led to established empirical forecasts of either the AO or
43 the NAO that are based also on indices of snow cover variability or support indi-
44 rectly the role of the snow [e.g. Wang et al., 2017, Cohen and Jones, 2011]. More
45 recently, snow cover has been successfully exploited in the implementation of a
46 hybrid statistical-dynamical forecast of the NAO via a redefinition (subsampling)
47 of the ensemble that involves the state of snow cover [Dobrynin et al., 2018]. A
48 number of studies demonstrated a snow-AO mechanism in idealised simulations
49 [e.g. Smith et al., 2010, Fletcher et al., 2009, Orsolini and Kvamstø, 2009]. These
50 studies find an impact of the snow cover on the circulation over the Pacific and
51 North America that leads to enhanced wave propagation into the stratosphere.
52 Nonetheless, Hardiman et al. [2008] showed that CMIP3 climate models do not
53 reproduce the observed snow-stratosphere linkage and therefore fail to reproduce
54 the impact on the AO/NAO. More recently, Gastineau et al. [2017] identified a
55 snow-AO relationship in a small subset of CMIP5 models and called for targeted
56 experiments to assess the causal relationship. Furtado et al. [2015] also confirmed
57 that the relationship is not robust in CMIP5 models but emerges in some time in-
58 tervals, indicating a non-stationarity of the process, and revealed a common bias in
59 the modelled snow variability. Along these lines, recent studies [Peings et al., 2013,
60 Douville et al., 2017] highlighted the non-stationarity of the observed snow-AO re-
61 lationship in past decades. Li et al. [2019] examined the role of snow initialisation
62 in a seasonal forecast system and suggested a compromise of a weak feedback of
63 the snow onto the AO/NAO [see also Orsolini et al., 2016, Garfinkel et al., 2020].
64 Non-stationarity and model dependence of the snow-AO relationship pose ques-
65 tions on the role of snow in modern dynamical seasonal forecasts and the lack of a
66 multi-model assessment in an operational context is potentially a gap in our view
67 of the linkage. In this study we investigate the relationship between Eurasian snow
68 cover variability in a set of state-of-art operational forecast systems. As discussed
69 in section 2, by using the Copernicus Climate Change Service (C3S) retrospective
70 forecasts, we can take advantage of large ensembles and high resolution models.
71 We demonstrate consensus and realism in how models reproduce the modulation
72 of the surface energy balance by snow cover variability. As the hindcast time series
73 is short and the sampling of the state of the climate system is poor, signals in the
74 atmosphere can be explained by a forcing other than the snow or more generally
75 by a predictable component of the system. One example is variability of sea ice in

76 the Eurasian sector of the Arctic, which is known to be covariant with Eurasian
77 snow [Cohen et al., 2014b, Gastineau et al., 2017] and has been linked with cir-
78 culation changes in polar and subpolar regions [Ruggieri et al., 2016] on seasonal
79 time scales. Similarly, the winter circulation in the Northern Hemisphere can be
80 affected by a range of phenomena. Noticeable examples are El Nino Southern Os-
81 cillation and the Pacific-Decadal Oscillation [Benassi et al., 2021, Rao et al., 2019],
82 the Quasi-Biennial Oscillation [Rao et al., 2020], the Madden-Julian Oscillation
83 [Wang et al., 2020], the Atlantic Multidecadal Variability [Ruggieri et al., 2021,
84 Ruprich-Robert et al., 2017]. To account for the shortness of the hindcast period
85 and corroborate the attribution of the remote atmospheric adjustment, idealised
86 simulations with prescribed snow cover are performed. The synthesis of this com-
87 bined approach allows to identify robust features of the atmospheric response to
88 snow cover and disclose potential applications. Results are presented in section 3
89 and discussed in section 4.

90 2 Methodology

91 2.1 Reanalysis, seasonal forecast and idealised simulations

92 The ERA5 [Hersbach et al., 2020] and ERA5-Land [Muñoz-Sabater et al., 2021]
93 reanalyses of the European Centre for Medium-Range Weather Forecast are used
94 in this study to obtain an estimate of the state of the surface energy balance and of
95 the atmospheric circulation associated with Eurasian snow cover variability. Data
96 used include snow cover, two-metre temperature, sea level pressure, geopotential
97 height, surface thermal and shortwave net radiation, surface latent and sensible
98 heat flux. Data are obtained on a monthly basis from the Copernicus Data Store
99 on the native grid. Results presented in the study also include the analysis of re-
100 spective forecasts performed with 5 operational forecast systems contributing
101 to the Copernicus Climate Change Service (C3S). The retrospective forecasts or
102 hindcasts cover the period from 1993 to 2016 and we use the nominal start date
103 of October 1st. Models used are summarised in table 1 and details on model fea-
104 tures and initialisation techniques are available through the C3S documentation
105 (<https://confluence.ecmwf.int/display/CKB/C3S+Seasonal+Forecasts>). A quali-
106 tative description of the snow cover initialisation technique is given in table 1.
107 Some models are initialised via a reanalysis field, other via a forced run, one in-
108 directly via a coupled run with assimilation in the atmosphere. One advantage
109 of these forecast systems is the relatively high number of ensemble members that
110 allows to identify small signals. A drawback of the approach is that the time series
111 is short, a fact that undermines a clean attribution of the signal to snow cover vari-
112 ability. To corroborate findings obtained with seasonal forecast and to facilitate
113 the attribution of atmospheric signals to snow cover related surface forcing, we
114 use idealized simulations with atmospheric general circulation models (AGCMs)
115 performed within the MEDSCOPE project (<https://www.medscope-project.eu/>).
116 Models used are the atmospheric component of CMCC-SPS3 [Sanna et al., 2016, ,
117 one of the models contributing to the C3S forecast] hereafter referred to as CMCC-
118 AGCM, and the ARPEGE-Climat 6.3 [Roehrig et al., 2020] hereafter referred to
119 as MF-AGCM. A reference control (hereafter CONTROL) simulation and a forced
120 experiment (hereafter SNOW), designed to mimic a snow cover increase in Eurasia

121 (i.e. 42.5 to 72.5 °N, 40 to 180°E), have been performed. CONTROL is a 50-year
122 integration with climatological monthly SSTs [computed over the period 1981-
123 2010 from the HadiSST ocean analysis, Rayner et al., 2003] and using perpetual
124 radiative forcing at year 2000 (i.e. present-day conditions), after spinup. On the
125 other hand, SNOW is a 50-member ensemble of 6-month long integrations from
126 the 1st of October with initial conditions taken from CONTROL. To prescribe an
127 arbitrary snow cover field, the models land surface schemes have been modified
128 to inhibit the interactive snow formation, accumulation, and melt over the region
129 of interest. The methodology adapts the approach applied by Hauser et al. [2017]
130 on soil moisture to snow water equivalent. With this method we restore the snow
131 cover to a stationary state for the first two months of integration, while from De-
132 cember this constraint is not further applied and the system is free to evolve. The
133 snow water equivalent values are prescribed to the average November snow water
134 equivalent computed over the upper 25th percentile of the observed distribution
135 in the 1913-2012 period from the NOAA-20CR reanalysis [as in Douville et al.,
136 2017]. The quality of snow cover data in this reanalysis is also discussed in Douville
137 et al. [2017]. As shown in the Results section, this definition imposes a forcing that
138 mimics observed features of snow cover variability.

139 2.2 Methods

140 To quantify the variability of snow cover in Eurasia we introduce the Eurasian
141 Snow Cover (ESC) index adapted from Douville et al. [2017]. Compared to Douville
142 et al. [2017], we apply a small adjustment to the latitudinal boundaries to facilitate
143 the implementation of a snow cover restoring in idealised simulations. The ESC
144 index is defined as the area-weighted average of the snow cover area fraction (snow
145 cover) of ERA5-Land in the domain 40-180°E, 42.5-72.5°N in October. This index
146 is therefore proportional to the fraction of the domain covered by snow. Note
147 that in the reanalysis the snow cover is diagnosed from other prognostic variables.
148 The choice of the the snow cover over other snow-related variables is guided by its
149 established impact with the near-surface climate and is common to previous studies
150 [e.g. Douville et al., 2017, Gastineau et al., 2017]. For models, linear regressions
151 are computed using the ESC index of the reanalysis as predictor and the ensemble
152 mean of a modelled quantity as predictand. Statistical significance for regressions
153 has been computed using the Wald test [Wald and Wolfowitz, 1940] to reject the
154 null hypothesis that the slope is zero. The atmospheric response to snow forcing
155 in idealised experiments is defined as the difference between the ensemble mean of
156 SNOW and the ensemble mean of CONTROL. Statistical significance for idealised
157 experiments is assessed using a t-test to reject the null hypothesis that the means
158 belong to the same distribution. The AO pattern is obtained as the first empirical
159 orthogonal function of monthly sea level pressure, computed separately for each
160 month. The strength of the stratospheric polar vortex is diagnosed with the average
161 of the zonal mean zonal wind at 10 hPa in the band 50-65 °N [see e.g. Palmeiro
162 et al., 2015].

3 Results

This section is divided into three parts. The first one where the relationship between snow cover, the local surface energy balance and the circulation in seasons ahead is investigated using a reanalysis. In the second one, linear regressions based on seasonal hindcasts are analysed to identify a predictable component of the atmospheric response to snow cover variability. In the third and final one, results presented in previous sections 3.1 and 3.2 are discussed in view of a set of idealised sensitivity model simulations in which snow cover is increased in Eurasia.

3.1 Snow cover variability and teleconnection in the reanalysis

The time series of the ESC index in the period covered by C3S hindcasts is shown in Fig. 1 by a solid black line. We note a peak around year 2000 followed by a drop between 2003 and 2012 and a second peak at the end of the time series. This behaviour is broadly in agreement with satellite-derived estimates of snow cover [not shown, Estilow et al., 2015]. The role of snow cover in the surface energy balance was examined by Cohen and Rind [1991] and the impact of the snow cover variability on surface energy fluxes is demonstrated by Fletcher et al. [2009, see their figure 1]. The presence of snow increases the reflected amount of shortwave radiation by increasing the surface albedo and cools the surface. A cooler surface leads to reduced thermal and turbulent heat fluxes. Overall the combined effect of these changes cools the atmosphere and reduces the near surface air temperature. We see in Fig. 1 that the ESC index explains a large fraction of surface flux variability in its domain, with the correlation between ESC and the net shortwave flux reaching 0.75 (for the other components it is about -0.66). Note that in Fig. 1 the sign of the sum of thermal radiation and turbulent heat flux has been reversed to facilitate the comparison with the other time series. The correlation between the ESC index and grid point values of snow cover is shown in Fig. 2. It can be seen that the variability in the period 1993-2016 described by the ESC index is mostly representative of the western part of the sector, although positive correlations are found over more than 90% of the land grid points in the domain. Fig. 2 also suggests that October snow cover anomalies persist in November in some regions, and up to winter in the Ural region. We conclude that the ESC index is associated with regional snow cover variability in a vast portion of Eurasia and is closely representative of snow cover variability in the Ural-Western-Siberia region. We examine hereafter the surface energy balance and to account for the persistence of the snow cover anomalies (Fig. 2) we look at October-November bimonthly averages.

The relationship between the ESC index and the local surface energy balance in the reanalysis is examined in Fig. 3. Grey rectangles encompass the area used to compute averaged quantities displayed in Fig. 1. We can identify three regions where a considerable fraction of the variability of surface fluxes is explained by the ESC index: in the Ural region, in Eastern Siberia and in Western Siberia. In these regions, correlations reach about ± 0.75 and a linear regression suggests that the ESC index can explain up to about 10 Wm^{-2} .

The surface shortwave flux is positively correlated also in smaller scale regions in East Asia and a negative correlation is found in a small area to the south

208 and to the east of the Central Siberian Plateau. The negative correlation can be
209 interpreted in terms of a weaker snow cover variability in this region. The sum
210 of other components of the energy flux (Fig. 3,c) confirms the regional signature
211 of snow cover variability found in Fig. 3,a. These findings hold for the individual
212 addenda of Fig. 3,c (not shown). Such alteration of the surface energy balance
213 leads to a cooler atmosphere, as confirmed by Fig. 3,d, where we see that negative
214 correlations with near-surface air temperature cover most of the domain. The snow
215 cover variability captured by the ESC index can modulate the near surface air
216 temperature in Eurasia by 4 K, presumably with a stronger impact on the western
217 part of the domain, where also changes in surface fluxes are stronger. A fraction of
218 the temperature signal is likely explained by the atmospheric circulation in October
219 (Fig. 4,a), that features a barotropic low over western Siberia and the northerly
220 flow over the Ural region, coincident with the most negative correlation in Fig. 3,d.
221 A similar pattern, prior to episodes of snow cover increase in western Siberia on
222 an intraseasonal time scale, is discussed by Song and Wu [2019]. Moreover, we see
223 a barotropic high in the Arctic, from the Barents to the Laptev sea, and a surface
224 high in Eastern Siberia, both are statistically significant at the chosen confidence
225 level and the latter is likely explained by a shallow thermodynamic adjustment
226 to local snow cover increase, perhaps in phase with an independent surface high.
227 With the same rationale, the low over Siberia could be masking the snow feedback
228 on sea level pressure. These statements are discussed more in details in section 3.3.
229 In November (not shown) a signal is found in the extratropical Pacific with a deep
230 low covering most of the Northern part of the ocean and high located to the North.
231 A shallow high over Barents sea and Urals is also found and the barotropic signal
232 in the Pacific is persistent up to December and is detectable in seasonal average
233 of NDJ (Fig. 4,b). The lack of a negative AO response in winter is noticeable
234 and is common also to late winter (Fig. 4,c and d). The picture given by Fig. 4 is
235 likely to be contaminated by atmospheric noise that is unrelated to nor induced by
236 snow cover or any deterministic surface forcing. Ensembles of dynamical seasonal
237 forecast can help us detect a predictable component of the feedback of the snow
238 onto the atmosphere and they are analysed in section 3.2.

239 3.2 Eurasian snow cover in recent seasonal forecasts

240 Results presented in this section are based on regressions onto the ERA5-Land
241 ESC index of atmospheric and surface variables in a multi-model ensemble of sea-
242 sonal forecasts. The regressions is computed for ensemble mean quantities of each
243 model separately, over the period (1993-2016). Models used are described in sec-
244 tion 2.1 and in table 1. In Figs. 5 and 6 we can see that all models simulate the
245 observed modulation of the surface energy balance by snow cover and they repro-
246 duce regional features with good accuracy. In seasonal hindcasts, as for ERA5,
247 the surface forcing to the atmosphere associated with snow cover variability is
248 particularly strong in the Ural-Western-Siberia region. This analysis shows that
249 the current generation of dynamical seasonal forecasts reproduces the signature
250 of snow cover variability on a regional scale. We can therefore investigate the de-
251 pendence of the forecast of the tropospheric circulation on the snow cover extent
252 in Eurasia. In Fig. 7 we see the multi-model mean of the regression of Z500 and
253 SLP, i.e. the mean of the 5 regression coefficients computed for each model alone.

254 We can see that in regions where a signal emerges all models agree on the sign of
255 the regression (stippled regions). This result indicates that models agree on simu-
256 lating a linear dependence of the forecast on the Eurasian snow cover variability
257 captured by the ESC index. It also reassures that signals in the multimodel mean
258 are not dominated by one or some of the models of the ensemble. It can be seen
259 that, in October (Fig. 7,a), all models show a wave-train from the North Atlantic
260 which at 500 hPa peaks with a low over Siberia, where there is no signal in the
261 sea level pressure. The surface pressure peaks instead in the Barents sea and in
262 Scandinavia. The Scandinavian pattern and the low over western Siberia are pre-
263 cursors of snow cover increase in the region [Gastineau et al., 2017, Song and Wu,
264 2019], but a similar pattern emerges also in response to snow cover increase, as
265 shown later, and it is found in the reanalysis (Fig. 4). This configuration can be in-
266 terpreted assuming that the models predict the linear, local response over Eurasia
267 plus a predictable signal confined in the first weeks. The statement on the linear
268 response is confirmed by results presented in section 3.3. After the first month,
269 seasonal averages are dominated by the negative barotropic anomaly in the Pacific
270 and the positive one further downstream over the American continent. A shallow
271 high from the Barents to the Laptev sea develops throughout the seasons, deepens
272 and enlarges into the Labrador sea.

273 While the snow can be linked with the local adjustment of the pressure and of
274 the geopotential, the robust remote response in the Arctic and in the Pacific can
275 be explained also by a different forcing. Large ensembles of seasonal forecasts are
276 indeed a valuable tool to identify the predictable signal even if the magnitude of
277 the signal is small compared to that of the noise, but are so far available for a
278 relatively short period. The shortness of the time series introduces a major lim-
279 itation on the attribution of the signal. To mitigate this issue a set of idealised
280 simulations has been performed and results based on these idealised simulations
281 are presented in the following section.

282 3.3 Idealised simulations

283 Idealised simulations with two AGCMs have been performed to corroborate find-
284 ings obtained with the seasonal forecasts. The setup is such that the control simu-
285 lation performed with climatological sea ice and SSTs and freely evolving snow
286 cover (CONTROL) is compared with a perturbed run where the snow extent in
287 Eurasia is imposed to be substantially larger in October and November (SNOW).
288 Both are 6-month ensemble simulations starting in October. Details on the setup
289 and the models are given in section 2.1. The analysis is based on differences be-
290 tween the ensemble mean of the SNOW run minus that of the CONTROL run.

291 It can be seen from Fig. 8 that this idealised setup implies increased snow cover in
292 the southern and in the western part of the domain and this is a desired feature of
293 the surface forcing, in view of the results presented in Figs. 2, 4 and 5. However,
294 the forcing is particularly strong in the southern boundary of the domain and
295 covers also east Asia, which does not resemble closely the observed variability in
296 the hindcast period. The intensity of the forcing decreases substantially between
297 October and November especially for the CMCC model and a residual anomalous
298 snow cover in the western part of the domain persists into winter, up to January
299 (not shown). Overall this behaviour agrees qualitatively with the reanalysis. The

300 total snow cover extent of the domain in October is however exaggerated if com-
301 pared to the interannual variability of seasonal forecasts used in this study. Surface
302 fluxes (Fig. 9) are perturbed by the increased snow in a way that resembles what
303 found in sections 3.1 and 3.2 in reanalysis and seasonal forecasts. The associated
304 spatial patterns are not shown as they follow the pattern of snow cover change. We
305 note the positive heat flux anomaly in the western Pacific, that can be explained
306 in terms of a stronger heating by the ocean due to a colder overlaying atmosphere,
307 which in turn is explained by advection by the predominantly westerly wind of air
308 that is cooled by the presence of the snow. The abrupt fluctuation between 40 and
309 60 °E is explained by the intensified heat transfer from water to air in the Caspian
310 sea.

311 To condense and compare results obtained by the two idealised simulations, in Fig.
312 10 we show atmospheric fields averaged in the midlatitudes, precisely between 42.5
313 °N and 72.5 °N, which is the latitudinal extent of the domain used to impose the
314 snow cover perturbation and used in Fig. 3. As in forecast and reanalysis, the
315 surface cooling is stronger in the western part of the domain, by construction
316 in this case, and differences between the two models can be explained by differ-
317 ences in the surface flux change that in turn is proportional to a different snow
318 cover anomaly. The temperature change is found in the boundary layer and in
319 the free troposphere above and it persists but substantially decays in November.
320 The equivalent analysis for the circulation is rather interesting. In Fig. 10,c the
321 dominant feature is a strongly baroclinic response that can be understood in the
322 framework of a steady, linear adjustment a la Hoskins and Karoly [1981]. Both
323 models reproduce a shallow high to the east of the cooling, advection of warm
324 air from lower latitudes in the region of strong cooling (50-90 °E) and the vortic-
325 ity anomaly increasing with height as implied by the negative Z500 signal. This
326 regime of the response has been described for surface heating/cooling associated
327 with high-latitude SST and sea ice variability [Deser et al., 2010, Ruggieri et al.,
328 2017]. These studies found that the linear response is established within 1-2 days
329 and dominates the anomalous circulation up to two-three weeks. After that, a re-
330 sponse projecting onto a dominant mode of variability (AO-NAO) emerges. In our
331 case, the linear response indeed dominates in the first month and is confined in
332 the first 3 weeks of simulation (not shown). After one month, in November, albeit
333 a persistent, residual forcing, there is no evidence of a surface cooling out of the
334 boundary layer and of any signal attributable to the linear response. On the other
335 hand, the two models show a coherent signal in the Pacific and in North America,
336 broadly (but not exactly) coincident with a weakened Aleutian low. These findings
337 highlight the emergence of a lagged and remote response that is arguably crucial
338 for the predictability in the seasonal range.

339 The comparison between idealised experiments and seasonal forecasts is indeed
340 particularly insightful. Fig. 11 shows a 1-2 month lagged atmospheric response to
341 snow cover increase deduced with two different methods. In panel a), it is assessed
342 using the multi-model (C3S) regression on the ESC index based on ERA5-Land. In
343 panel b), it is assessed through the two-model ensemble mean difference in the ide-
344 alised sensitivity experiments. Substantial and intriguing similarities can be noted,
345 primarily in the Pacific-American sector. There is a wave-train peaking with a neg-
346 ative anomaly in the Eastern Pacific. In the Arctic, a rather shallow signal in the
347 Kara and Laptev seas is found. On a finer scale, there are important differences:
348 both are compatible with a local linear response to cooling at the surface in Eura-

349 sia, but this signature is remarkably stronger in the idealised experiments. This
350 is expected and consistent with a stronger surface forcing. Finally over the North
351 Atlantic and Europe, here it appears that uncertainty is larger, compared to other
352 regions. Both approaches feature a low over the subpolar-gyre/Nordic-Seas and
353 high over central and western Mediterranean. However the anomalous circulation
354 is substantially different in the Euro-Atlantic. Perhaps this region is more sensi-
355 tive to small differences in surface forcing, but in any case this sensitivity foresees
356 large intrinsic uncertainty. Robust features in the Arctic and in the Pacific are
357 confirmed by analysis of individual models. For instance, in Fig. 12, we can see
358 the comparison of the regression with the CMCC forecast system and with the
359 CMCC-AGCM, that is the atmospheric component of the forecast systems itself.
360 The structure with a low over Siberia, a high over the Arctic and a low over the
361 Eastern Pacific is confirmed. We also note that the CMCC and ECMWF fore-
362 cast systems feature a stronger signal in the Pacific and in the Arctic compared
363 to other models (not shown). This is in agreement with a stronger signal in the
364 surface fluxes (Figs. 5 and 6). Fig. 12 also confirms a large uncertainty for the
365 response in the Atlantic.

366 As discussed in the Introduction, previous studies have identified a stratospheric
367 pathway for a delayed atmospheric response to snow cover. The timing of the ro-
368 bust features of the response can help understand whether the signal is related
369 to a stratospheric pathway. The North Pacific response (Fig. 13a,b) is relatively
370 fast, peaks in November and can be seen in bimonthly averages up to November-
371 December, models agree on the timing of the signal. The shallow Arctic high (Fig.
372 13c,d) peaks later through the season and models slightly disagree on the timing
373 with CMCC-AGCM showing the peak in December and MF-AGCM in January.
374 Bimonthly averages show clearly that this signal persists into winter. In the strato-
375 sphere there is no evidence of any significant modification of the intensity of the
376 polar vortex. We measure it with the average zonal mean zonal wind in the polar
377 vortex edge (Fig. 13e,f, Palmeiro et al. [2015]). There is no evidence of a role for
378 the stratosphere in the mechanism, but the tropospheric signal projects onto the
379 model AO pattern (see table 2). Indeed, the peak of the signal in the Arctic (Fig.
380 13c,d) is associated with a weak negative AO signal.

381 Forecasts and idealised simulations coherently suggest a potential for the pre-
382 dictability of temperature and precipitation over land driven by snow cover vari-
383 ability in some regions of Europe and North America. The latter can be seen in
384 Fig. 14, where again forecast and idealised simulations provide a similar picture
385 for the precipitation response over some areas of the extra-tropical North America.
386 We note in particular sporadic signals of continental drying and a dipolar pattern
387 of precipitation along the East coast of the continent. Nonetheless, there is lit-
388 tle agreement with the corresponding regression with reanalysis. Differences with
389 reanalysis (not shown) are attributable to a different tilt of the Pacific wave-train.

390 4 Concluding remarks

391 In this study retrospective seasonal forecasts with operational systems are com-
392 bined with idealised simulations with state-of-the-art AGCMs to assess the atmo-
393 spheric response to Eurasian snow cover variability in recent decades (1993-2016).
394 Seasonal hindcasts reproduce regional features of Eurasian snow cover variabil-

395 ity and the associated change in the surface energy balance. Reanalysis, seasonal
396 forecasts and idealised experiments converge in indicating that an anomalous cir-
397 culation in the Pacific is established after the first month and persists for about
398 two months (Figs. 4, 7, 11, 13). Forecasts identify a robust relationship between
399 snow cover and atmospheric circulation that can be attributed to snow cover vari-
400 ability by AGCM idealised simulations (Fig. 11). The modelled response projects
401 onto the annular mode of variability of the model, but notably there is no evidence
402 of changes in the stratospheric circulation (Fig. 13).

403 The geographical position of the surface forcing determines how the atmospheric
404 flow responds. For the case of the snow, this can be understood primarily in terms
405 of a linear interference between anomalous and climatological waves [Smith et al.,
406 2010], but in also in view of how synoptic eddies respond [Ruggieri et al., 2019].
407 For the period considered in our study, autumn snow cover variability in Eurasia is
408 dominated by western Siberia, but this does not generally apply to the twentieth
409 century [Brown, 2000]. It follows that the lack of a stratospheric response could
410 be linked with this specific feature of the forcing (e.g. whether it is stronger in
411 western or eastern Siberia). On the other hand, part of the uncertainty of the re-
412 sponse could be mitigated if we were able to sample the role of regional snow cover
413 variability in sub-domains of Eurasia. In this sense, it could be that operational
414 forecasts draw predictability from snow cover well beyond the picture given by
415 Fig. 11. An interesting question is whether the snow-cover forcing closer to winter
416 shows an impact on the polar vortex. Portal et al. [2021] showed that November
417 start dates of the same seasonal forecasts used in this study are able to simulate
418 realistically the variability of the polar vortex, the upward wave propagation and
419 their relationship. It is therefore unlikely that the lack of the stratospheric path-
420 way in this study is explained by model biases in the troposphere-stratosphere
421 interaction. Portal et al. [2021] also find a weak and model dependent relation-
422 ship between snow cover variability and the intensity of the polar vortex in the
423 examined period for the November start date, that is consistent with results of
424 Garfinkel et al. [2020] and Orsolini et al. [2016]. It is therefore plausible that start
425 dates closer to winter may reveal a different picture.

426 Some inconsistencies between models are found in regional features of the surface
427 flux variability associated with snow variability (Fig. 6). This can be partly ex-
428 plained by differences in the snow initialisation techniques. The use of reanalysis
429 initialization for land surface could generate inconsistencies, resulting in a degra-
430 dation of the forecast quality [Materia et al., 2014]. Improvements in the land
431 initialisation and in land-atmosphere coupled data assimilation may reduce the
432 model uncertainty for the atmospheric response to Eurasian snow cover variability
433 and the value of development in this direction is advocated by our analysis. We
434 find regional impacts in other regions, one example is the western Mediterranean
435 where models show a surface high (Fig. 11) and reduced precipitation (not shown).
436 But as this signal comes with a model dependent large scale response in Europe,
437 associated impacts should be mentioned with caution. The substantial agreement
438 between forecast, idealised simulations and to some extent reanalysis gives a clear
439 picture of the atmospheric response to snow cover variability for many regions but
440 the North Atlantic. This is arguably closely linked with the lack of role for the
441 stratosphere, which also limits the lead time of the predictable component of the
442 response to 2-3 months after initialisation. It is likely that persistence and predic-
443 tion of snow cover anomalies will be crucial for practical purposes. In terms of the

444 Arctic Oscillation, our study indicates a weak but robust tropospheric feedback
445 from the snow to the AO.

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Table 1 Description of the seasonal prediction systems used in this study. Columns indicate, from left to right, the modelling centre and the system version, the resolution, the size of the ensemble the method used to initialize snow cover, the land model and a reference for the land model. The JRA55 reanalysis is documented in Kobayashi et al. [2015]. The ERA-Interim (ERA-I) reanalysis is documented in Dee et al. [2011]. Here a forced run is an offline land-only run forced with meteorological forcing.

System	Resolution	Ens. Size	Snow I.C.	Land model	Ref.
CMCC 3	1×1 L46	40	Forced run	CLM4.5	Oleson et al. [2013]
MF 6	TL359 L91	25	ERA-I	SURFEX v8.1	Le Moigne et al. [2009]
ECMWF 5	T _{CO} 319 L91	25	Forced run	IFS 43r1	Johnson et al. [2019]
DWD 2	T127 L95	30	Indirect	JSBACH	Brovkin et al. [2009]
UKMO 13	N216 L95	21	JRA55	JULES GL 6	Walters et al. [2017]

Table 2 Projection onto the model AO pattern of the response (SNOW minus CONTROL) in idealised simulations with the CMCC-AGCM model and the MF-AGCM model.

Month	CMCC	MF
OCT	-0.17	0.23
NOV	-0.33	0.53
DEC	-1.22	-0.10
JAN	-0.05	-0.71
FEB	0.29	-0.23

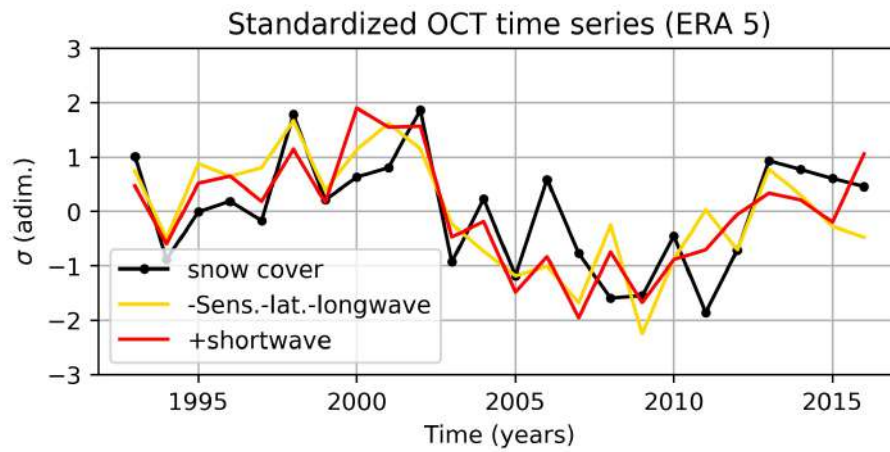


Fig. 1 Standardized time series of October snow covered area fraction (snow cover, black line) and October surface heat fluxes over Eurasia (42.5°N-72.5°N 40°E-180°E, only grid point entirely over land are used) derived from ERA5-Land in the period covered by C3S seasonal forecast. Fluxes are defined as positive when upward. The red line indicates the shortwave flux, while the yellow line indicates the sum of sensible, latent and longwave flux with a minus sign.

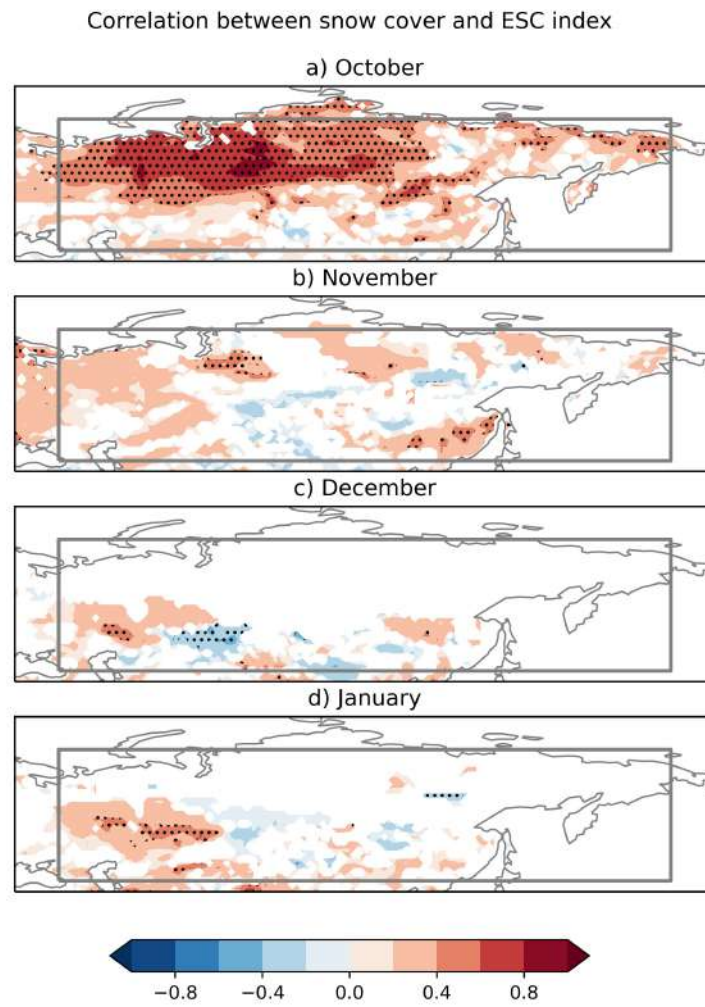


Fig. 2 Correlation between monthly mean snow cover in ERA5-Land and the October ESC index. Values between -0.1 and 0.1 are not displayed. Grey rectangles encompass the area used to compute averaged quantities displayed in Fig. 1.

Correlation (shading) and regression with ESC index

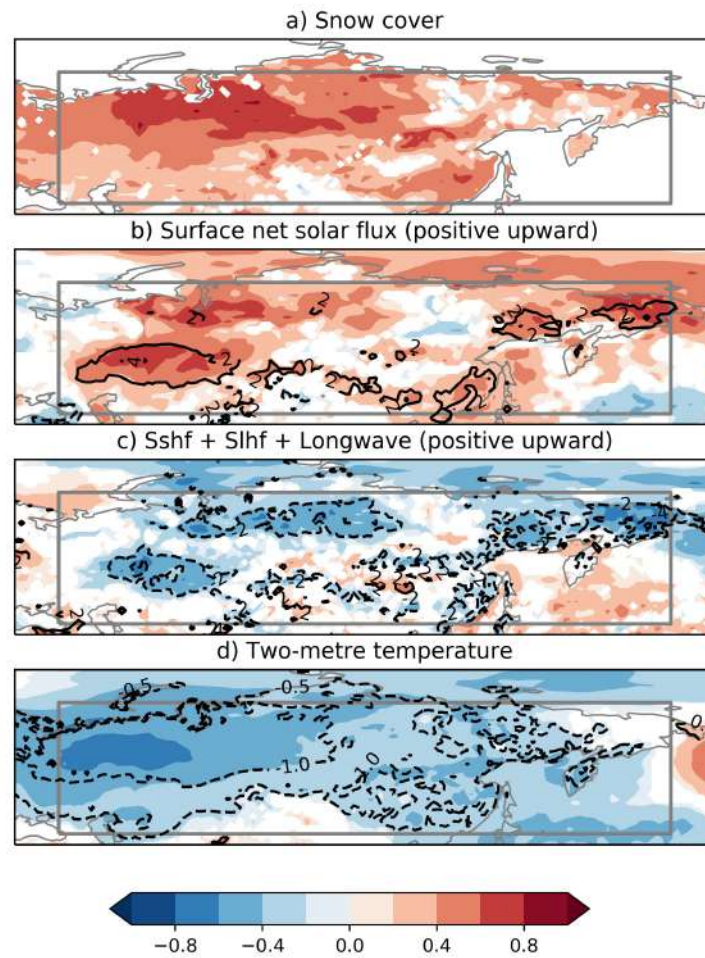


Fig. 3 Correlation (shading) and regression (contours) of October-November mean ERA5-land snow cover, and ERA5 surface fluxes and two-metre temperature (T2m) and the October ESC index in ERA5-Land, shown in Fig. 1 with a black line. All fluxes are defined as positive when upward. a) Correlation of snow cover. b) Correlation and regression of the sum of sensible heat flux, latent heat flux and net longwave flux. c) as in b) but for the net shortwave flux and d) for two-metre temperature. Units for the regression are Wm^{-2} in b) and c) and K in d). Values between -0.1 and 0.1 are not displayed.

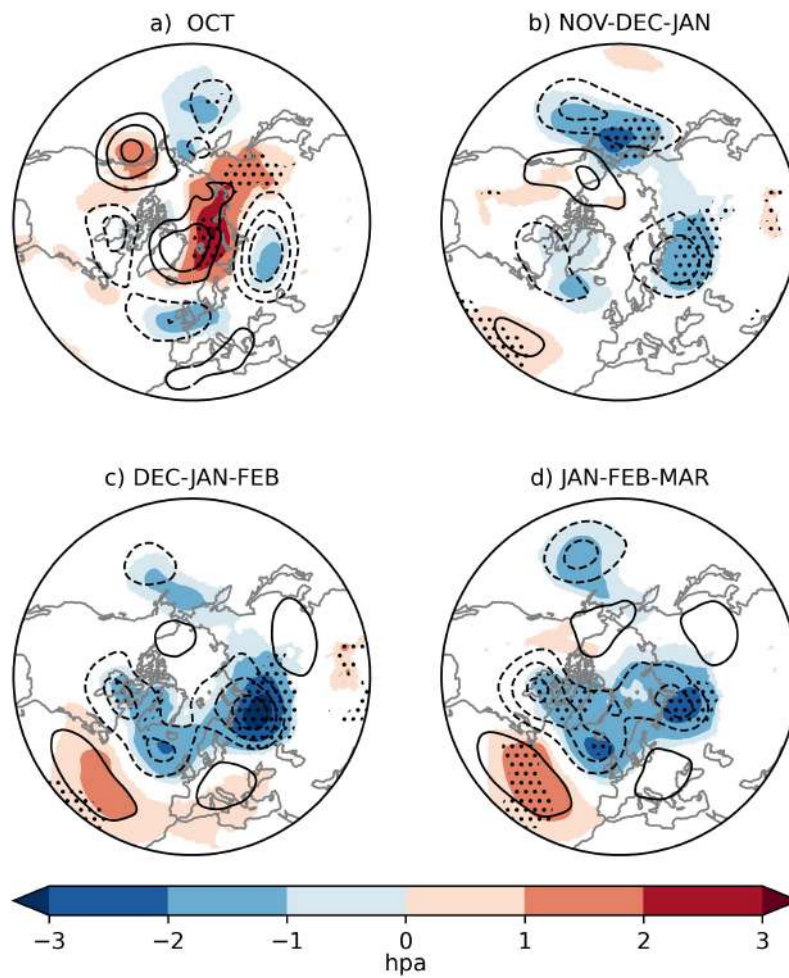
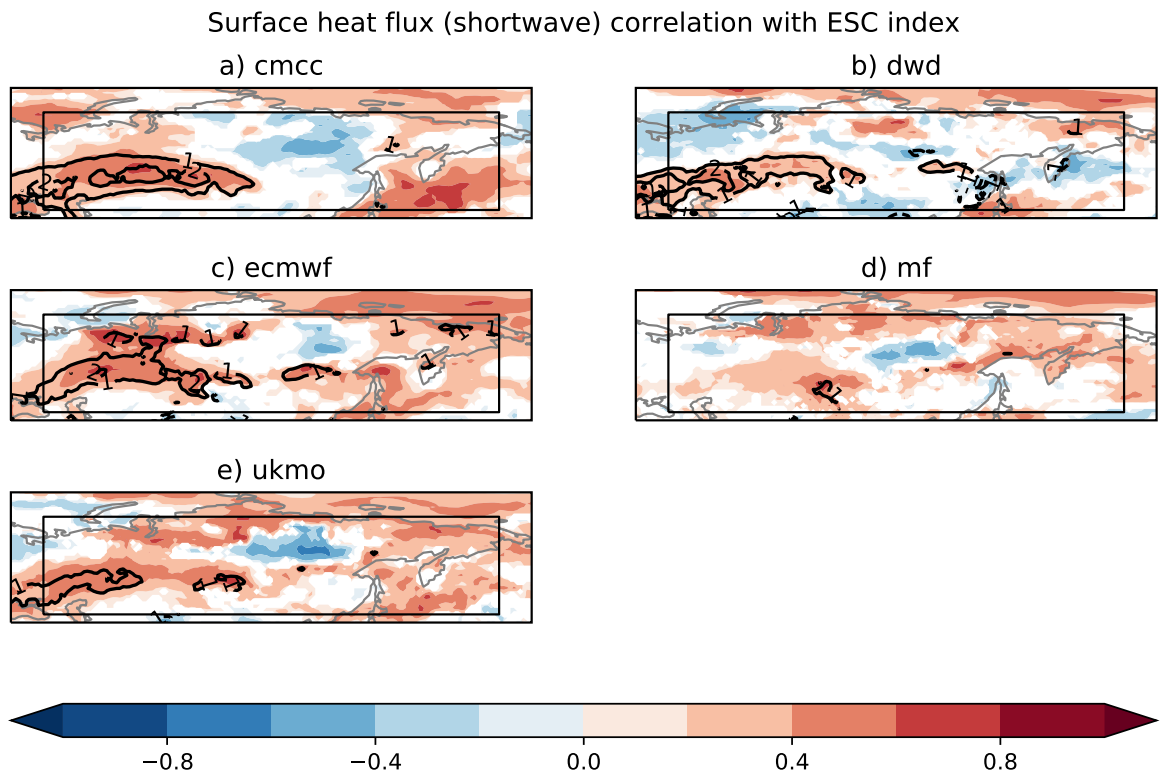
ERA5 SLP (hPa/ σ) and Z500 (m/ σ) regression on ESC index

Fig. 4 ERA5 regression coefficients on the October ESC index of Z500 (contours drawn at ± 10 , 20 and 30 m) and SLP (shadings, hPa) averaged in a) October, b) OND, c) NDJ and d) DJF. Stippling indicates region where the regression is statistically significant at 95% confidence level. Values between -0.5 and 0.5 hPa are not displayed.



-0.8 -0.4 0.0 0.4 0.8

Fig. 5 Correlation (shadings) and regression (contours, drawn at $\pm 1, 2 \text{ Wm}^{-2}$) between the flux of shortwave radiation and the October ESC index in ERA5-Land for 5 seasonal hindcasts. Values between -0.1 and 0.1 are not displayed.

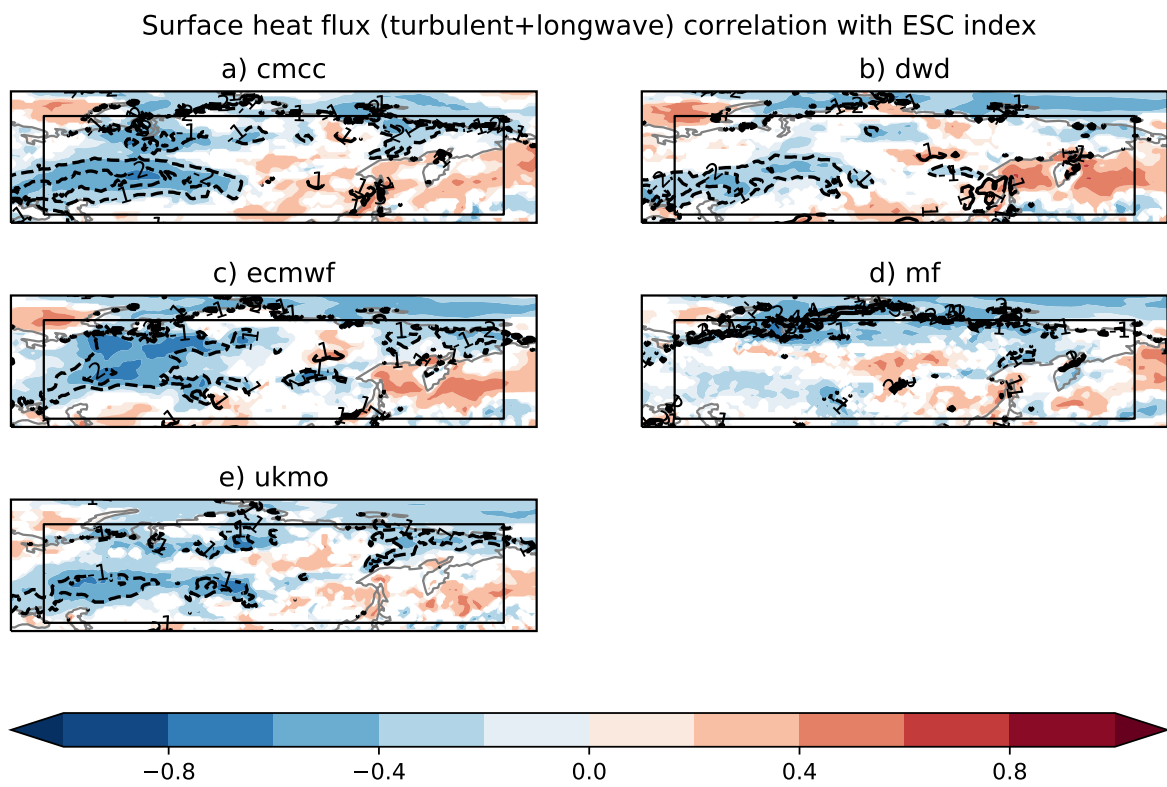


Fig. 6 As in Fig. 5 but for the sum of sensible heat, latent heat and thermal radiation.

Multi model mean of SLP (hPa/ σ) and Z500 (m/ σ) regression on ESC index

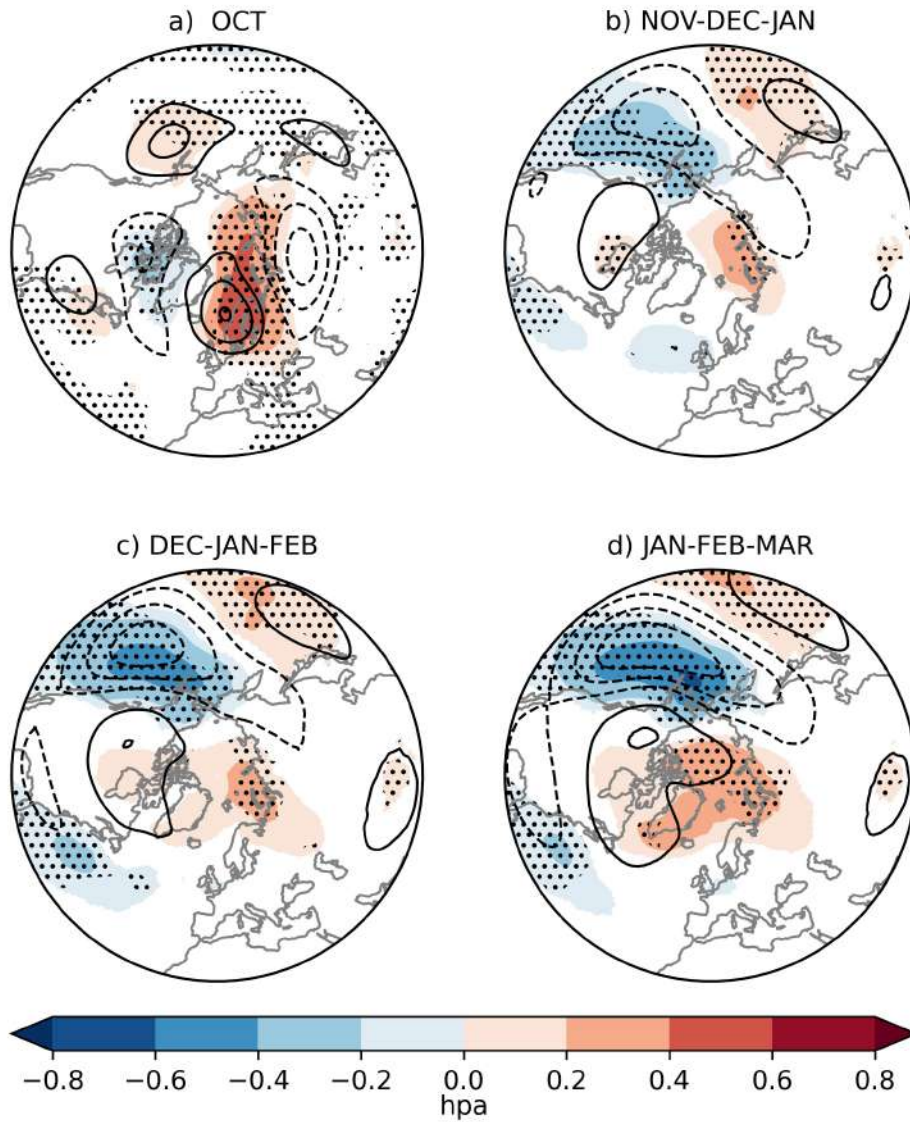


Fig. 7 Multi-model (5 seasonal hindcasts) mean of regression coefficients on the October ESC index in ERA5-Land of Z500 (contours drawn $\pm 6, 12$ and 18 m in a) and at $\pm 2, 4$ and 6 m in b),c) d)) and SLP (shadings, hPa) averaged over a) October, b) NDJ, c) DJF and d) JFM. Stippling indicates region where all models have the same sign for SLP. SLP values between -0.1 and 0.1 hPa are not displayed and SLP values in a) have been scaled by a factor of $1/3$.

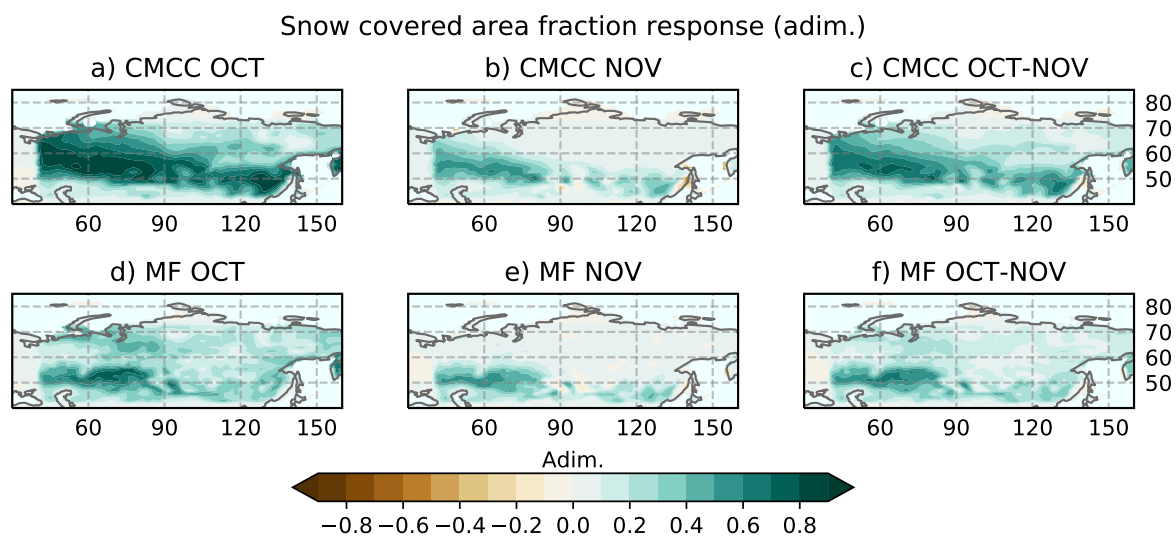


Fig. 8 Response of the snow cover in the idealized AGCM experiments in October (a, d), November (b, e) and their average (c, f) for the CMCC-AGCM model (top row) and the MF-AGCM model (bottom row).

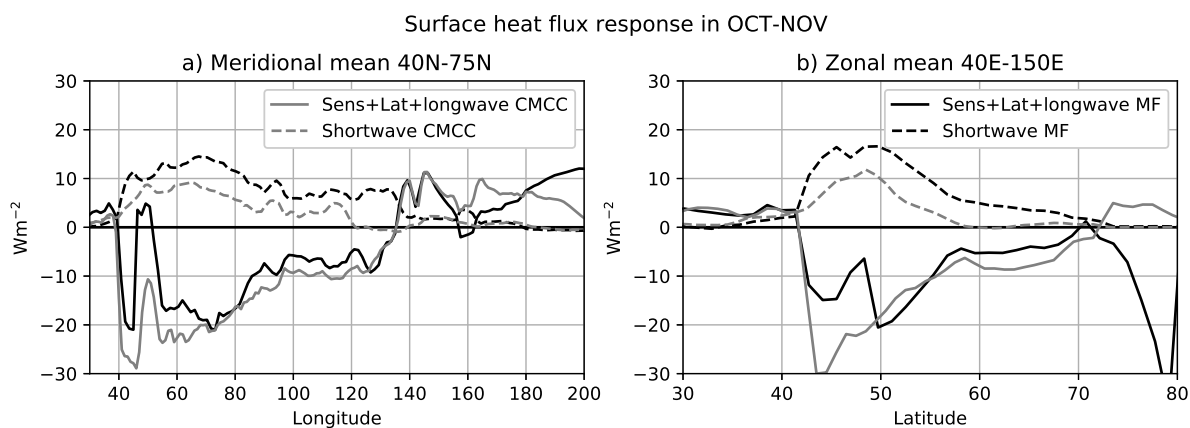


Fig. 9 Surface heat flux response in October and November in the idealised experiments with the CMCC-AGCM (grey lines) and MF-AGCM (black lines) model. The dashed lines are used for shortwave radiation and the solid lines for the sum of longwave, sensible and latent heat fluxes. In panel a) fluxes are averaged meridionally (weighted by the area) between 40 and 70 °N, in b) they are averaged zonally between 40 and 150 °E.

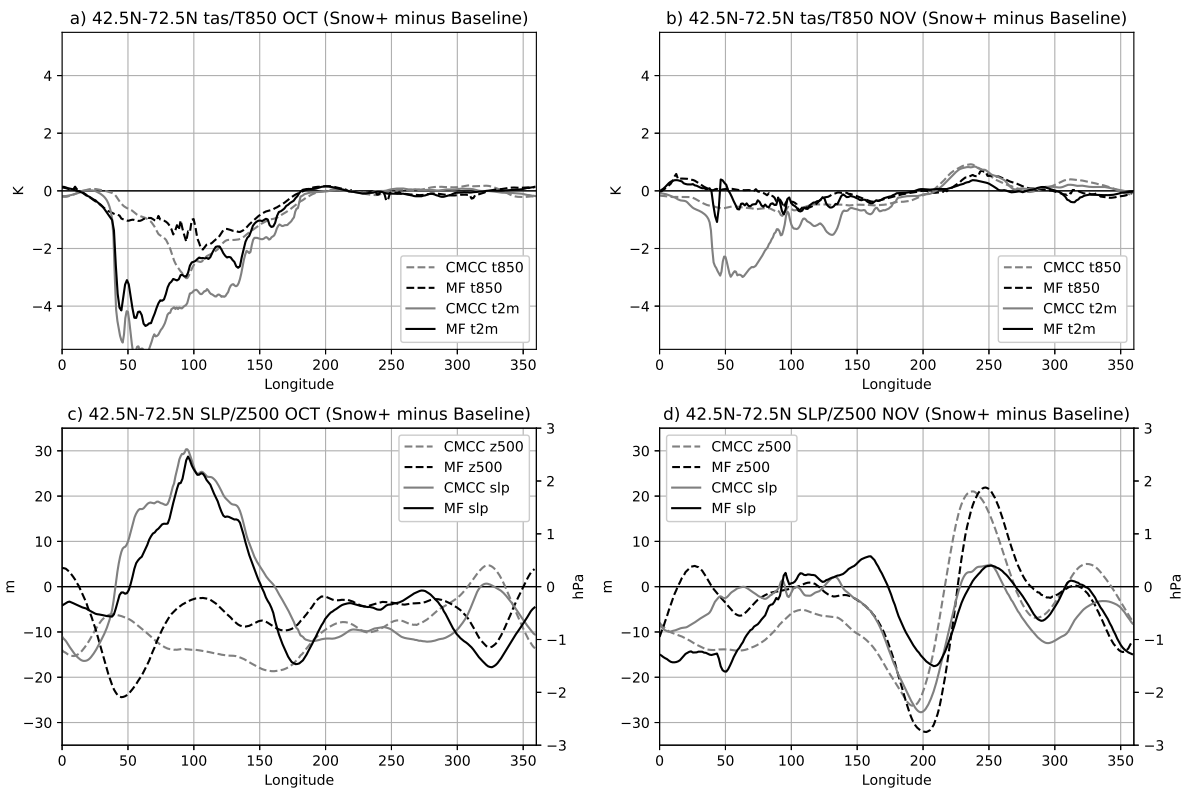


Fig. 10 Meridional mean (area-weighted) between 42.5N and 72.5N of T2m (solid lines, K) and T850 (dashed lines, K) in a) OCT and b) NOV for the idealized AGCM experiments with the CMCC-AGCM model (grey lines) and the MF-AGCM (black lines) model. c) and d) As in a) and b) but for SLP (solid lines, hPa) and Z500 (dashed lines, m).

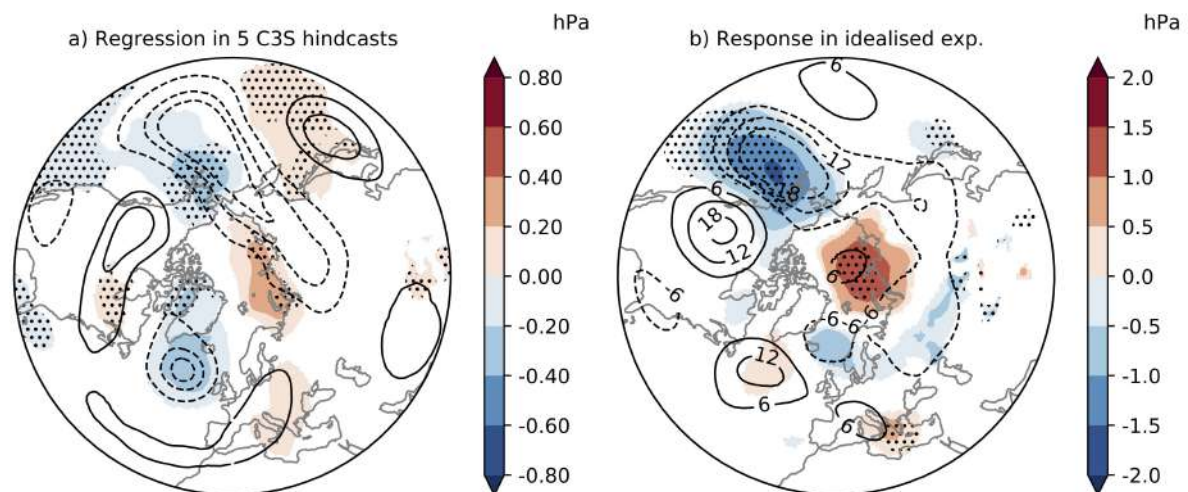


Fig. 11 November-December Atmospheric response to October snow cover increase inferred from C3S hindcasts (a) and idealised simulations (b). In a), the multi-model mean regression coefficient is shown for Z500 (contours drawn at $\pm 1.5, 2.25, 3, 3.75$ m) and SLP (shadings). Stippling indicates regions where all models agree on the sign of the regression. In b), the response to snow cover defined two-model mean of the ensemble mean difference (SNOW-CONTROL). As in a), Z500 is shown by contours drawn at $\pm 6, 12, 18$ and SLP is shown with shadings. Stippling indicates regions where the two models have the same sign of the response.

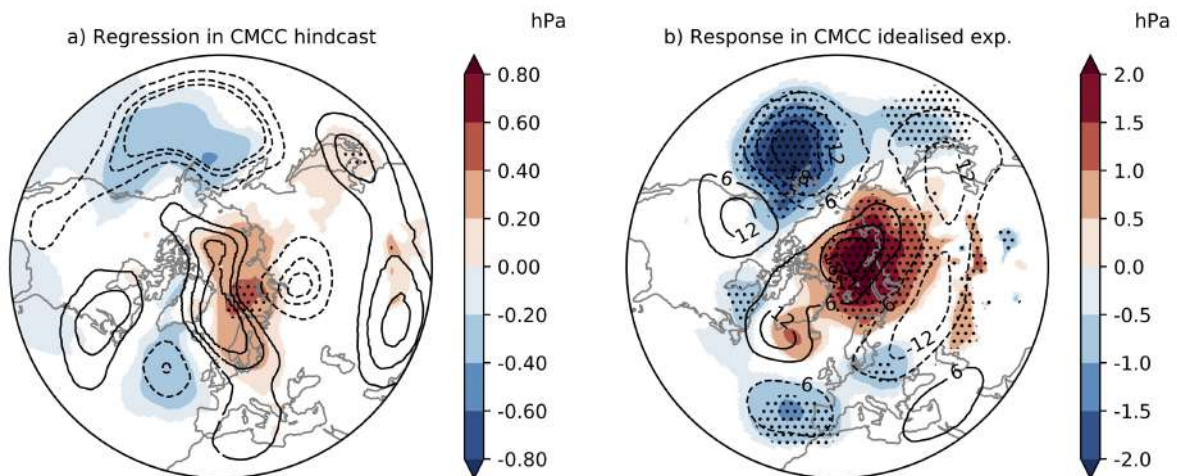


Fig. 12 As in Fig. 11 but using the CMCC-AGCM model instead of the multi-model mean. Stippling indicates statistically significant values at 95% confidence level.

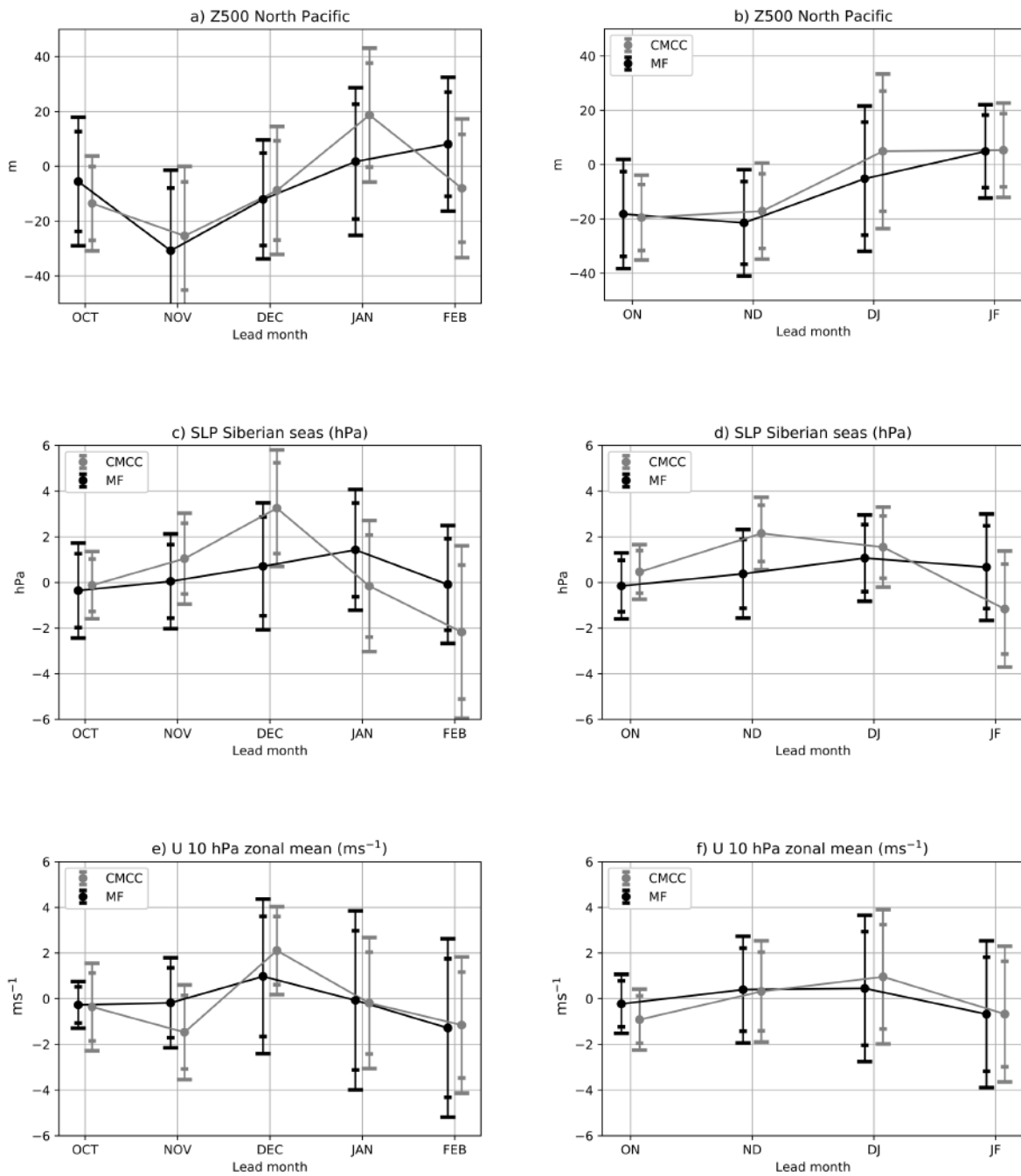


Fig. 13 Circulation response in the idealised simulations with the CMCC-AGCM (grey lines) and MF-AGCM (black lines) models. Bars indicate the confidence interval at 80% and 90% confidence level. Values are averaged monthly in the left column and bimonthly in the right column. a) and b) show Z500 averaged in the Northern extratropical Pacific (45-65 °N, 180-210 °E). c) and d) show SLP averaged in the Siberian sector of the Arctic ocean (75-90 °N, 55-150 °E, over the East Siberian sea and the Laptev sea). e) and f) zonal mean of the zonal wind at 10 hPa between 55 and 65 °N.

Multi model mean of Total Precip. (mm day^{-1}) and Z500 (m)

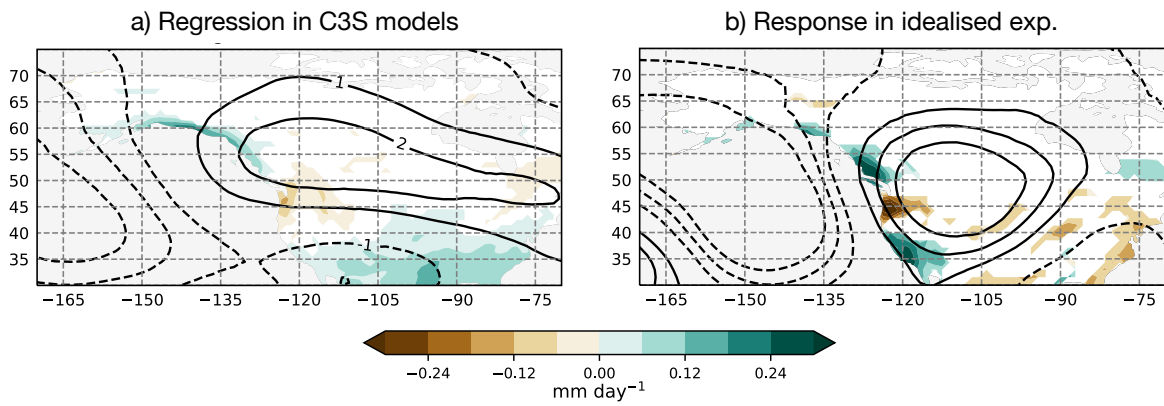


Fig. 14 a) Multi-model mean regression coefficient with the ESC index in ERA5-Land of Z500 (contours, drawn every 1 m, zero line omitted) and precipitation (shadings, mm day^{-1}) in seasonal hindcasts. b) Multi-model mean response to snow cover increase in idealised simulation. Contours of Z500 are drawn every 6 m (zero line omitted) and shadings are used for precipitation.