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**Atmospheric impacts of Arctic sea-ice loss, 1979-2009: Separating forced change from
atmospheric internal variability**

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28 **Abstract**

29

30 The ongoing loss of Arctic sea-ice cover has implications for the wider climate system. The
31 detection and importance of the atmospheric impacts of sea-ice loss depends, in part, on the relative
32 magnitudes of the sea-ice forced change compared to natural atmospheric internal variability (AIV).
33 This study analyses large ensembles of two independent atmospheric general circulation models in
34 order to separate the forced response to historical Arctic sea-ice loss (1979-2009) from AIV, and to
35 quantify signal-to-noise ratios. We also present results from a simulation with the sea-ice forcing
36 roughly doubled in magnitude. In proximity to regions of sea-ice loss, we identify statistically
37 significant near-surface atmospheric warming and precipitation increases, in autumn and winter in
38 both models. In winter, both models exhibit a significant lowering of sea level pressure and
39 geopotential height over the Arctic. All of these responses are broadly similar, but strengthened
40 and/or more geographically extensive, when the sea-ice forcing is doubled in magnitude. Signal-to-
41 noise ratios differ considerably between variables and locations. The temperature and precipitation
42 responses are significantly easier to detect (higher signal-to-noise ratio) than the sea level pressure
43 or geopotential height responses. Equally, the local response (i.e., in the vicinity of sea-ice loss) is
44 easier to detect than the mid-latitude or upper-level responses. Based on our estimates of signal-to-
45 noise, we conjecture that the local near-surface temperature and precipitation responses to past
46 Arctic sea-ice loss exceed AIV and are detectable in observed records, but that the potential
47 atmospheric circulation, upper-level and remote responses may be partially or wholly masked by
48 AIV.

49 **Introduction**

50

51 One of the clearest manifestations of recent climate change is the loss of summer and autumn sea-
52 ice cover in the Arctic (Stroeve et al., 2011). During the 2012 melt season, the Arctic sea-ice extent
53 shrunk to the lowest value in the satellite record, which began in 1979 (Zhang et al., 2013;
54 Parkinson and Comiso, 2013). Especially rapid sea-ice melt occurred during August 2012 at the
55 time of a ferocious storm (Simmonds and Rudeva, 2012), though model hindcasts suggest that a
56 new sea-ice minimum would have been recorded even without this storm (Zhang et al., 2013). The
57 last six years (2007-2012) have witnessed the six lowest September sea-ice extents on record,
58 possibly suggesting a “tipping point” has been passed (Livina and Lenton, 2013). Recent dramatic
59 sea-ice reductions augment longer-term trends, but statistically significant sea-ice extent reductions
60 are apparent in all calendar months even if the last six years are excluded (Kay et al., 2011).

61

62 The ongoing retreat of Arctic sea-ice has implications for the climate system. In order to better
63 understand these, a number of studies have perturbed sea-ice conditions in atmospheric general
64 circulation models (AGCMs) and examined the atmospheric response (e.g., Singarayer et al., 2006;
65 Seierstad and Bader, 2009; Deser et al., 2010; Strey et al., 2010; Bluthgen et al., 2012; Orsolini et
66 al., 2012; Ghatak et al., 2012; Porter et al., 2012; Screen et al., 2012; 2013). In a model setting, the
67 sea-ice cover can be manipulated in a controlled manner to reveal how and by what processes it
68 affects the wider climate system. These studies have identified some robust and reasonably well-
69 understood features of the local atmospheric response to sea-ice loss (i.e., impacts proximate to
70 regions of sea-ice loss). These include warming and moistening of the lower troposphere and
71 increases in cloud cover and precipitation.

72

73 The impacts of Arctic sea-ice loss may not be limited to the high latitudes. Increasing attention is
74 now turning to the potential remote impacts of Arctic sea-ice loss, including possible changes in

75 mid-latitude weather (Honda et al., 2009; Petoukhov and Semenov, 2012; Liu et al., 2012; Francis
76 and Vavrus, 2012; Screen and Simmonds, 2013a,b). Progress in understanding the potential large-
77 scale or remote impacts of Arctic sea-ice loss is hampered by large uncertainties in the atmospheric
78 circulation response to sea-ice loss. Observational studies suggest links between autumn sea-ice loss
79 and circulation patterns in the following winter (Francis et al., 2009; Overland and Wang, 2010; Wu
80 and Zhang, 2010; Strong et al., 2010; Jaiser et al., 2012), but the statistical significance of these
81 linkages has been questioned (Hopsch et al., 2012), causality is unclear and the mechanisms are
82 poorly understood. In model simulations, the spatial pattern, strength, statistical significance and
83 timing of the circulation response to sea-ice loss differs considerably between studies, and can be
84 hard to disentangle from atmospheric internal variability (AIV). AIV, also known as “climate
85 noise”, arises from non-linear dynamical processes intrinsic to the atmosphere (see, e.g., Deser et
86 al., 2012 and references therein).

87
88 In an attempt to better separate, and quantify, the potential forced response to Arctic sea-ice loss
89 and AIV, this manuscript presents results from large ensembles with two independent models, in
90 which the only prescribed forcing was observed Arctic sea-ice loss. Both models have been run
91 multiple times with identical surface boundary conditions and external forcing, with each run
92 beginning from a different atmospheric initial state. Therefore, the differences between the
93 simulated atmospheric states of each of the ensemble members arise only due to AIV. These
94 ensembles are approximately a factor of ten larger than those used in Screen et al. (2013), and
95 appreciably larger than in most of the studies mentioned above. In part, we seek to answer the
96 question: how many ensemble members are required to detect a significant response (in a particular
97 variable) to Arctic sea-ice loss, if indeed it is possible to detect a significant response at all? This is
98 pertinent to assessing the strength of the forced change compared to AIV and hence, whether it may
99 be observable in the real world.

101 **Data and Methods**

102

103 *Simulations*

104

105 We utilise two independent AGCMs: the UK-Australian Unified Model (UM) version 7.3 and the
106 US National Center for Atmospheric Research (NCAR) Community Atmosphere Model (CAM)
107 version 3. The UM has been developed by the UK Meteorological Office Hadley Centre and is the
108 atmospheric model used in their Global Environmental Model version 2 (HadGEM2) and in the
109 Australian Community Climate and Earth System Simulator (ACCESS). Both HadGEM2 and
110 ACCESS are participating models in the fifth Coupled Model Intercomparison Project (CMIP).
111 The configuration of the UM used here has 38 vertical levels with a horizontal resolution of 1.25
112 degrees of latitude by 1.875 degrees of longitude. CAM is the atmospheric component of the NCAR
113 Community Climate System Model version 3 (CCSM3), which participated in the third CMIP. The
114 version used here has 26 vertical levels and a spectral resolution of T42, roughly equivalent to 2.8
115 degrees of latitude and longitude. For further details the reader is directed to Martin et al. (2011)
116 and Bi et al. (2013) for the UM/ACCESS and Collins et al. (2006) for the CAM.

117

118 We primarily analyse two distinct simulations, performed identically with each model, termed the
119 control (CTRL) and perturbation (PERT) simulations. In CTRL, the models were prescribed with
120 an annually-repeating monthly cycle of climatological (CLM) sea-ice concentration (*SIC*) and sea
121 surface temperature (*SST*). Monthly-mean *SIC* and *SST* were taken from the Hurrell et al. (2008)
122 data set, updated to 2009, which is derived from a combination of in situ and remotely-sensed
123 observations. In PERT, the linear trend (TRD) in *SIC* over 1979-2009 for each month was added to
124 the climatological monthly values and these CLM+TRD values were prescribed in the models. The
125 prescription of *SST* in PERT was based on the approach introduced by Screen et al. (2013) and was
126 as follows. In grid-boxes and months where the *SIC* TRD is not zero, then the CLM+TRD *SST* were

prescribed. Elsewhere, CLM *SST* was prescribed. This approach captures *SST* changes directly related to *SIC* changes, but does not include *SST* changes outside the sea-ice zone (see Screen et al., 2013 for further details and justification). CTRL and PERT were run for 100 years in the UM and for 60 years in CAM. Since the prescribed surface forcing repeats annually, but the atmospheric initial conditions vary, each year is considered to be an independent ensemble member (atmospheric “memory” is negligible from year-to-year).

A further simulation has been performed with the UM, termed the PERT*2 simulation, in which the linear trends in *SIC* were doubled before being added to climatological values and these CLIM+(TRD*2) values were prescribed in the model. *SST* were prescribed as above, but this time with CLM+(TRD*2) values in place of CLM+TRD values. PERT*2 was run for 100 years in the UM only.

To isolate the atmospheric impacts of sea-ice loss, we compare the ensemble-mean of a particular variable in CTRL with the ensemble-mean in PERT or PERT*2. The ensemble-mean difference, PERT-CTRL or PERT*2-CTRL, is referred to as the “response” to Arctic sea-ice loss in the single- or double-perturbation experiments. We refer to “local” and “remote” responses, by which we mean responses that are in close proximity to sea-ice changes and those that are geographically distant from sea-ice changes, respectively.

Statistical methods

To test the statistical significance of the ensemble-mean differences we compute the Student's *t*-statistic, *t*, using the difference of means test (Von Storch and Zwiers, 1999),

$$t = \frac{|\bar{x} - \bar{y}|}{s_p \times \sqrt{\frac{2}{N}}} \quad (1)$$

153

154 where x is the ensemble-mean from PERT (or PERT*2), y is the ensemble-mean from CTRL, N is
 155 the ensemble size and s_p is the pooled standard deviation, given by,

156

$$s_p = \sqrt{\frac{\sum_{i=1}^n (x_i - \bar{x})^2 + \sum_{i=1}^m (y_i - \bar{y})^2}{n + m - 2}} \quad (2)$$

158

159 where x_i is an individual ensemble member from PERT (or PERT*2), y_i is an individual ensemble
 160 member from CTRL and n and m are the respective ensemble sizes (in our cases $N=n=m$). The
 161 ensemble-mean difference is considered statistically significant when $t \geq t_c$ where t_c is the cutoff
 162 value of the Student's t-distribution for a two-tailed probability of 0.025 (i.e., 95% confidence
 163 interval) and $n+m-2$ degrees of freedom.

164

165 To calculate the minimum ensemble-size required to detect a statistically significant ensemble-
 166 mean difference, N_{min} , we re-write (1) replacing t with t_c and N with N_{min} ,

167

$$t_c = \frac{|\bar{x} - \bar{y}|}{s_p \times \sqrt{\frac{2}{N_{min}}}} \quad (3)$$

169

170 and re-arrange to give,

171

$$N_{min} = 2t_c^2 \times \left(\frac{s_p}{\bar{x} - \bar{y}} \right)^2 \quad (4)$$

173

174 It can be seen from combining (1) and (3) that when $t \geq t_c$ then $N_{min} \leq N$. N_{min} can be considered a
175 measure of the signal-to-noise ratio, with small values of N_{min} implying a large signal-to-noise ratio
176 and large values of N_{min} implying a small signal-to-noise ratio. A similar approach for computing
177 N_{min} was used in Deser et al. (2012) and Terray et al. (2012). Equation (4) assumes that s_p is
178 insensitive to the ensemble size (i.e., s_p for N_{min} is equal to s_p for N). This assumption is
179 approximately valid, except for small values of N_{min} (when s_p for N_{min} is generally lower than s_p for
180 N ; not shown). When N_{min} is small however, the denominator in (4) is appreciably larger than the
181 numerator ($s_p \ll x - y$) and thus, N_{min} is relatively insensitive to discrepancies in s_p .

182

183 In this manuscript we focus on autumn (September-November; SON) and winter (December-
184 February; DJF) as the atmospheric response to sea-ice loss is largest in these two seasons (e.g.,
185 Deser et al., 2010; Porter et al., 2012; Screen et al., 2013) and five key atmospheric variables: near-
186 surface (defined as 1.5 m in the UM and 2 m in CAM) air temperature (T_{ref}), air temperature on
187 constant pressure levels (T), precipitation (P), sea level pressure (SLP) and geopotential height on
188 constant pressure levels (Z).

189

190 **Results**

191

192 Figure 1a shows the SON *SIC* differences (PERT-CTRL) in the single-perturbation experiment. *SIC*
193 is reduced over most of the Arctic marginal seas, with the greatest losses in the Beaufort, Chukchi
194 and East Siberian Seas. By design, this pattern closely matches the *SIC* trends observed over the
195 period 1979-2009. In DJF, *SIC* reductions are most pronounced over the Barents Sea, Sea of
196 Okhotsk, Hudson Bay and the Labrador Sea (Figure 1b). Small *SIC* increases are located along the
197 east coast of Greenland and south of the Bering Strait. The difference in sea-ice area between
198 CTRL and PERT is 1.73 and 0.98 million km² in SON and DJF, respectively, and between CTRL

199 and PERT*2 is 2.53 and 1.48 million km², respectively, in SON and DJF. Note that the loss of sea-
 200 ice area in the double-perturbation experiment is less than twice that in the single-perturbation
 201 experiment because the *SIC* in any grid-box cannot be lower than zero. Recall that the boundary
 202 conditions in PERT are based on *SIC* trends from 1979 through to 2009, which was the last full
 203 year of *SIC* data when the model runs were initiated. The past 3 years (2010, 2011 and 2012) have
 204 had low sea-ice coverage, with summer 2012 a new record minimum (Zhang et al., 2013; Parkinson
 205 and Comiso, 2013), enhancing the long-term trend. The observed sea-ice area loss from 1979 to
 206 2012, based on the National Snow and Ice Data Center (NSIDC) sea-ice index
 207 (<http://nsidc.org/data/G02135>), is 2.40 and 1.34 million km² in SON and DJF, respectively. Thus,
 208 the single-forcing experiment represents a smaller (by 28% and 27% in SON and DJF) loss of sea-
 209 ice than observed from 1979 to 2012 and the double-forcing experiment represents a slightly larger
 210 (by 5% and 10% in SON and DJF) loss of sea-ice than observed from 1979 to 2012. Figure 1c and
 211 d show the corresponding differences in *SST* for SON and DJF, respectively. In general, the *SST*
 212 warms where *SIC* decreases, and vice versa. By design, *SST* is unchanged in regions of constant or
 213 zero *SIC* change. The *SIC* and *SST* differences in the double-perturbation experiment have the same
 214 spatial patterns as in Figure 1, but with differences that are larger in magnitude (not shown).

215
 216 Figure 2 shows the ensemble-mean T_{ref} responses (a-c; g-i) and associated values of N_{min} (d-f; j-l),
 217 with the panels arranged as follows. The first (a-c) and second (d-f) rows correspond to SON and
 218 the third (g-i) and fourth (j-l) rows to DJF. The first (a, d, g, j) and second (b, e, h, k) columns are
 219 for the single-perturbation experiment in the CAM and UM, respectively, and the third column (c, f,
 220 i, l) is for the double-perturbation experiment.

221
 222 In SON, both models show widespread and significant warming over the Arctic Ocean and adjacent
 223 continents (Figure 2a, b). Unsurprisingly, warming is largest over the regions of greatest ice loss
 224 (cf. Figure 1a, c). The models are in very close agreement. The most obvious difference is that the

225 warming extends further over Scandinavia and northeastern Russia in the CAM than UM. The DJF
226 responses in both models show four warming centres: the Barents Sea, Hudson Bay, northern
227 Bering Sea and the Sea of Okhotsk (Figure 2g, h). These regions correspond to areas of winter sea-
228 ice loss and associated *SST* warming (cf. Figure 1b, d). The atmospheric warming is largely
229 confined to maritime regions in the case of the Bering Sea and the Sea of Okhotsk, but spreads to
230 neighbouring land masses around the Barents Sea and Hudson Bay. Farther away from the regions
231 of sea-ice loss, there are very few areas of significant T_{ref} response in either model. The UM depicts
232 significant cooling over the Caspian Sea and CAM depicts warming over central Asia.

233

234 As might be expected, the local T_{ref} response is larger in the double-perturbation experiment than in
235 the single-perturbation experiment (cf. Figures 2b and c; h and i). Additionally, the T_{ref} response is
236 larger over the high-northern continents and significant T_{ref} responses are detectable at lower
237 latitudes. This suggests that if Arctic sea-ice loss continues unabated, the geographical area affected
238 by sea-ice loss induced warming will increase. Under doubled forcing, there is a weak cooling
239 response over mid-latitude Eurasia in DJF, but this is only significant over a limited area
240 surrounding the Caspian Sea (Figure 2i).

241

242 In terms of N_{min} , five or fewer ensemble members are required to detect a statistically significant T_{ref}
243 response in the proximity of sea-ice loss, irrespective of the model, season or the magnitude of
244 forcing. Away from the regions of ice loss, approximately 30-50 ensemble members are required to
245 detect a significant response. The response over regions adjacent to ice loss is likely mediated by
246 horizontal advection due to synoptic systems (Deser et al., 2010), so it follows that the non-local
247 response will be weaker and subject to larger AIV (i.e., lower signal-to-noise ratio, higher N_{min})
248 than the local response that is directly driven by surface heat flux changes (e.g., Deser et al., 2010;
249 Screen et al., 2013). Under doubled forcing, the T_{ref} response over the high-northern continents is
250 easier to detect (lower N_{min} than with single forcing) and is detectable further south (cf. Figures 2e

251 and f; k and l).

252

253 Figure 3 shows the latitudinal and vertical structure of the zonal-mean T response. Consistent with
254 earlier work (Screen et al., 2012; 2013), Arctic warming due to observed sea-ice loss is strongest in
255 the lowermost atmosphere and is almost entirely confined to below 700 hPa in both seasons and
256 models (Figure 3a-b, g-h). In the double-perturbation experiment, the T response is stronger in the
257 near-surface levels, but the response remains trapped in the lower troposphere (Figure 3c, i). This
258 implies that Arctic sea-ice loss has no discernable influence on T aloft. N_{min} generally increases with
259 altitude. At the few locations where there is significant zonal-mean warming above 700 hPa, N_{min} is
260 50 or more.

261

262 Figure 4 shows the ensemble-mean P responses, arranged as in Figure 2. In SON, widespread P
263 increases are found over the Beaufort, Chukchi and East Siberia Seas in the UM (Figure 4b). In the
264 CAM, P also increases in these regions but with less spatial coherence (Figure 4a). In both models,
265 the P increases are associated with significant increases in cloud cover, principally low cloud, but
266 the cloud responses are weaker in the CAM than UM (not shown). The weaker P and cloud cover
267 responses in the CAM versus UM was previously noted by Screen et al. (2013) and appears to relate
268 to problems with the cloud cover scheme in CAM version 3. In the double-forcing experiment, the
269 P increases are stronger and are significant over most of the Arctic Ocean (Figure 4c). Away from
270 the Arctic Ocean there are isolated patches of significant P response in both models, but no large-
271 scale features even in the double-perturbation experiment. In DJF, both models show significant P
272 increases over the regions of winter sea-ice loss: the Barents Sea, Sea of Okhotsk and Hudson Bay
273 (Figures 4g, h). Remote P decreases occur over the North Atlantic and Pacific in CAM, but they
274 are only significant in small areas (Figure 4g). In the double-forcing experiment, a similar spatial
275 pattern is found, but with increased magnitude (Figure 4i). Over regions of maximum sea-ice loss,
276 N_{min} for P is less than 10 and over other regions of sea-ice loss it is around 10-30 (Figures 4d-f, j-l).

277 The majority of grid-boxes with a significant P response have an associated N_{min} of less than 40,
278 with the main exceptions being the sporadic remote P responses.

279

280 Figure 5 shows the SLP responses. In SON, SLP decreases significantly over the Beaufort, Chukchi
281 and East Siberian Seas in the UM (Figure 5b). A second low SLP centre is located over the Baltic
282 countries. In CAM, two regions of lowered SLP are identified in broadly similar, but non-identical,
283 locations (Figure 5a). The first low is shifted to the southwest to be centred over Alaska and the
284 second low is shifted westward to be located over Scandinavia. Away from these limited regions,
285 the SLP response is statistically insignificant in both models. The spatial patterns of the SLP
286 responses are largely consistent between the single- and double-perturbation experiments, but there
287 are differences in the magnitudes and significance of the responses (cf. Figure 5b and c). In the
288 double-perturbation experiment, significantly lowered SLP is found over a larger area, including
289 most of the Arctic Ocean, the Canadian Archipelago and Hudson Bay. The low-pressure centre over
290 the Baltic countries that is significant in the single-perturbation is statistically insignificant in the
291 double-perturbation experiment. Conversely, SLP increases over Europe and East Asia become
292 significant in the double-perturbation experiment.

293

294 In DJF, significant large-scale SLP decreases are found over the Arctic Ocean, Hudson Bay and
295 eastern Canada in the UM (Figure 5h). Isolated regions of significant SLP reductions are also
296 identified over the Sea of Okhotsk and central North America. SLP is increased over Europe, but
297 this feature is not statistically significant. In the CAM, SLP decreases significantly over Hudson
298 Bay, Greenland and the Atlantic-side of the Arctic Ocean (Figure 5g). SLP increases significantly
299 over the Bering Sea. Whilst the two models exhibit broadly similar SLP responses in the Arctic and
300 sub-Arctic Canada, the mid-latitude responses are rather different (cf. Figure 5g and h). In
301 particular, CAM depicts larger SLP increases over the north Atlantic and north Pacific than does the
302 UM, and the responses over the United States are opposite in sign between the two models.

303 However, the mid-latitude responses are predominantly statistically insignificant in both models, so
304 these discrepancies can be explained by AIV. No regions show significant responses of opposite
305 sign between the models. The spatial patterns of the DJF *SLP* responses are similar in the single-
306 and double-perturbation experiments (cf. Figures 5h and i). The *SLP* decrease over the Arctic and
307 Hudson Bay is larger in magnitude in the latter, but the geographical extent of the significant *SLP*
308 response is not overly different. Three small regions show significant responses in the double-
309 perturbation experiment that are not significant in the single-perturbation experiment. These are
310 *SLP* increases over the Bering Sea, eastern Europe and eastern China. The region of weak, but
311 significant, *SLP* decrease over central North America in the single-perturbation experiment is not
312 significant in the double-perturbation experiment.

313

314 N_{min} for the *SLP* response is as low as 10 in the UM over regions of maximum ice loss, especially in
315 the double-perturbation case, but N_{min} values this low are only found in very limited geographical
316 regions (Figure 5d-f; j-l). Generally, approximately 30-50 ensemble members are required to detect
317 a significant *SLP* response, and upwards of 50 members are required to detect a significant response
318 in remote regions. It is notable that even with 100 ensemble members in the UM, very few mid-
319 latitude regions show a significant *SLP* response in the single-perturbation experiment. Further,
320 despite larger mid-latitude responses in CAM, an ensemble size of 60 is insufficient for these
321 achieve statistical significance. This implies that the remote *SLP* response to recent Arctic sea-ice
322 loss is considerably smaller than AIV.

323

324 Figure 6 shows the zonal-mean *Z* responses. In SON, the high-latitude response is baroclinic with *Z*
325 decreases in the lowermost atmosphere and *Z* increases aloft. Significant zonal-mean *Z* responses
326 are only found at 1000 hPa. The vertical profile is fairly consistent across the models and
327 experiments. Taken together, the *SLP* and *Z* responses in SON are suggestive of a shallow thermal
328 (heat) low in response to sea-ice loss. Thermal lows can occur when cold air overlies warmer water,

329 as is the case in regions of sea-ice loss (Higgins and Cassano, 2009; Deser et al., 2010; Strey et al.,
330 2010; Orsolini et al., 2012). In DJF, the vertical profile of the Z response is completely different.
331 Both models show a quasi-barotropic Z decrease over high northern latitudes. This high-latitude Z
332 decrease is significant in the UM below 500 hPa, but only at 1000 hPa in CAM. Both models show
333 Z increases over mid-latitudes. In CAM, these extend throughout the troposphere, but are only
334 significant above 700 hPa. In the UM, Z increases are found aloft but not at 1000 hPa, and are
335 shifted polewards in comparison to those in CAM. They are insignificant in the single-perturbation
336 experiment, but significant above 850 hPa in the double-perturbation experiment. In all other
337 respects, the Z responses in the single- and double-perturbation are very similar. N_{min} for Z is high,
338 typically 50 or above in the single-perturbation experiment and only slightly lower in the double-
339 perturbation experiment.

340
341 In summary, the SLP and Z responses point to rather different spatial and vertical structures to the
342 circulation responses in SON and DJF. In SON, the response is baroclinic (restricted to the near-
343 surface levels) and localised. Similar local circulation responses to sea-ice loss have been identified
344 in other simulations (Higgins and Cassano, 2009; Deser et al., 2010; Strey et al., 2010; Orsolini et
345 al., 2012). By contrast in DJF, the circulation response is fairly barotropic and more spatially
346 extensive. This seasonal transition from a local baroclinic response to a larger-scale barotropic
347 response was also noted by Deser et al. (2010), although the horizontal structure of their winter
348 responses are rather different to that found here. In our CAM simulations, the DJF responses project
349 onto the positive phase of the Arctic Oscillation (AO). This is in contrast to the negative-type AO
350 responses found in February by Deser et al. (2010) and in DJF by Liu et al. (2012), both using
351 CAM but in response to projected future and past sea-ice trends, respectively. Screen et al. (2013)
352 reported a negative North Atlantic Oscillation (NAO) response in early-winter (November-
353 December) in the CAM and UM, but cautioned that the response was weak and often exceeded by
354 AIV. The larger ensembles presented here do not support a shift towards to negative phase of the

355 NAO in response to observed sea-ice loss. Instead, in CAM the response projects onto the positive
356 NAO phase and in the UM the response is not NAO-like. Thus, the wintertime circulation responses
357 (and their interactions with the large-scale modes of atmospheric variability) are not robust across
358 simulations, even those using the same models.

359

360 So far we have considered a limited number of atmospheric variables. For a wider perspective,
361 Table 1 provides the mean N_{min} for a broad selection of atmospheric variables. The values given in
362 Table 1 are averages of N_{min} across all grid-points that exhibit a significant response in that variable
363 (recall N_{min} is undefined where the response is insignificant), all experiments, models (UM and
364 CAM) and seasons (SON and DJF). For example, the mean N_{min} for T_{ref} is the average of all the
365 values in Figures 2d-f and j-l. Table 1 also provides the mean percentage area of northern
366 hemisphere extra tropics ($>30^{\circ}\text{N}$) exhibiting a significant response in each variable. Both the mean
367 N_{min} and area metrics mask substantial spatial, seasonal and inter-model variability, so the precise
368 numbers must be interpreted with caution. However, comparison of the mean N_{min} between
369 variables is insightful as it clearly demonstrates that the responses in certain variables are easier to
370 detect than others. To aid interpretation, the variables in Table 1 are listed in order of ascending
371 mean N_{min} . Recall, smaller values indicate that the response is easier to detect than larger values.
372 The ranked variables can be split into four categories of increasing mean N_{min} . This ranking is
373 largely insensitive to whether or not the double-perturbation experiment is included in the analysis
374 (not shown). The variables with smallest values ($N_{min} < 30$) are the surface heat fluxes and T_{ref} . The
375 next group ($30 < N_{min} < 50$) includes variables related to clouds, precipitation and radiation. A third
376 group ($50 < N_{min} < 60$) contains variables related to surface atmospheric circulation, including SLP
377 and near-surface wind. The hardest responses to detect ($N_{min} > 60$) are in upper-level variables, for
378 example, mid-tropospheric (500 hPa) temperature (T_{500}) and geopotential height (Z_{500}) and
379 jetstream-level (250 hPa) wind (U_{250} , V_{250}). Although we have not considered stratospheric
380 variables here, Cai et al. (2012) found that the stratospheric response to sea-ice loss is small

381 compared to the tropospheric response.

382

383 Clearly, AIV is a key source of uncertainty in the simulated atmospheric response to Arctic sea-ice
384 loss. Larger ensembles can reduce this uncertainty by averaging out, to some extent, the effects of
385 AIV. Figure 7 quantifies the reduction in uncertainty in the response to Arctic sea-ice loss, due to
386 AIV, as the ensemble size increases. To construct this figure, we have sub-sampled our large
387 ensembles into smaller sub-ensembles of varying size. For each sub-ensemble size, a large number
388 (100,000) of unique combinations are sampled to produce a large set of sub-ensemble mean
389 responses. For example, for a sub-ensemble size of 5 we sub-sampled 100,000 unique combinations
390 of 5 members from the full set. For each combination, we averaged the selected members to
391 produce a sub-ensemble mean. This results in a set of 100,000 sub-ensemble mean responses. The
392 spread (difference between maximum and minimum values) of these sub-ensemble mean responses
393 provides a measure of the uncertainty in the response due to AIV, for an ensemble of that size.
394 Figures 7a and b shows examples for the Arctic-mean ($>70^{\circ}\text{N}$) SON T_{ref} response and DJF SLP
395 response, respectively, but qualitatively similar results are found for other seasons and variables.

396

397 Uncertainty due to AIV, as estimated by the spread of sub-ensemble mean responses, can be seen to
398 decrease almost exponentially as the ensemble size increases. This implies that to reduce
399 uncertainty by one half, the ensemble size has to be doubled. In absolute terms, uncertainty due to
400 AIV decreases rapidly as the ensemble-size increases from 5 (or fewer) to 20 members, and then
401 continues to reduce more slowly as further ensemble members are added. This behaviour is very
402 similar in the two models and in both the single- and double-perturbation experiments, however,
403 CAM has larger AIV than the UM for both T_{ref} and SLP . In the UM a point is reached, around 50-
404 60 ensemble members, where adding further ensemble members has almost no impact on the
405 uncertainty due to AIV (spread of responses). We assume a similar point would occur in the CAM,
406 but we have insufficient ensemble members to confirm this.

407

408 **Discussion and Conclusions**

409

410 Arguably, one of the most surprising aspects of our results is that over the Arctic we have identified
411 a robust (in the sense that it is statistically significant in both models, and in both the single- and
412 double-perturbation experiments) lowering of *SLP* in response to Arctic sea-ice loss. This is
413 surprising because this high-latitude DJF *SLP* response is opposite to that suggested in some
414 empirical studies (e.g., Francis et al., 2009; Jaiser et al., 2012) and to that found in other modelling
415 studies, for example, Deser et al (2010) and Liu et al. (2012). It is of especial interest to draw
416 comparisons with Liu et al. because that study performed very similar experiments to those
417 presented here. It used the same model (CAM version 3 at T42 resolution), experimental set-up and
418 forcing based on observed sea-ice trends. However, Liu et al. report a significant increase in DJF
419 *SLP* in response to Arctic sea-ice loss, in stark contrast to the decrease shown here (cf. our Figure
420 3g and their Figure 4c). One notable difference between the two studies is that Liu et al. used an
421 ensemble of 20 CAM simulations compared to our ensemble of 60 CAM simulations.

422

423 It is plausible that 20 ensemble members are insufficient to accurately separate the forced signal
424 from AIV (N_{min} for DJF *SLP* is generally larger than 20). To test this hypothesis, we considered if it
425 is possible to derive the Liu et al. result from a subset of our larger CAM ensemble. Figure 8 shows
426 probability distribution functions (PDFs) for the Arctic-mean DJF sub-ensemble mean *SLP*
427 responses (i.e., the set of 100,000 sub-sampled responses). The PDFs narrow as the ensemble size
428 increases, implying that larger ensembles yield more precise responses to Arctic sea-ice loss. For a
429 sub-ensemble size of 10, it is possible to obtain both positive and negative sub-ensemble mean *SLP*
430 responses. This shows that AIV influences both the magnitude and sign of *SLP* responses in small
431 ensembles. However for a sub-ensemble size of 20, an Arctic-mean increase in *SLP* is found in less
432 than 0.1% of cases. This implies that the Liu et al. result cannot be derived from a 20-member

subset of our larger ensemble. Thus, it is very unlikely that the discrepancy between the Arctic winter *SLP* responses in this study and in Liu et al. is due to AIV alone. A further difference between this study and Liu et al. is the magnitude of the sea-ice forcing. Although the spatial pattern of the sea-ice forcing is highly similar between the two studies, our forcing is approximately two-to-three times larger magnitude than that in Liu et al. (cf. our Figures 1a ,b and their Figures 4a, b; note the different colour scales). Whether or not this is the cause of the opposing winter *SLP* responses is unclear, but the discrepancy highlights that experimental differences between simulations, even with the same model, can lead to fundamentally different responses.

We now return to the question posed earlier: how many ensemble members are required to detect a significant response to Arctic sea-ice loss? In reality, there is no simple answer to this question as N_{min} varies considerably in space and by variable. As a general rule of thumb, we suggest that detection of the thermo-dynamical (e.g., *SLP*, wind) response requires an ensemble size approximately twice as large as the thermal response (e.g., surface heat fluxes, *T*). The hydrological response (e.g., cloud, *P*) lays in-between, which likely reflects influences of both thermal and thermo-dynamical factors on these variables. Deser et al. (2012) reported very similar findings in the context of the coupled climate response to greenhouse gas forcing, as did Wehner (2000) and Taschetto and England (2008) both in the context of the atmospheric response to global *SST* and *SIC* trends. The upper-level response to Arctic sea-ice loss (e.g., Z_{500} , T_{500} , U_{250} , V_{250}) is harder to detect than the near-surface response as the signal-to-noise ratio decreases with altitude. Equally, the remote T_{ref} or *P* responses (that are mediated by thermo-dynamical processes) are harder to detect than the local T_{ref} and *P* responses (that are primarily-driven by surface fluxes). Figure 7 suggests large gains, in terms of reduced uncertainty, by increasing from a small (less than 20 members) to moderate-sized (20-50 members) ensemble. Further increases in ensemble size represent a case of “diminishing returns” with smaller reductions in uncertainty per additional ensemble member. On this basis, we argue that an ensemble size of around 50 members is

desirable. This is considerably larger than the typical ensemble size used in past studies of the atmospheric response to observed Arctic sea-ice loss (e.g., 5 in Ghatak et al. (2012) and Orsolini et al. (2012), 5/8 in Screen et al. (2013), 10 in Strey et al. (2010), 15 in Porter et al. (2012), 20 in Liu et al. (2012)).

The values of N_{min} have implications not only for modelling studies, but also for what aspects of the simulated Arctic sea-ice response may be observable in the real world. Since each ensemble member is one-year of simulation, N_{min} can also be thought of as approximate measure of the minimum number of years required to detect a significant response due to Arctic sea-ice loss, assuming the rate of loss is linear. The differences in prescribed boundary conditions between CTRL and PERT have, in reality, occurred over a 31-year period (1979-2009). Assuming that the models are realistic in their depiction of the forced response and AIV, a N_{min} of 31 or less suggests that the simulated response should be observable in nature over the period 1979-2009. Conversely, a N_{min} of greater than 31 suggests that more than 31 years are required to separate the forced response from AIV and therefore, the response to past sea-ice loss would not be expected to be detectable in observed records. Accordingly, we argue that the simulated local T_{ref} and P responses to Arctic sea-ice loss should be detectable, but that the atmospheric circulation (e.g., SLP , 10 m wind), upper-level (e.g., Z_{500} , T_{500} , U_{250} , V_{250}) and remote responses may be partially or wholly masked by AIV. In practise, the detection and importance of the atmospheric impacts of sea-ice loss not only depend on the relative magnitudes of the sea-ice forced change compared to AIV, but also on the relative magnitudes of sea-ice forced response to other forced responses.

Figures 9a and d show observed (from ERA-Interim; Dee et al., 2011) trends in T_{ref} over the period 1979-2009 for SON and DJF, respectively. These are highly similar to the simulated T_{ref} responses to Arctic sea-ice loss (Figure 2), suggesting that the T_{ref} response is indeed detectable, consistent with previous studies (Screen and Simmonds, 2010a; 2010b; 2012). Figures 9b and e show

485 observed trends in P for SON and DJF, respectively, taken from the Global Precipitation
486 Climatology Project (GPCP) data set (Adler et al., 2003). These can be compared to the simulated P
487 responses in Figure 4. Although the moderate values of N_{min} in Figure 4 suggested that the local P
488 response may be detectable, the observed trends are not in agreement with the simulated P
489 responses. We propose that there are two likely reasons for this apparent disparity. One reason is
490 that detection of the atmospheric impacts of sea-ice loss not only depend on the relative magnitudes
491 of the sea-ice forced change compared to AIV, but also on the relative magnitudes of sea-ice forced
492 response to other forced responses. It is likely that the observed P trends are forced by factors other
493 than, or in addition to, sea-ice loss. Secondly, there is considerable uncertainty as to the sign and
494 magnitude of observed P trends over the poorly observed Arctic region. P trends from alternative
495 observationally constrained data sources – for example the CPC Merged Analysis of Precipitation
496 (CMAP; Xie and Arkin, 1997) or ERA-Interim - depict rather different patterns of P change over
497 the Arctic Ocean (not shown). It is possible therefore, that the simulated P response to sea-ice loss
498 is undetectable in observations because of considerable observational uncertainty. Observed (ERA-
499 Interim) SON and DJF SLP trends are shown in Figures 9c and f. Neither resemble the simulated
500 SLP responses in any of the models/experiments (Figure 5), which is consistent with the conclusion
501 that the SLP response to Arctic sea-ice loss is masked by AIV, or SLP trends due to other forcing
502 factors. We note that the observed trends are only one realisation (effectively one ensemble
503 member) and likely contain a sizeable component of natural (unforced) variability. Thus, the
504 observed trends would not be expected to match the ensemble-mean simulated responses.

505

506 We close by emphasising two arguably obvious, but nonetheless important, considerations. Firstly,
507 this study has only considered the first-order “direct” atmospheric response to Arctic sea-ice loss.
508 The fully coupled climate system response to Arctic sea-ice loss may be different to that shown
509 here. Secondly, Arctic sea-ice loss is only one forcing factor that may be relevant to northern
510 hemisphere climate variability and change. Observed trends reflect changes in multiple forcing

511 factors and the complex interactions between them.

512

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520

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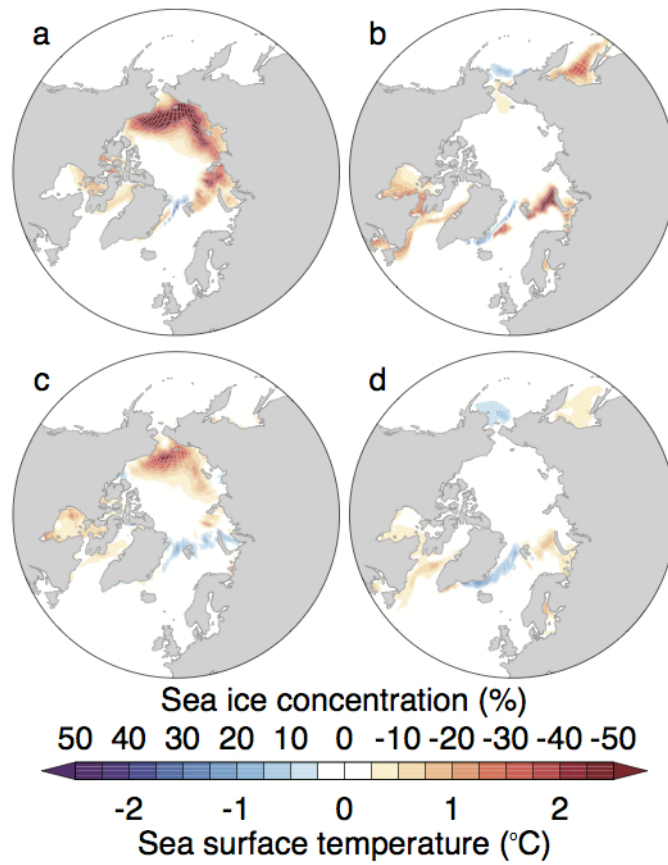
625

Variable	Mean N_{min}	Mean area
Surface sensible heat flux	21.9	19.1
Surface latent heat flux	24.5	19.7
1.5 m air temperature (T_{ref})	26.0	28.9
Low cloud cover	29.9	17.5
925 hPa air temperature (T_{925})	31.1	25.7
Net surface short-wave radiation	32.6	12.9
Net surface long-wave radiation	32.8	15.2
Total cloud cover	33.3	16.5
Precipitation	40.3	11.4
Sea level pressure (SLP)	50.3	11.9
10 m meridional wind speed	50.7	9.0
10 m zonal wind speed	50.8	9.4
500-1000 hPa thickness	56.1	7.0
250 hPa zonal wind speed (U_{250})	65.4	5.5
250-1000 hPa thickness	67.6	4.7
250 hPa meridional wind speed (V_{250})	68.0	3.2
500 hPa air temperature (T_{500})	70.0	4.3
500 hPa geopotential height (Z_{500})	71.9	5.0
250 hPa geopotential height (Z_{250})	73.3	4.8

627

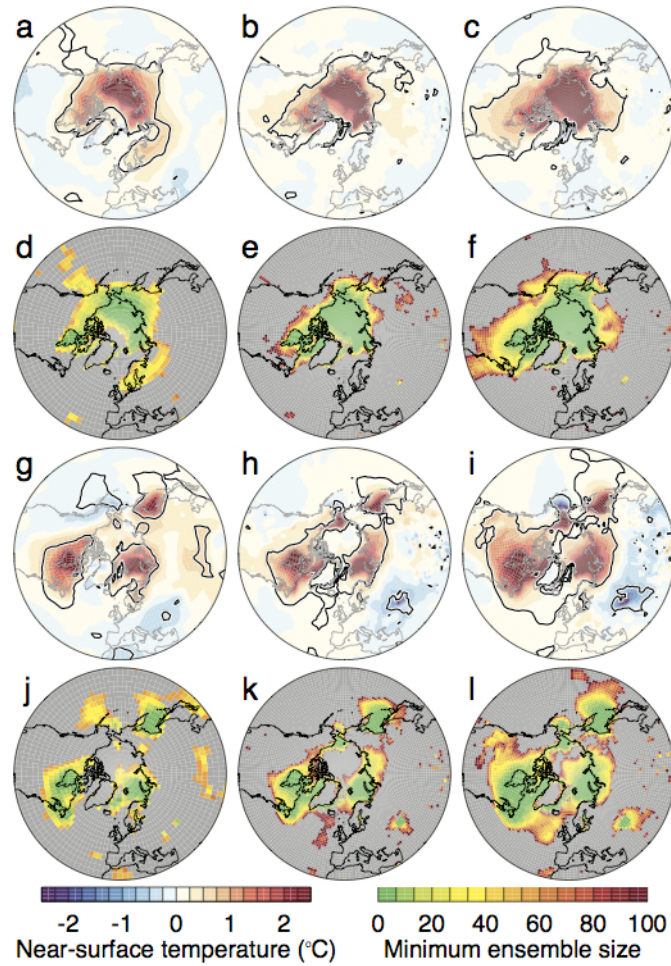
628 **Table 1:** Mean N_{min} for a selection of atmospheric variables. For each variable, the value of N_{min}
629 given is the average over all grid-boxes (with a significant response in that variable), all models and
630 experiments, and both autumn and winter. The right-hand column shows the mean percentage area
631 of northern hemisphere extra tropics ($>30^{\circ}\text{N}$) exhibiting a significant response in that variable
632 (again averaged across models, experiments and seasons). To aid interpretation, the variables are
633 listed in order of ascending mean N_{min} . Note that low values imply a response that is easier to detect
634 than high values.

635



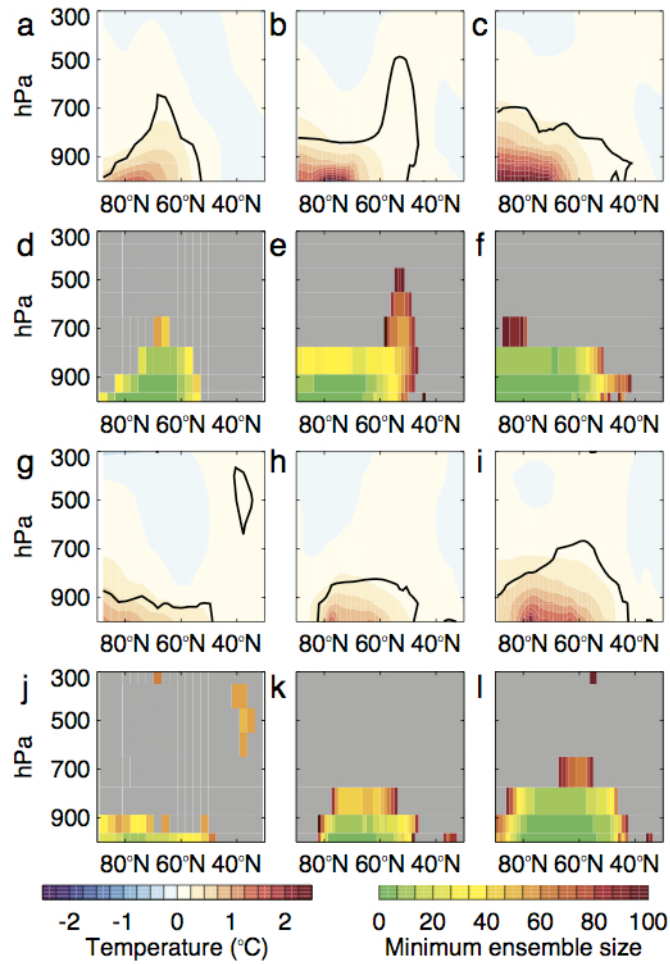
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637 **Figure 1:** Ensemble-mean differences (PERT-CTRL) in sea-ice concentration (*SIC*) for (a) autumn
 638 and (b) winter. (c-d) As (a-b), but for sea surface temperature (*SST*). Note the inverse scale for *SIC*.



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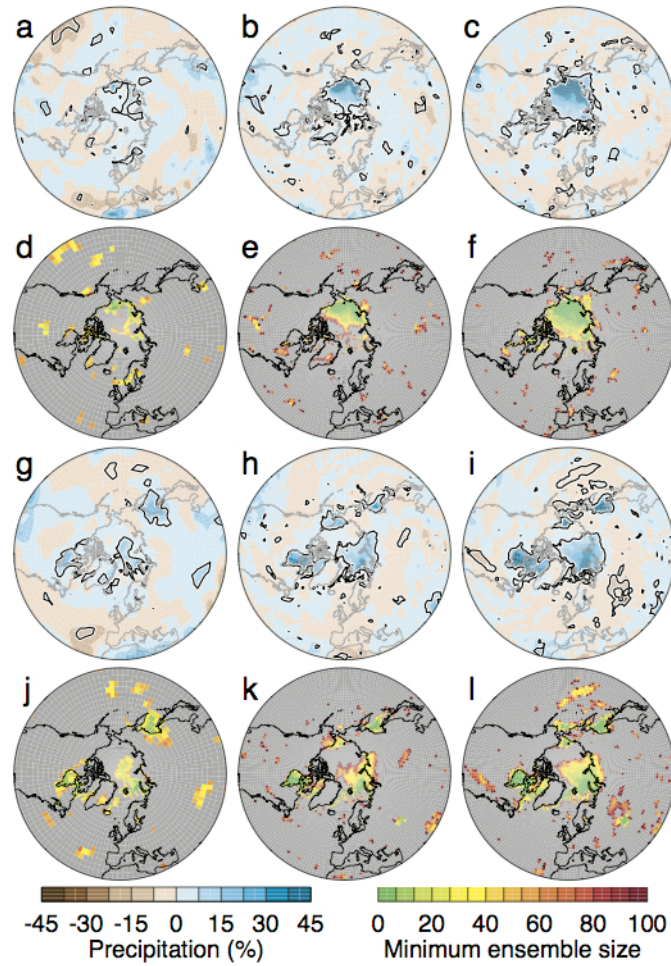
640 **Figure 2:** Ensemble-mean differences in autumn near-surface air temperature (T_{ref}) for (a) CAM
 641 PERT-CTRL, (b) UM PERT-CTRL and (c) UM PERT*2-CTRL. Statistically significant
 642 differences (at the $p \leq 0.05$ level) are enclosed by black contours. (d-f) N_{min} for the differences
 643 shown in (a-c), respectively. Grey shading denotes an insignificant ensemble-mean difference. (g-l)
 644 As (a-f), but for winter.



645

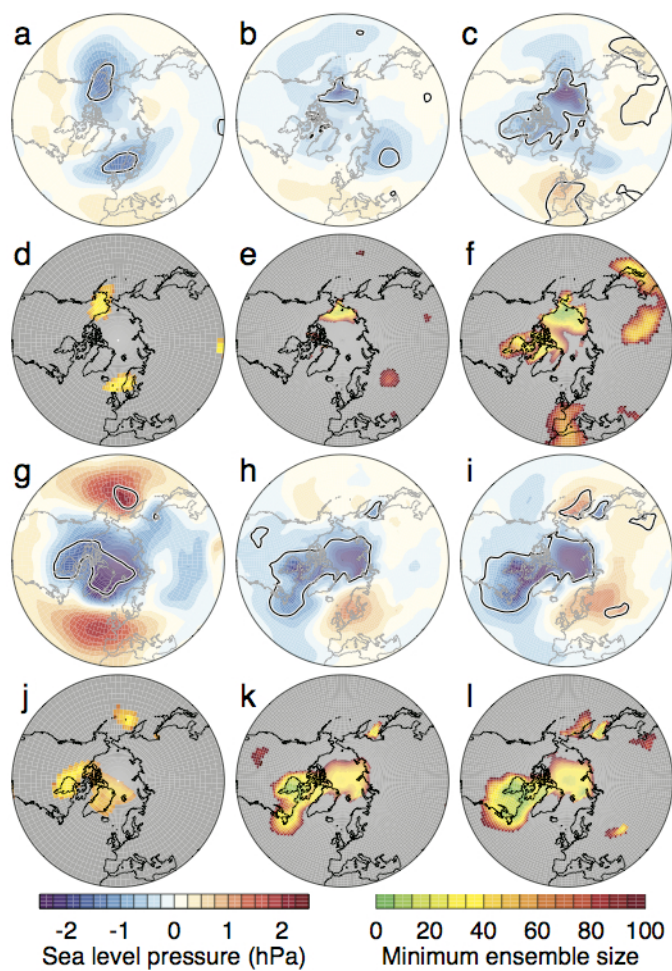
646 **Figure 3:** Zonal-mean ensemble-mean differences in autumn air temperature (T) for (a) CAM
 647 PERT-CTRL, (b) UM PERT-CTRL and (c) UM PERT*2-CTRL. Statistically significant
 648 differences are enclosed by black contours. (d-f) N_{min} for the differences shown in (a-c),
 649 respectively. Grey shading denotes an insignificant ensemble-mean difference. (g-l) As (a-f), but for
 650 winter.

651



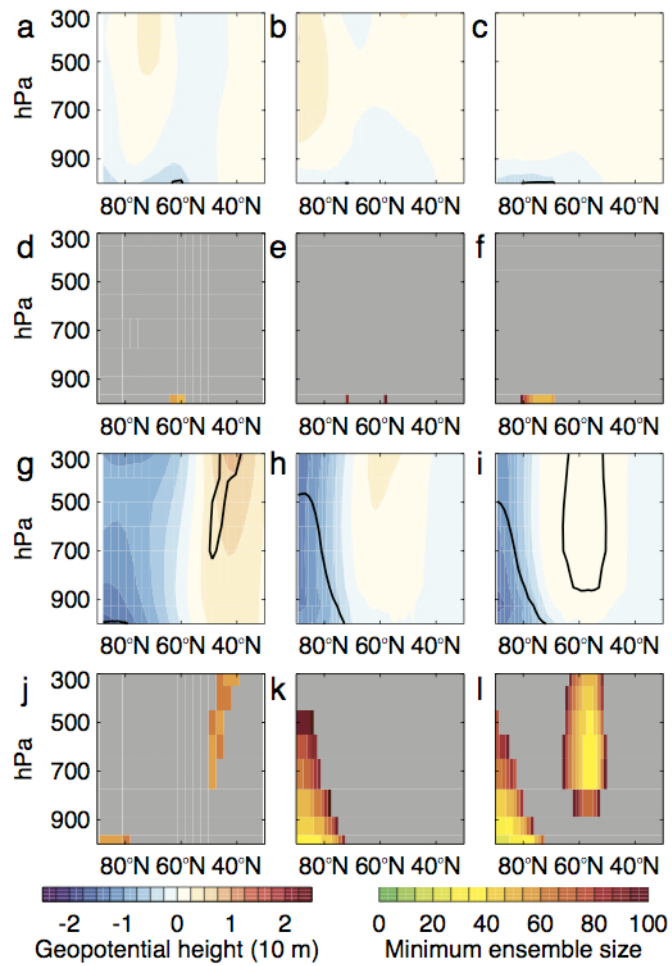
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653 **Figure 4:** As Figure 2, but for precipitation (P). The P differences are expressed as percentages
 654 relative to the ensemble-means in CTRL.



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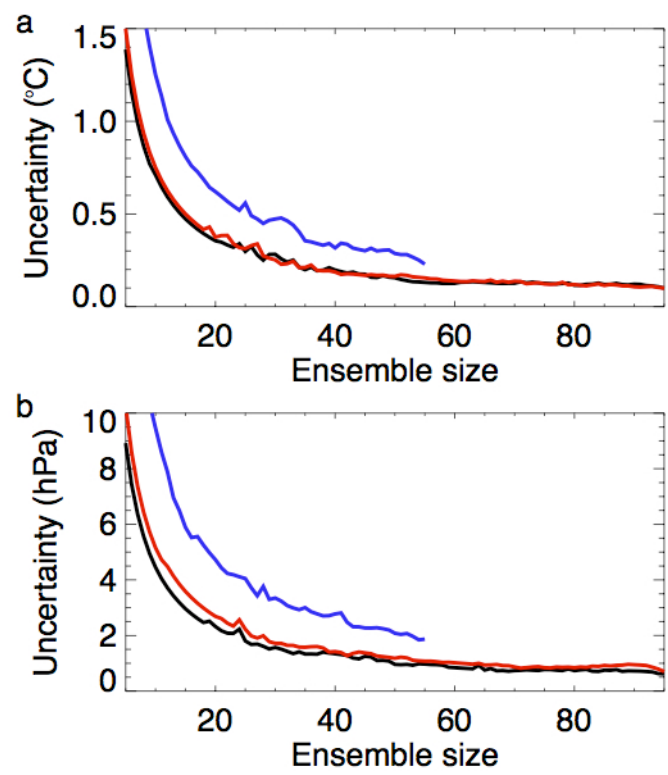
656 **Figure 5:** As Figure 2, but for sea level pressure (*SLP*).



657

658 **Figure 6:** As Figure 3, but for geopotential height (Z).

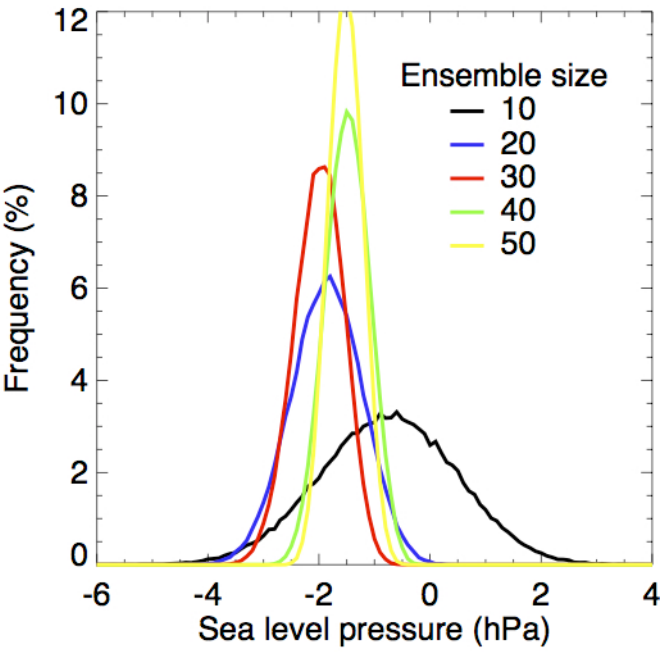
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661

662 **Figure 7:** Uncertainty due to atmospheric internal variability [see text for details] as a function of
663 ensemble size for the Arctic-mean (a) autumn near-surface air temperature (T_{ref}) response and (b)
664 winter sea level pressure (SLP) response. Black and blue lines correspond to PERT-CTRL in the
665 UM and CAM respectively, and the red lines to PERT*2-CTRL.

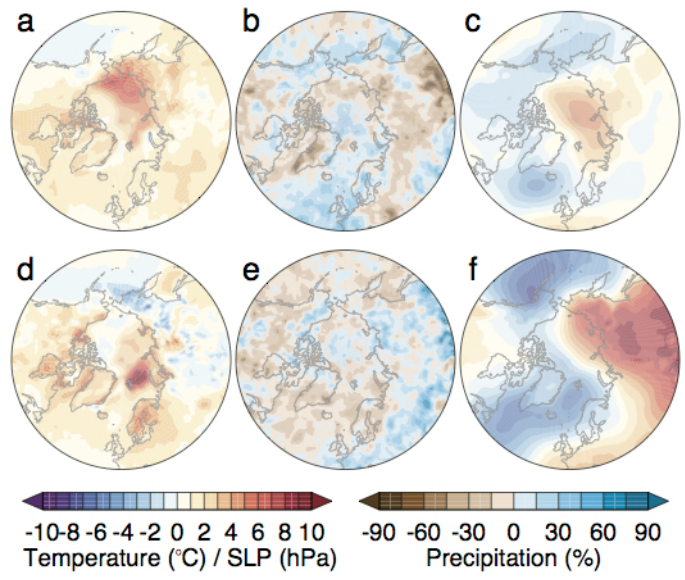
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668

669 **Figure 8:** Probability distribution functions (PDF) for the winter Arctic-mean sea level pressure
670 (*SLP*) responses in sub-ensembles of varying size. Each PDF is constructed from 100,000 unique
671 combinations sub-sampled from the 60-member CAM ensemble. For example, the blue line
672 represents sub-ensemble means for 100,000 unique combinations of 20 CAM members sampled
673 from the full set of 60 CAM members [see text for further details].

674



677 **Figure 9:** Observed trends in autumn (a) near-surface temperature (T_{ref}), (b) precipitation (P) and
678 (c) sea level pressure (SLP) for the period 1979-2009. Precipitation trends are expressed as
679 percentages relative to the climatological-means. (d-f) As (a-c), but for winter. T_{ref} and SLP data are
680 from the ERA-Interim reanalysis and P data are from the GPCP product.