

Deconstructing the climate change response of the Northern Hemisphere wintertime storm tracks

Article

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1 **Deconstructing the climate change response of the Northern**
2 **Hemisphere wintertime storm tracks**

3 **B. J. Harvey · L. C. Shaffrey · T. J. Woollings**

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6 **Abstract** There are large uncertainties in the circulation response of the atmosphere to climate change.
7 One manifestation of this is the substantial spread in projections for the extratropical storm tracks made
8 by different state-of-the-art climate models. In this study we perform a series of sensitivity experiments,
9 with the atmosphere component of a single climate model, in order to identify the causes of the differences
10 between storm track responses in different models. In particular, the Northern Hemisphere wintertime
11 storm tracks in the CMIP3 multi-model ensemble are considered. A number of potential physical drivers
12 of storm track change are identified and their influence on the storm tracks is assessed. The experimental
13 design aims to perturb the different physical drivers independently, by magnitudes representative of the
14 range of values present in the CMIP3 model runs, and this is achieved via perturbations to the sea
15 surface temperature and the sea-ice concentration forcing fields. We ask the question: can the spread of

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16 projections for the extratropical storm tracks present in the CMIP3 models be accounted for in a simple
17 way by any of the identified drivers? The results suggest that, whilst the changes in the upper-tropospheric
18 equator-to-pole temperature difference have an influence on the storm track response to climate change,
19 the large spread of projections for the extratropical storm track present in the northern North Atlantic in
20 particular is more strongly associated with changes in the lower-tropospheric equator-to-pole temperature
21 difference.

22 **Keywords** Storm tracks · Climate change · CMIP3 · Drivers of change · Polar amplification

23 1 Introduction

24 Extratropical cyclones are an important component of the climate system. They play a substantial role
25 in the general circulation of the atmosphere through the transport of heat, momentum and moisture.
26 They also play a role in the societal impacts of weather in the mid-latitudes through their associated
27 precipitation, clouds and surface winds. It is therefore important to understand how the extratropical
28 storm tracks, and the properties of extratropical cyclones, may change in the future.

29 Many studies have addressed this by either (i) analysing projections of the storm tracks simulated by
30 one or more of the latest state-of-the-art climate models, to quantify and understand the range of changes
31 that may take place, or (ii) by performing specific climate model experiments to understand individual
32 mechanisms that might drive the changes. Over recent years the consensus has arisen that the zonal-mean
33 properties of the storm tracks, and their associated eddy-driven surface winds, will move poleward under
34 the projected greenhouse gas forcing (Yin 2005; Solomon et al. 2007). The precise mechanism causing
35 such a change, however, is still under debate (Chen et al. 2008; Rivière 2011; Kidston et al. 2011). Recent
36 studies (Woollings & Blackburn 2012; Barnes & Polvani 2013) have shown that the poleward shift of the
37 eddy-driven jets in response to anthropogenic greenhouse gas forcing is remarkably robust in different

38 models, ocean basins and seasons, with the notable exception of the North Atlantic region in winter. In
39 that case the multi-model mean response is small in comparison to the spread between different model
40 responses. Other studies have investigated the responses of the storm tracks directly in large multi-model
41 ensembles (Ulbrich et al. 2008; Harvey et al. 2012; Zappa et al. 2013b; Woollings et al. 2012; Harvey
42 et al. 2013) and found less of a consensus. For the North Atlantic wintertime storm track in particular,
43 these studies suggest that, rather than a poleward shift, there will be an intensification of its southern
44 flank, effectively extending the storm track zonally towards Europe. However, there is a large spread in
45 the projections of different models (e.g. La  n   et al. 2009).

46 As an illustration of the problem, Figure 1 shows the multi-model mean climate change responses
47 of several wintertime climate variables from the World Climate Research Programme’s Coupled Model
48 Intercomparison Project phase 3 (CMIP3) multi-model dataset. Here the climate change response is
49 defined as the difference between 2081-2100 from the SRESA1B experiment and 1961-2000 from the
50 20C3M experiment. Panel **a** shows a measure of the storm activity, the standard deviation of the 2-
51 6 day bandpass-filtered daily-mean mean sea level pressure (MSLP) field. This commonly-used diagnostic
52 provides a simple assessment of synoptic-scale activity using only daily-mean MSLP data (Hoskins &
53 Hodges 2002; Chang 2009), and will be referred to simply as the storm track in the following. The
54 response of this measure of storm activity is in general agreement with the studies cited above. Here we
55 note the key points relevant to later discussion: (i) there is a poleward shift and intensification of the
56 multi-model mean North Pacific storm track in response to climate change, (ii) there is no corresponding
57 poleward shift in the North Atlantic storm track, instead there is an intensification on its southern flank
58 resulting in a zonal extension of the storm track towards Europe, and (iii) there is a large inter-model
59 standard deviation of the responses, particularly in the North Atlantic region (Figure **1b**). It is this large
60 spread that we wish to understand.

61 On the global scale, the zonal-mean temperature response to increased GHGs consists of a warming
62 throughout the troposphere with regions of enhanced warming in the tropics in the upper-troposphere and
63 in the wintertime polar regions near the surface, as shown in Figure 1c. Many studies have suggested that
64 both regions of warming may influence the storm tracks in different ways, for instance Harvey et al. (2013)
65 show using CMIP5 data that a substantial fraction of the inter-model spread in wintertime storm track
66 projections can be accounted for by a simple linear regression model based on differences in the equator-to-
67 pole temperature difference at upper- and lower- tropospheric levels, with larger temperature differences
68 at either level associated with stronger storm activity. A number of mechanisms have been suggested
69 for how the warming influences the storm tracks in each case, including: changes in the baroclinicity, via
70 changes in the meridional temperature gradient or the vertical static stability (Frierson 2006; Butler et al.
71 2010) or the height of the tropopause (Williams 2006; Lorenz & DeWeaver 2007) directly influencing the
72 eddies, or changes in the mean circulation influencing the propagation of the eddies (Hartmann et al.
73 2000; Kidston et al. 2011) or their wave-breaking behaviour (Chen et al. 2008; Rivière 2011).

74 On a more local scale, Wilson et al. (2009) and Brayshaw et al. (2011) present evidence that local
75 sea surface temperatures (SSTs) can influence the North Atlantic storm track, and Deser et al. (2000)
76 and Magnusdottir et al. (2004) show that the distribution of Arctic sea ice concentration (SIC) is also
77 important. Future changes in the SST and sea-ice distributions may be linked via changes in the Atlantic
78 Meridional Overturning Circulation (AMOC, see Laine et al. (2009); Brayshaw et al. (2009)), and indeed
79 Woollings et al. (2012) show that there is a strong correlation between the responses of the AMOC and
80 the North Atlantic storm track in the CMIP3 models. However, several questions remain: Is the inter-
81 model spread in the responses of either the North Atlantic SSTs or the Arctic sea ice responsible for the
82 large spread in the storm track responses?, and is the link between the responses of the AMOC and the
83 North Atlantic storm track responses due to changes in North Atlantic SSTs, Arctic sea-ice, or both?

84 In this paper we attempt to deconstruct the climate change projections of the Northern Hemisphere
85 storm track/eddy-driven jet system, and in particular aim to understand the causes of the inter-model
86 differences in the CMIP3 multi-model dataset. In particular we ask the question: can the differences
87 between the model projections be accounted for by differences in a small number of physical ‘drivers’
88 of the storm tracks? If so, which drivers are most important, and what is their impact on the storm
89 tracks? This work is motivated by the studies of Woollings et al. (2012) and Harvey et al. (2013), which
90 suggest respectively the AMOC and both the lower- and the upper-tropospheric large-scale equator-to-
91 pole temperature differences as possible physical drivers of storm track change. These studies relied on
92 inter-model correlations in multi-model ensembles and were therefore unable to infer any causality. In the
93 present study, a series of atmosphere-only GCM experiments are presented, using a single model forced
94 by perturbations to the SST and SIC fields. The perturbations are designed to isolate the effects of the
95 drivers of storm track change, and reflect the spread of these drivers between the different CMIP3 models
96 projections. In this way, the dominant drivers of the inter-model spread in the storm track response are
97 determined. Whilst we restrict attention to the CMIP3 models in the present study, we note that both the
98 magnitude of the mean storm track response in the CMIP5 models and the inter-model spread between
99 the CMIP5 models have similar magnitudes and spatial distributions as the CMIP3 models (see Harvey
100 et al. 2013, Figure 2b), and therefore the results of this study are most likely also relevant to the CMIP5
101 ensemble.

102 The experimental design is described in Section 2, followed by a description of the control experiments
103 in Section 3. Two sets of perturbation experiments are presented: the first aims to test the hypothesis
104 discussed in Harvey et al. (2013) that the large-scale equator-to-pole temperature difference at upper-
105 and lower-tropospheric levels influences the storm track, and these are presented in Section 4. The second

106 set of perturbation experiments consider the North Atlantic in more detail and these are presented in
107 Section 5. A summary and discussion are presented in Section 6.

108 **2 Model and Experimental Design**

109 **2.1 HadGAM1**

110 The model used in this study is HadGAM1, the atmospheric component of the Hadley Centre’s HadGEM1
111 global climate model (Johns et al. 2006; Martin et al. 2006) which was used in the CMIP3 ensemble. It
112 has a grid-point resolution of $1.875^\circ \times 1.25^\circ$, and 38 layers between the surface and the model top at 39
113 km altitude. **Around 10 of the layers are globally above the tropopause.**

114 The experiments differ only in the prescribed SST, SIC and well-mixed greenhouse gas (GHG) fields
115 (the land surface is allowed to evolve). The GHGs are kept fixed during each experiment and are specified
116 using either late 20th century or projected late 21st century values from the SRESA1B scenario as
117 appropriate. Annually-repeating monthly SST and SIC fields are constructed for each experiment from
118 CMIP3 output as described in Section 2.2. All other elements of the model setup are unchanged including
119 the land-use specifications, and the aerosol and ozone concentrations.

120 Each experiment consists of two 21 year runs (single 21 year runs were found to be insufficient to
121 provide robust statistics), all starting from the same pair of initial conditions and all having the first
122 12 months discarded as spin-up. Therefore there are 40 complete years of data for each experiment.
123 Additional runs of 21 years are performed for the two control experiments, as described below, and in
124 those cases 60 years of data are used.

125 2.2 Experimental design

126 Table 1 details the experiments presented in this paper, and the relationships between them. There
127 are two control runs (CON-20C and CON-A1B) representing the 20C3M and SRESA1B mean climates
128 respectively, and four pairs of perturbation runs (ARC \pm , UFM \pm , UFM3 \pm and NATL \pm) designed to
129 isolate different drivers of storm track change. Various methods have been used to construct the SST
130 and SIC fields for the perturbation experiments, as described below. In each case the aim is to capture
131 one standard deviation of the spread in the CMIP3 model responses of a particular driver of storm track
132 change. **In this way, the relative contributions of the different drivers to the overall spread of the CMIP3**
133 **ensemble are assessed.**

134 Data from 21 CMIP3 models is used to construct the SST and SIC fields, as listed in Appendix A.
135 One ensemble member per model is used in order to provide equal model weighting in the multi-model
136 mean. Data from two periods is used: years 1961-2000 of the 20C3M scenario and years 2061-2100 of the
137 SRESA1B scenario. Monthly SST and SIC climatologies are calculated for both 40 year periods of each
138 model, and these are interpolated onto the HadGAM1 grid. Nearest grid point extrapolation is used to
139 fill gaps resulting from land-sea mask mismatches between the models.

140 2.2.1 Control experiments (CON-20C and CON-A1B)

141 The CON-20C and CON-A1B experiments represent the CMIP3 20C3M and SRESA1B climates respec-
142 tively. The SST fields used are the multi-model mean values from the 20C3M and SRESA1B experiments
143 respectively. A similar method is desired for creating monthly SIC fields; however, due to the large range
144 of 20C3M ice extents in the CMIP3 models (see, e.g., Stroeve et al. 2007), simply taking the multi-model
145 mean of the SIC values produces an ice distribution with an unrealistically smooth ice edge region. To
146 avoid this problem, artificial SIC distributions have been created to mimic the typical ice distribution

147 of the CMIP3 models in the 20C3M and SRESA1B scenarios whilst retaining a realistic ice-edge struc-
148 ture. The details of this procedure are presented in Appendix B. In brief, a multi-model mean ice edge
149 position is constructed from the 20C3M ice distributions and then converted into an ice distribution
150 which transitions smoothly from 100% within the ice edge to 0% outside of it. This distribution is used
151 for the CON-20C experiment. The CON-20C ice edge position is then retreated by a distance which is
152 representative of the ice retreats in the models, and then similar converted into a smooth ice distribution
153 which is used for the CON-A1B experiment. This process is carried out separately in the Northern and
154 Southern Hemispheres.

155 Figures 2a and c show the resulting DJF mean SST and SIC values for the CON-20C experiment as
156 well as the CON-A1B minus CON-20C difference. The SST increases almost everywhere between CON-
157 20C and CON-A1B, with a peak around 2.3 deg C in the equatorial regions. However, there are spatial
158 variations and in particular there is a small region of cooling in the west side of the North-Atlantic sub-
159 polar gyre, surrounded by a region of mediated warming covering much of the sub-polar gyre. This region
160 of cooling has been linked to a slow down of the AMOC in the future, associated with changes in wind
161 and precipitation in the regions of deep oceanic convection (Drijfhout et al. 2012). The SIC decreases
162 between CON-20C and CON-A1B near the ice edge at all longitudes, with the largest changes in the
163 Barents Sea and the Bering Strait.

164 2.2.2 Global perturbation experiments (ARC, UFM and UFM3)

165 Two pairs of perturbation experiments, $ARC\pm$ and $UFM\pm$, are configured to reflect the spread in the
166 lower- and upper-tropospheric equator-to-pole temperature differences respectively. In the $ARC\pm$ exper-
167 iments a perturbation is made to the Arctic sea-ice distributions, as described below. A sea ice reduction
168 generally acts to warm the lower atmosphere in the Arctic, and therefore reduces the lower-tropospheric

169 equator-to-pole temperature gradient. The aim here is to modify the lower-tropospheric equator-to-pole
170 temperature difference by approximately one standard deviation of the spread of the responses of the
171 lower-tropospheric equator-to-pole temperature differences in the CMIP3 models. The Antarctic sea-ice
172 extent is not changed as the focus here is on the Northern Hemisphere circulation. The method for
173 perturbing the Arctic sea-ice extent is similar to that used to generate the CON-A1B ice distribution.
174 The distance retreated for the ARC+ and ARC- experiments is equal to the distance retreated for the
175 CON-A1B experiment plus and minus one standard deviation of the range of distances retreated by the
176 ice edges in the CMIP3 models (see Appendix B for details). These experiments are therefore designed
177 to capture the spread in the ice edge responses present in the CMIP3 models. The difference between the
178 ARC+ and ARC- DJF mean SIC fields are shown in Figure 2d. In addition, the SST field is perturbed in
179 the ice edge region as described below in section 2.2.3. The SST perturbations act to fill-in the region of
180 exposed sea surface in these experiments, whilst also modifying the surrounding SST in expected sense,
181 with ARC+ having a warmer ice edge region than A1B, and ARC- a cooler ice edge region. We do not
182 attempt to separate the relative influence of the SST and ice perturbations as it is their combined effect
183 that is of interest to this study.

184 Turning to the UFM \pm experiments, Harvey et al. (2013) show that the upper-tropospheric equator-
185 to-pole temperature difference is closely related to the tropical SSTs in the CMIP5 models. Therefore, to
186 modify the upper-tropospheric equator-to-pole temperature difference, a uniform SST anomaly ΔT_{UFM}
187 is applied globally to the CON-A1B SST field. The magnitude of ΔT_{UFM} is chosen initially as the inter-
188 model standard deviation of the tropical SST responses in the CMIP3 models, where the Tropics are
189 defined here as the region with $|\text{latitude}| < 30$ degrees. The value of this is $\Delta T_{\text{UFM}} = 0.29\text{K}$ (see Figure
190 4b), and the UFM+ and UFM- experiment SST values are given by $SST_{\text{CON-A1B}} \pm \Delta T_{\text{UFM}}$ respectively.
191 In addition to these, a further pair of experiments are performed, UFM3+ and UFM3-, with SST values

192 given by $SST_{\text{CON-A1B}} \pm 3\Delta T_{\text{UFM}}$ respectively. Other studies which consider the influence of a uniform
193 SST perturbation on the atmosphere include Kodama & Iwasaki (2009) and Chen et al. (2013) (both
194 using aquaplanet models), and Graff & LaCasce (2012) (using a full AGCM). It should be noted that as
195 well as perturbing the upper-level equator-to-pole temperature difference this experimental design may
196 introduce influences from other mechanisms, most notably changes in diabatic heating within the storm
197 tracks due to changes in precipitable water (as discussed in Schneider et al. 2010) and changes in the
198 lower-level equator-to-pole temperature difference due to remotely-forced polar amplification (see, e.g.,
199 Screen et al. 2012).

200 We note that additional factors may also influence the climate change response of the storm tracks.
201 For instance Scaife et al (2012) showed that the representation of the stratosphere can have a large
202 impact on the response of the North Atlantic storm track in particular. We do not attempt to capture
203 this effect directly, since none of the CMIP3 models have well-resolved stratospheres, however any impact
204 of this process may be at least partially captured via its impact of the upper-tropospheric equator-to-pole
205 temperature difference.

206 2.2.3 North Atlantic perturbation experiments (NATL)

207 There is a particularly large spread of North Atlantic SST responses in the CMIP3 models. This spread
208 may be interpreted as a combination of differences in the magnitude and location of the region of mediated
209 warming in the sub-polar gyre (see Figure 2a), along with differences associated with sea-ice retreat
210 (Woollings et al. (2012)). As a result there is spatial structure present in the inter-model spread of the
211 SST responses. To capture this spatial structure, the SST pattern which explains the maximum of the
212 variance between the SST response fields of the 21 CMIP3 models is calculated (this is effectively the

213 first “inter-model EOF” of the responses). This is denoted $\Delta T_{\text{EOF}}(\text{lon}, \text{lat})$ and is illustrated in Figure
214 **2b**.

215 The pattern of ΔT_{EOF} is close to zero along the southern boundary of the domain, which justifies
216 the choice of domain used, and has two distinct maxima: one in the sub-polar gyre region and the other
217 along the ice edge. Both of these maxima are present in composites representing the impact of AMOC
218 variability on the North Atlantic SSTs (see, e.g., Drijfhout et al. 2012, Figure 2), so their presence is
219 consistent with the inter-model SST differences being associated with variations in the amount of AMOC
220 reduction between the models. Since the maxima along the ice edge is most likely associated with ice
221 retreat, the ΔT_{EOF} pattern has been split into two regions (indicated A and B in the figure) capturing
222 the SST changes in the sub polar gyre and along the ice edge respectively. Perturbation experiments are
223 performed with SST values using that part of ΔT_{EOF} in region A, denoted ΔT_{NATL} . These are denoted
224 as NATL+ and NATL- and have the SST distributions $SST_{\text{CON-A1B}} \pm \Delta T_{\text{NATL}}$ respectively. The part of
225 ΔT_{EOF} in region B is used as the SST perturbation around the ice edge region in the ARC+ and ARC-
226 experiments, thus providing an SST value in regions of ice retreat.

227 **3 Results I: Control experiments (CON-20C and CON-A1B)**

228 Before considering the results of the perturbation experiments, we first consider the extent to which the
229 CON-20C and CON-A1B experiments reproduce the corresponding CMIP3 multi-model mean 20C3M
230 and SRESA1B climates.

231 **3.1 The CON-20C climate**

232 The CON-20C DJF climate is shown by the *grey contours* in Figure 3, which shows the zonal mean
233 temperature (panel **a**), the zonal wind at 850 hPa (panel **b**) and the storm track (panel **c**). These three

234 variables are used throughout as they provide a simple overview of the thermodynamic and circulation
235 response of the atmosphere. The CON-20C DJF zonal mean temperature is very similar to the CMIP3
236 multi-model mean zonal mean temperature (*contours* in Figure 1c). The CON-20C DJF U850 and storm
237 track (panels **b** and **c**) are also both broadly similar to the corresponding CMIP3 multi-model mean values
238 (*contours* in Figures 1d and **a** respectively). **However in both ocean basins the CON-20C DJF U850 and**
239 **storm track are slightly stronger than the corresponding CMIP3 multi-model mean values, and in addition**
240 **the North Atlantic and North Pacific U850, and the North Atlantic storm track, have stronger SW-NE**
241 **tilts in the CON-20C experiment than in the CMIP3 multi-model mean.** It is a well-known problem that
242 many climate models have jet streams and storm tracks that are oriented too zonally, particularly in the
243 North Atlantic region (Zappa et al. 2013a). HadGAM1 appears to here have a more strongly-tilted jet
244 and storm track than the CMIP3 multi-model mean, despite being forced with CMIP3 mean SST and
245 SIC fields. Other features of the CON-20C climate are broadly similar to the CMIP3 multi-model mean,
246 justifying its use here as a control with which to compare the perturbation experiments.

247 3.2 The difference between the CON-A1B and CON-20C climates

248 The difference between the CON-A1B and CON-20C DJF climates is shown by the *shading* in Figure 3.
249 There is warming throughout the troposphere, and cooling in the stratosphere, accompanied by regions
250 of enhanced tropospheric warming in the tropical upper-troposphere and in the Arctic lower-troposphere
251 (panel **a**). The spatial structure and magnitudes of these features closely match those of the CMIP3
252 multi-model mean response (Figure 1c), giving confidence that the thermal structure of the atmosphere
253 in the HadGAM1 model is responding in a similar manner to that seen in the CMIP3 models.

254 The difference between the CON-A1B and CON-20C DJF U850 fields (Figure 3b) also shares similarity
255 with the CMIP3 multi-model mean U850 response (Figure 1c). The spatial pattern and magnitude of

256 the response is reproduced. In addition, the difference between the CON-A1B and CON-20C storm
257 tracks (Figure 3c) is also remarkably similar to the CMIP3 multi-model mean storm track response. The
258 North Pacific storm track intensifies and shifts poleward, whereas the North Atlantic storm track has an
259 intensification on its southern flank. However, the magnitudes of these features in the CON-A1B minus
260 CON-20C difference is larger than in the CMIP3 multi-model mean response, and the North Atlantic
261 intensification over Europe is more localised spatially and located further north. Some of these differences
262 may be expected since (i) a single model experiment (HadGAM1) is being compared with the smoother
263 CMIP3 multi-model mean, and (ii) the northerly position of the intensification over western Europe in
264 Figure 1a is consistent with the stronger SW-NE tilt of the North Atlantic storm track in the CON-20C
265 experiment. Therefore we conclude that the CON-A1B minus CON-20C difference is qualitatively similar
266 to the CMIP3 multi-model mean response, and whilst these are not identical they are similar enough to
267 provide confidence in the results of the perturbation experiments.

268 **4 Results II: Global perturbation experiments (UFM, ARC and UFM3)**

269 In this section results from the three global perturbation experiments (ARC, UFM and UFM3) are
270 presented.

271 4.1 ARC experiments

The DJF temperature, circulation and storm track for the ARC experiments are shown in Figures 5a-c. These plots show the difference between the ARC+ and ARC- experiments (which will be referred to as the ‘ARC response’) and therefore represent a two standard deviation perturbation to the sea ice edge position, centred on the CON-A1B climate. Before discussing these plots we first consider Figure 4 which shows the DJF climate change responses of several key variables in both the CMIP3 ensemble and in the

experiments presented in this paper. In particular, panel **a** shows the SRESA1B minus 20C3M difference for each CMIP3 model (*diamonds*) and the EXPT minus CON-20C difference for the ARC+, CON-A1B and ARC- experiments (*squares with labels*). The variables shown are the change in Arctic sea ice extent (defined as the area with SIC greater than 15%) and the following measure of the lower-tropospheric equator-to-pole temperature difference (as used in Harvey et al. (2013)):

$$\Delta T_{850} = T_{850(30S-30N)} - T_{850(60N-90N)}, \quad (1)$$

272 where $T_{850(30S-30N)}$ and $T_{850(60N-90N)}$ are the area average zonal mean temperature at the 850 hPa
273 level in the latitude ranges indicated.

274 There is a wide range of sea-ice extent responses in the CMIP3 ensemble, from around $-1 \times 10^6 \text{km}^2$ to
275 $-7 \times 10^6 \text{km}^2$. By construction, the sea-ice extents of the two ARC experiments span a large part of this
276 range: the difference between ARC+ and ARC- is a little over two standard deviations of the responses
277 in the CMIP3 ensemble. There is also a wide range of responses of the lower-tropospheric equator-to-pole
278 temperature difference in the CMIP3 models. The plot shows that the ARC + and ARC- experiments
279 capture a similar fraction the lower-tropospheric equator-to-pole temperature difference (as measured by
280 ΔT_{850}) with a span of slightly over two standard deviations of the CMIP3 ensemble. **Therefore the ARC**
281 **perturbation experiments are capturing the spread of the desired driver of storm track change.**

282 Turning to the spatial patterns of the ARC response, shown in Figures 5**a-c**, the ARC+ minus ARC-
283 zonal mean temperature difference has a localised warm anomaly in the lower-troposphere north of around
284 60N. The anomaly is strongest below 850hPa, although there is some penetration throughout the depth
285 of the Arctic troposphere. There are differences in both U850 and the storm track in the ARC response
286 (panels **b** and **c** respectively), particularly in the North Atlantic. There is a southward shift of the North
287 Atlantic jet accompanied by a reduction in storm activity over the northern North Atlantic, similar to
288 the negative phase of the NAO. In this case, however, the storm track response is widespread across all

289 of the sub-polar regions, and not confined solely to the North Atlantic sector. The sign of the storm track
290 response is consistent with the storm track responding directly to the reduced baroclinicity in the lower-
291 troposphere. The magnitude of the storm track response in the northern North Atlantic is of the order
292 0.6 hPa, which is larger than the inter-model standard deviation of the responses in the CMIP3 models,
293 whereas it is weak across the entire North Pacific. This suggests that the range of sea ice responses in the
294 CMIP3 models contributes towards the particularly large spread in the CMIP3 storm track responses in
295 the northern North Atlantic region, but not in the North Pacific.

296 Given that the response of the storm track is strong in the ARC experiments, we briefly consider
297 whether there is a relationship present between the responses of the lower-tropospheric equator-to-pole
298 temperature difference and the responses of the storm track in the CMIP3 models. To this end, Figure
299 **6a** shows the slope of the linear regression between the local storm track responses in each model and the
300 corresponding ΔT indices in each model. Similar plots are presented for the CMIP5 models and discussed
301 in detail in Harvey et al. (2013). The values in Figure 6 are scaled by the magnitude of two standard
302 deviations of the corresponding ΔT responses (as illustrated by the vertical bars in Figure 4) and the sign
303 of the mean ΔT response. Therefore the quantity shown in Figure **6a** may be expected to correspond in
304 both magnitude and sign with the ARC storm track response of Figure 5c. There is a close resemblance
305 between this regression plot and the experiment results of Figure 5c, and this provides further evidence
306 that the lower-tropospheric equator-to-pole temperature difference has an influence on the storm track,
307 with a reduction in the temperature difference resulting in a wide spread storm track reduction over
308 much of the hemisphere. Interestingly, there is only a weak signal in the North Atlantic region in Figure
309 **6a**. However, Harvey et al. (2013) show that using instead the equator-to-pole temperature difference
310 from the North Atlantic region only (rather than the full zonal mean) results in much stronger regression
311 values in the North Atlantic region, as this is also the case for CMIP3 models (not shown).

312 4.2 UFM and UFM3 experiments

Panel **b** of Figure 4 shows SRESA1B minus 20C3M differences for each CMIP3 model (*diamonds*) and the EXPT minus CON-20C differences for the CON-A1B, UFM \pm and UFM3 \pm experiments (*squares with labels*). The variables shown are the change in tropical SST and the following measure of the upper-tropospheric equator-to-pole temperature difference:

$$\Delta T_{250} = T_{250}(30S-30N) - T_{250}(60N-90N). \quad (2)$$

313 By construction, the UFM+ and UFM- experiments span exactly two standard deviations of the CMIP3
 314 tropical SST responses. However, the corresponding span of ΔT_{250} is less than one standard deviation
 315 of the CMIP3 models. Therefore the UFM+ and UFM- experiments fail to capture the spread in the
 316 desired driver of storm track change. This suggests that, whilst the tropical SST does play a role in the
 317 climate change response of the equator-to-pole temperature difference, other factors also contribute to
 318 the differences between the ΔT_{250} responses in the CMIP3 models. **For instance, the diverse range of**
 319 **subgrid parametrizations present amongst the CMIP3 models may amplify the spread from the tropical**
 320 **SST response to the upper-tropospheric temperature response. Alternatively, it may be the case that**
 321 **HadGAM1 only weakly responds to the tropical SST values compared to the other CMIP3 models.** In
 322 an attempt to capture the spread of ΔT_{250} , the pair of more-strongly-forced experiments UFM3 \pm are
 323 performed. These are identical to the UFM \pm experiments, except the uniform SST perturbation is three
 324 times stronger, and therefore the UFM3+ minus UFM3- difference spans six standard deviations of the
 325 CMIP3 tropical SST responses. This magnitude was chosen with the aim of spanning a similar fraction
 326 of the CMIP3 model spread in ΔT_{250} as the ARC experiments do in ΔT_{850} , that is, a little over two
 327 standard deviations. Figure 4**b** shows that this is achieved.

328 The spatial patterns of the UFM+ minus UFM- difference (the ‘UFM response’) are shown in Figures
329 **5d-f**, and the UFM3+ minus UFM3- difference (the ‘UFM3 response’) in Figures **5g-i**. The pattern of
330 tropospheric warming in each case (panels **d** and **g**) have the expected zonal mean structure: there is
331 warming throughout the troposphere with a maximum in the tropical upper-troposphere. The equator-
332 to-pole temperature difference therefore increases at upper-tropospheric levels in each case. There is also
333 a small amount of northern-hemisphere polar amplification present in the UFM3 response, resulting in a
334 slight reduction in the lower-tropospheric equator-to-pole temperature difference. Such remotely-forced
335 polar amplification is a well-known feature of the climate system and possible causes include changes in
336 the Northern Hemisphere poleward heat transport and changes in Arctic cloudiness (Screen et al. (2012)).
337 Given the results of the ARC experiments, this may be expected to have some influence on the circulation
338 and storm track.

339 The thermal changes are statistically significant for both the UFM and UFM3 responses. However, the
340 corresponding U850 and storm track responses in UFM (panels **e** and **f**) are too weak to be distinguished
341 from internal variability in these runs (as indicated by the limited areas of stippling in these panels), and
342 the pattern of the response of neither variable is robust to sub-sampling of the yearly data (not shown).
343 Therefore the UFM SST perturbations appear to be too small to significantly influence the storm activity
344 in this experimental setup.

345 In contrast, the UFM3 response does contain regions of distinguishable circulation response (Figures
346 **5h** and **i**). Both the U850 and the storm track exhibit a tri-polar pattern in the North Atlantic region,
347 with a decrease in both variables in the subtropics and the Arctic, and an increase over north-western
348 Europe. In the North Pacific there is an intensification of both U850 and the storm track on the southern
349 downstream flank. These patterns are qualitatively very similar to the uniform SST experiment of Graff
350 & LaCasce (2012) (e.g. their Figure 11b). It is of note that the spatial patterns of the UFM3 U850 and

351 storm track responses are very similar. This is in contrast to the ARC experiments where there was a
352 meridional shift in the U850 pattern and a general decrease in the storm track.

353 The UFM3 storm track differences are in some places of similar magnitude to the CMIP3 inter-model
354 standard deviation of Figure 1b. However, the magnitude is smaller than that of the ARC experiments.
355 In addition, the spatial pattern of the UFM3 storm track response does not match the pattern of the
356 regions of large inter-model spread, particularly in the northern North Atlantic. Both of these features
357 suggest that the upper-level equator-to-pole temperature difference is not the dominant driver of the
358 inter-model spread in the CMIP3 ensemble. Similar to the ARC response and Figure 6a, Figure 6b shows
359 the influence of ΔT_{250} on the storm track responses as diagnosed from a linear inter-model regression.
360 In this case, there is very little area exhibiting a significant correlation. This further suggests that the
361 spread in this variable is not a dominant driver of the spread between the CMIP3 model storm track
362 responses. Interestingly, Harvey et al. (2013) find some regions of significant correlation in the CMIP5
363 version of Figure 6b; the reason for this difference between CMIP3 and CMIP5 is not clear.

364 Finally, it is of note that the magnitude of the difference between the UFM3 SST fields is similar
365 to the CON-A1B minus CON-20C tropical SST difference (Figure 4a). Whilst the basic states of these
366 experiments are not the same, we briefly compare the CON-A1B minus CON-20C response of Figure 3
367 to the UFM3 response plots of Figure 5g-i. Which features, of any, of the mean climate change response
368 are produced in the uniform SST perturbation experiments? Interestingly, the UFM3 experiments do not
369 produce a poleward shift in the upstream end of the North Pacific storm track, a feature which is present
370 in both the CMIP3 multi-model mean and the CON-A1B minus CON-20C difference. This is unexpected
371 as (i) many idealised modelling studies have suggested a link between a poleward shifted storm track and
372 an increase in the upper-level equator-to-pole temperature difference, and (ii) the regression analysis of
373 Harvey et al. (2013) shows that those CMIP5 models which have larger upper-level ΔT_{250} increases tend

374 to have more poleward-shifted storm tracks. On inspection of the U850 responses in each case (Figures 3b
375 and 5h), however, there does appear to be similarities between the responses. Both have U850 increases
376 in two regions: one on the poleward side of the upstream end of the North Pacific jet, and the other on the
377 equatorward side at the downstream end. The weighting between these two regions is different in the two
378 cases. The reason for this is unclear, but may be related to differences between the SSTs in the tropical
379 Pacific driving differences in the tropical convection. In the North Atlantic, the storm track increase over
380 north-western Europe is remarkably similar to the CON-A1B minus CON-20C mean response. Several
381 studies have suggested that this pattern of response, an intensification on the southern flank of the North
382 Atlantic storm track, may be due to changes in the local SST gradient influencing the baroclinicity. These
383 experiments (and those of Graff & LaCasce (2012)) suggest however, that such a response can also be
384 achieved by a globally uniform SST perturbation. This point is discussed further below.

385 5 Results III: North Atlantic perturbation experiments (NATL)

386 The global perturbation experiments presented above show that the particularly large spread in the
387 CMIP3 storm track responses in the North Atlantic region is associated with differences in the responses of
388 the equator-to-pole temperature difference, particularly at low levels. [The study of Woollings et al. \(2012\)](#)
389 [showed that the large spread is also correlated with changes in the AMOC in the models.](#) However, it is
390 not clear from that study whether the AMOC influences the storm track via the large scale temperature
391 structure or via more local SST changes. The NATL experiments test this by perturbing the North
392 Atlantic SSTs, capturing the spread in the changes in the models without changing the equator-to-pole
393 temperature difference. The SST perturbation used is described above in Section 2.2.3, and plotted in
394 Figure 2b.

395 Figure 7 shows the difference between the pair of North Atlantic perturbation experiments (NATL+
396 and NATL-). Panel **a** shows the spatial distribution of the lower-tropospheric temperature difference
397 (T850), and panels **b** and **c** show the U850 and storm track responses. To compare the magnitude of
398 the NATL SST perturbations with the CMIP3 inter-model spread, Figure 4c shows the responses of the
399 area-average SST in the region indicated by the box in Figure 7a. The pair of NATL experiments span
400 almost two standard deviations of this measure of the CMIP3 inter-model spread.

401 The NATL T850 response consists of a warming above the North Atlantic SST anomaly, and little
402 change elsewhere (Figure 7a). However, the magnitude of the warming at 850 hPa is much smaller than
403 the polar warming in the ARC experiments. There is very little significant response in the U850 or storm
404 track in the NATL experiments (panels **b** and **c**): there is a weak equatorward shift of the jet, and an
405 increase in storm activity coinciding with, and extending downstream of, the maximum in warming.

406 Woollings et al. (2012) showed that the responses of the North Atlantic wintertime storm track in
407 the CMIP3 models depend strongly on changes in the AMOC, and suggested that the North Atlantic
408 SSTs may play a role in this relationship. The small response of the circulation in the NATL experiments
409 compared with the ARC experiments suggests that this is not the case in our experimental setup, and
410 instead the mechanism behind the AMOC link is via changes in Arctic sea ice and/or the SST anomalies
411 in the ice edge region.

412 A caveat is that there may be important features of the North Atlantic SST which are not captured
413 by our experimental design. For instance, the regions of strongest SST gradient are smoothed out in the
414 multi-model mean SST fields used here and this may reduce the sensitivity of the storm track to SST
415 perturbations. Alternatively, the atmospheric response to a North Atlantic SST perturbation may depend
416 on the position and/or tilt of the North Atlantic storm track. As noted above, the CMIP3 multi-model

417 mean has a more zonal North Atlantic storm track than is present in our experimental setup, and this
418 may align more closely with the SST gradients in the NATL experiments.

419 **6 Summary and Discussion**

420 The experiments presented here aim to elucidate the causes of the inter-model differences in the responses
421 of the Northern Hemisphere storm tracks in the CMIP3 multi-model ensemble, with a particular focus on
422 the North Atlantic region where the inter-model spread is largest. A number of possible ‘drivers’ of the
423 storm track response are considered in turn, and their role in causing the inter-model differences assessed:

- 424 – The lower-tropospheric equator-to-pole temperature difference has a strong impact on the storm track,
425 as inferred from an experiment forced by a perturbation to the Arctic sea ice. The magnitude and
426 spatial pattern of the response to this forcing suggest that there is a substantial contribution from this
427 driver to the inter-model differences in CMIP3, particularly in the northern North Atlantic region.
- 428 – The upper-tropospheric equator-to-pole temperature difference is found to have some impact on the
429 storm track, as inferred from an experiment forced by a globally-uniform SST perturbation. However,
430 the magnitude and spatial pattern of the response to this forcing are not consistent with this being
431 the dominant driver of the inter-model differences in CMIP3.
- 432 – A local SST anomaly in the North Atlantic ocean, reminiscent of the sub-polar gyre part of the pattern
433 of mediated warming in the North Atlantic associated with an AMOC reduction, has little impact on
434 the North Atlantic storm track.

435 In each case, the magnitude of the perturbations were chosen with the aim of representing one standard
436 deviation of the spread in the CMIP3 models of the corresponding driver of change. However, as different
437 mechanisms cannot be completely isolated in this experimental setup, no attempt is made to precisely

438 quantify the impacts of each driver, and instead we restrict attention to a more qualitative assessment of
439 the responses in each case.

440 The spatial pattern of the response of the circulation to the globally-uniform SST perturbation cap-
441 tures features of the mean climate change response, particularly in the North Atlantic region where there
442 is a tri-polar pattern in U850 and storm track responses. The response to the Arctic sea-ice perturbations,
443 in contrast, has an NAO-like pattern which is qualitatively different to the mean response in the CMIP3
444 models. Therefore whilst the experiments show that the loss of Arctic sea ice contributes to the large
445 inter-model differences in the CMIP3 storm track responses, it is not the dominant cause of the mean
446 climate change response. In addition there are some regions, particularly in the latitude band 30N-40N,
447 where neither perturbation captures the magnitude of the inter-model spread, indicating that additional
448 factors influence the storm track responses there. It is interesting that the eddy-driven jet and the storm
449 track responses have very similar spatial patterns in both the control climate change experiment and
450 the globally-uniform SST experiments, whereas there is little relationship between them in the ARC
451 experiments.

452 The lack of impact of the local North Atlantic SST anomaly suggests that the relationship between
453 the responses of the AMOC and the responses of the North Atlantic storm track (Woollings et al. (2012))
454 is via changes in Arctic sea ice and related SSTs, rather than sub-polar gyre SST anomalies. However,
455 there are two important caveats to this conclusion: model biases in the position and tilt of the North
456 Atlantic storm track may cause the atmospheric response to a given SST anomaly to be model dependent,
457 and the experimental design used here may not capture the spread in the most relevant drivers of storm
458 track change.

459 This study has focused on the CMIP3 models, however both the magnitude of the mean storm track
460 response in the CMIP5 models and the inter-model spread between the CMIP5 models have similar

461 magnitudes and spatial distributions as the CMIP3 models (see Harvey et al. 2013, Figure 2b). Therefore
462 the results of this study are most likely also relevant to the CMIP5 ensemble, although this remains to
463 be checked.

464 **A List of CMIP3 models used in this study**

465 Data from the following 21 models are used in this study. One run per model is used, and this is the run denoted ‘run1’ in
466 almost all cases. It is noted in the main text if particular fields are unavailable for any of the models.

467 BCCR-BCM2.0 (Bjerknes Centre for Climate Research), CGCM3.1(T47) and CGCM3.1(T63) (Canadian Centre for Cli-
468 mate Modelling & Analysis), CNRM-CM3 (Mètèo-France / Centre National de Recherches Mètèorologiques), CSIRO-Mk3.0
469 and CSIRO-Mk3.5 (CSIRO Atmospheric Research), ECHAM5/MPI-OM (Max Planck Institute for Meteorology), ECHO-G
470 (Meteorological Institute of the University of Bonn, Meteorological Research Institute of KMA, and Model and Data group),
471 GFDL-CM2.0 and GFDL-CM2.1 (US Dept. of Commerce / NOAA / Geophysical Fluid Dynamics Laboratory), GISS-AOM
472 and GISS-ER (NASA / Goddard Institute for Space Studies), INGV-SXG (Istituto Nazionale di Geofisica e Vulcanolo-
473 gia), INM-CM3.0 (Institute for Numerical Mathematics), IPSL-CM4 (Institut Pierre Simon Laplace), MIROC3.2(hires) and
474 MIROC3.2(medres) (Center for Climate System Research (The University of Tokyo), National Institute for Environmen-
475 tal Studies, and Frontier Research Center for Global Change (JAMSTEC)), MRI-CGCM2.3.2 (Meteorological Research
476 Institute), NCAR CCSM3 (National Center for Atmospheric Research), UKMO-HadCM3 and UKMO-HadGEM1 (Hadley
477 Centre for Climate Prediction and Research / Met Office).

478 **B Design of the sea-ice concentration fields**

479 As noted in Section 2.2, due to the large range of 20C3M ice extents in the CMIP3 models (Stroeve et al. 2007), simply taking
480 the multi-model mean of the SIC values (as was done for the SSTs) produces an ice distribution with an unrealistically smooth
481 ice edge region. To avoid this problem artificial SIC distributions have been created to mimic the typical ice distribution
482 of the CMIP3 models, whilst retaining a realistic ice-edge structure. Here we describe the algorithm for constructing these
483 SIC fields.

484 Four ice distributions are used in this study, one representing the 20C3M ice distribution (used in CON-20C), one
485 representing the mean SRESA1B response (used in CON-A1B), and two capturing the spread of the responses between the

486 models (used in $\text{ARC}\pm$). For each month of the CON-20C experiment, a multi-model mean ice edge position is defined as
487 the 50% contour of the multi-model mean SIC distribution. The artificial SIC distribution is then constructed to equal 100%
488 poleward of this ice edge, linearly reducing over a distance of 5 degrees to 0% equatorward of the ice edge. The width of this
489 ice edge region was chosen by a consideration of typical ice edge regions in the CMIP3 20C3M wintertime ice distributions.

490 The remaining three ice distributions are then constructed by retreating the ice edge in CON-20C towards the pole.
491 The distance retreated, a function of longitudinal grid point, is first calculated separately for each CMIP3 model. The
492 distances retreated for CON-A1B, ARC+ and ARC- are then taken as the median and standard deviation of the retreats
493 in the models. This process is illustrated in Figure 8 which shows, for the January values, the number of degrees latitude
494 retreated by each the ice edge in each model. Also shown, in the *lower panel*, is the resulting ice edge positions of the four
495 ice distributions. Once the position of the ice edge is found, artificial SIC distributions are then constructed, as above, to
496 equal 100% poleward of this ice edge, linearly reducing over a distance of 5 degrees to 0% equatorward of the ice edge.

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501 Program for Climate Model Diagnosis and Intercomparison provides coordinating support and led development of software
502 infrastructure in partnership with the Global Organization for Earth System Science Portals.

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581 **List of Figures**

- 582 1 Multi-model mean CMIP3 climate change response. *Shading* shows CMIP3 DJF values of **a**
583 the multi-model mean response of the storm track, **b** the inter-model standard deviation of
584 the storm track responses, **c** the multi-model mean response of the zonal mean temperature,
585 and **d** the multi-model mean response of the zonal wind at 850 hPa. *Contours* in **a** and **b**
586 show the multi-model mean 20C3M storm track (units: hPa), in **c** the multi-model mean
587 20C3M zonal mean temperature (units: 10 deg C), and in **d** the multi-model mean 20C3M
588 zonal wind at 850 hPa (units: 5 ms⁻¹). The stippling in **a**, **c** and **d** indicates regions where
589 the multi-model mean responses are non-zero at the 95% confidence level, relative to the
590 inter-model spread, according to a Student's t-test. One run per model is used, as listed
591 in Appendix A, with the exception of the two UKMO models which are omitted here due
592 to the lack of available storm track data. 30
- 593 2 DJF mean values of the SST and SIC forcing fields. **a** shows the CON-20C SST (*contours*;
594 units: 4C) and the CON-A1B minus CON-20C SST difference (*shading*), **b** shows the
595 CON-A1B SST (*contours*; units: 4C) and $2 \times \Delta T_{\text{EOF}}$ (*shading*) which equal to 2 x the first
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599 ice edge position (*contour*) and the ARC+ minus ARC- SIC difference (*shading*). 31
- 600 3 DJF mean values of the CON-A1B minus CON-20C difference (*shading*). The *panels* show
601 **a** the zonal mean temperature response, **b** the U850 response, and **c** the storm track
602 response. The *grey contours* show the corresponding climatology fields from CON-20C for
603 reference (*contour intervals*: 10 deg C, 5 ms⁻¹, and 1 hPa respectively), and the stippling
604 indicates regions where the two experiments differ at the 95% confidence level, relative to
605 the inter-annual variability, according to a Student's t-test. 32
- 606 4 The DJF climate change responses of several key variables in the CMIP3 models and in
607 the experiments presented in this paper. Diamonds indicate the responses of the individual
608 CMIP3 models (SRESA1B minus 20C3M), vertical bars indicate the mean and inter-model
609 standard deviation of the CMIP3 model responses, and the squares with labels indicate
610 the EXPT minus CON-20C difference for selected experiments. The variables shown are
611 defined in the main text. 33
- 612 5 DJF mean responses in the global perturbation experiments. The *columns* show, from
613 left to right, the zonal mean temperature response, the U850 response, and the 6-2 day
614 MSLP storm track response. The *rows* show, from top to bottom, the ARC+ minus ARC-
615 difference, the UFM+ minus UFM- difference, and the UFM3+ minus UFM3- difference.
616 The *grey contours* show the corresponding climatology fields from CON-A1B for reference,
617 with *contour intervals* as in Figure 3, and the stippling indicates regions where the two
618 experiments shown differ at the 95% confidence level, relative to the inter-annual variability,
619 according to a Student's t-test. 34
- 620 6 Inter-model regression between the local CMIP3 DJF storm track responses and the DJF
621 responses of **a** ΔT_{850} and **b** ΔT_{250} . The quantity shown is the regression slope multiplied
622 by two standard deviations of the corresponding ΔT responses and the sign of the mean
623 ΔT response. The stippling indicates a correlation greater than 0.47, which is the 95%
624 confidence level for a two-tailed test. 35

625	7	DJF mean responses in the North Atlantic perturbation experiments. The <i>panels</i> show a	
626		the T850 response b the U850 response c and the 6-2 day MSLP storm track response.	
627		The <i>grey contours</i> show the corresponding climatology fields from CON-A1B for reference,	
628		with <i>contour intervals</i> as in Figure 3, and the stippling indicates regions where the two	
629		experiments shown differ at the 95% confidence level, relative to the inter-annual variability,	
630		according to a Student's t-test. The box in a indicates the region used to define the North	
631		Atlantic SST index in Figure 4 (10-60W, 40-60N).	36
632	8	The <i>large panel</i> shows the distance of ice edge retreat, as a function of longitude, for the	
633		monthly-mean January ice distributions of each CMIP3 model. The retreat is defined as the	
634		difference in ice edge position between years 1961-2000 from 20C3M and years 2061-2100	
635		from SRESA1B and is measured in units of degrees of latitude. The models are ordered	
636		by the total area of monthly mean January ice in the 20C3M period, as indicated on the	
637		left-side axis; also indicated (right-side axis) is the change in January ice area between	
638		the 20C3M and SRESA1B periods. The <i>lower panel</i> indicates the constructed January ice	
639		edge positions for the CON-20C, CON-A1B, ARC- and ARC+ experiments, as indicated.	37

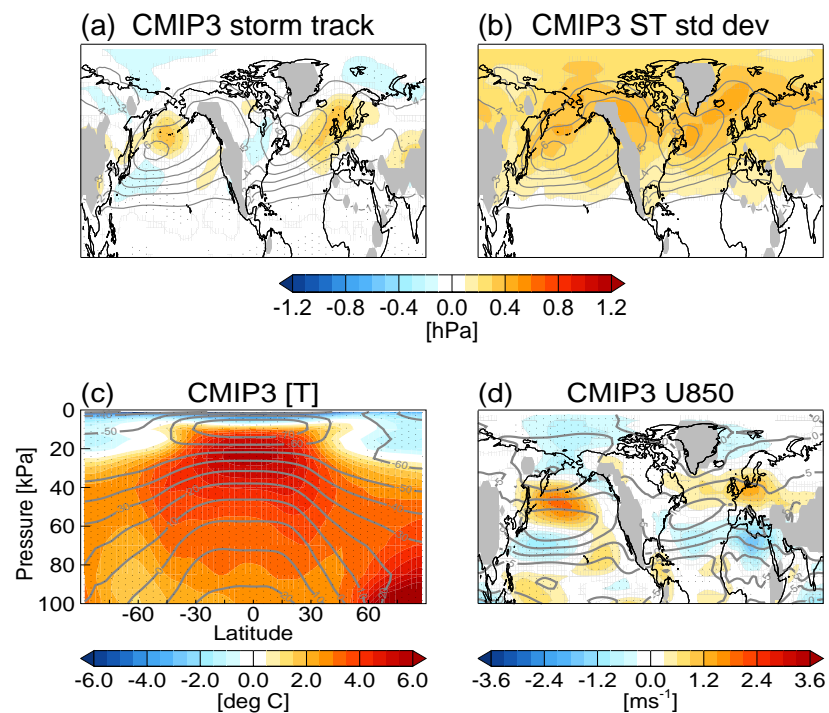


Fig. 1 Multi-model mean CMIP3 climate change response. *Shading* shows CMIP3 DJF values of **a** the multi-model mean response of the storm track, **b** the inter-model standard deviation of the storm track responses, **c** the multi-model mean response of the zonal mean temperature, and **d** the multi-model mean response of the zonal wind at 850 hPa. *Contours* in **a** and **b** show the multi-model mean 20C3M storm track (units: hPa), in **c** the multi-model mean 20C3M zonal mean temperature (units: 10 deg C), and in **d** the multi-model mean 20C3M zonal wind at 850 hPa (units: 5 ms^{-1}). The stippling in **a**, **c** and **d** indicates regions where the multi-model mean responses are non-zero at the 95% confidence level, relative to the inter-model spread, according to a Student's *t*-test. One run per model is used, as listed in Appendix A, with the exception of the two UKMO models which are omitted here due to the lack of available storm track data.

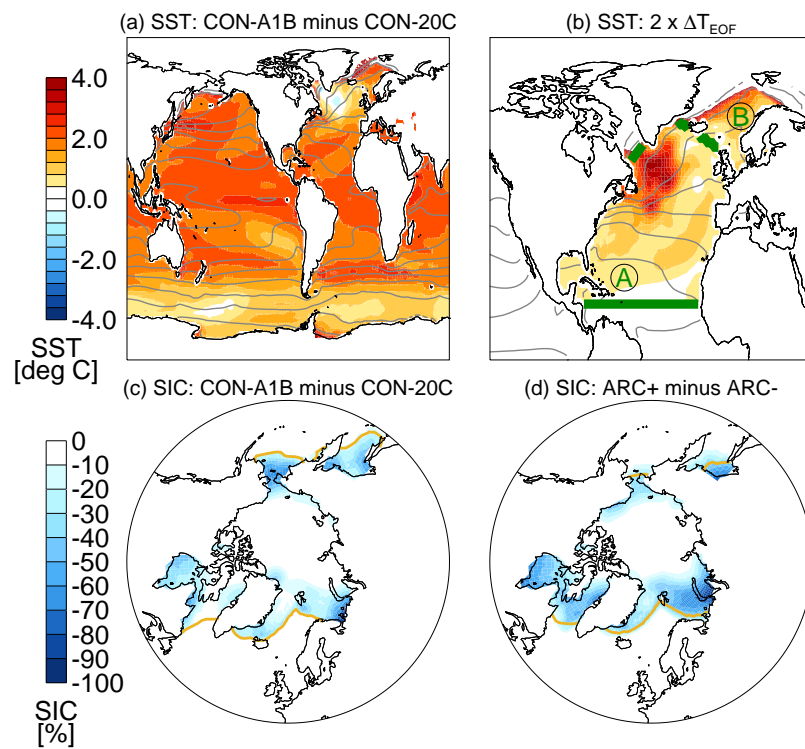


Fig. 2 DJF mean values of the SST and SIC forcing fields. **a** shows the CON-20C SST (*contours*; units: 4C) and the CON-A1B minus CON-20C SST difference (*shading*), **b** shows the CON-A1B SST (*contours*; units: 4C) and $2 \times \Delta T_{EOF}$ (*shading*) which equal to 2 x the first inter-model EOF of the CMIP3 SST responses (the EOF is normalised by the inter-model standard deviation - see text for details), **c** shows the CON-20C ice edge position (*contour*) and the CON-A1B minus CON-20C SIC difference (*shading*), and **d** shows the CON-A1B ice edge position (*contour*) and the ARC+ minus ARC-SIC difference (*shading*).

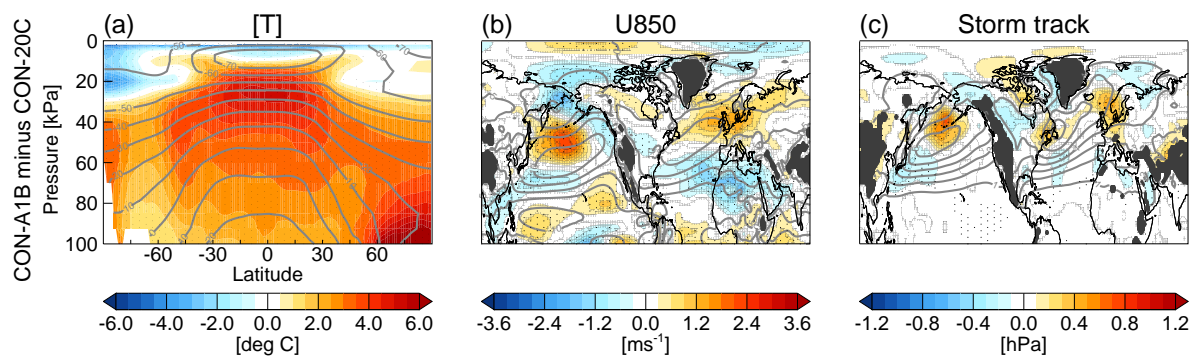


Fig. 3 DJF mean values of the CON-A1B minus CON-20C difference (*shading*). The *panels* show **a** the zonal mean temperature response, **b** the U850 response, and **c** the storm track response. The *grey contours* show the corresponding climatology fields from CON-20C for reference (*contour intervals*: 10 deg C, 5 ms^{-1} , and 1 hPa respectively), and the *stippling* indicates regions where the two experiments differ at the 95% confidence level, relative to the inter-annual variability, according to a Student's t-test.

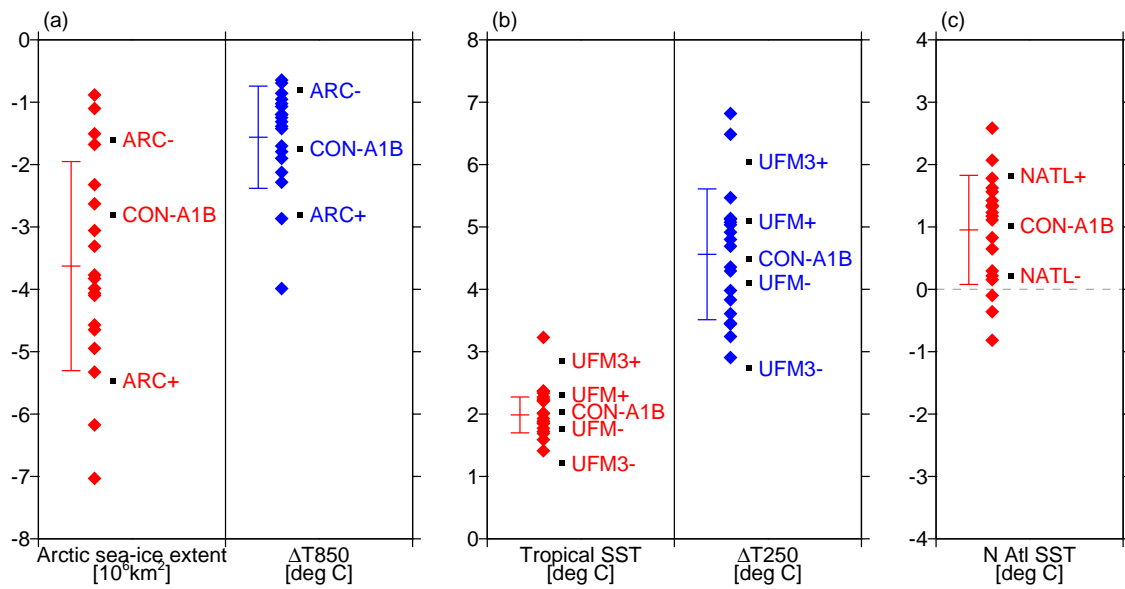


Fig. 4 The DJF climate change responses of several key variables in the CMIP3 models and in the experiments presented in this paper. Diamonds indicate the responses of the individual CMIP3 models (SRESA1B minus 20C3M), vertical bars indicate the mean and inter-model standard deviation of the CMIP3 model responses, and the squares with labels indicate the EXPT minus CON-20C difference for selected experiments. The variables shown are defined in the main text.

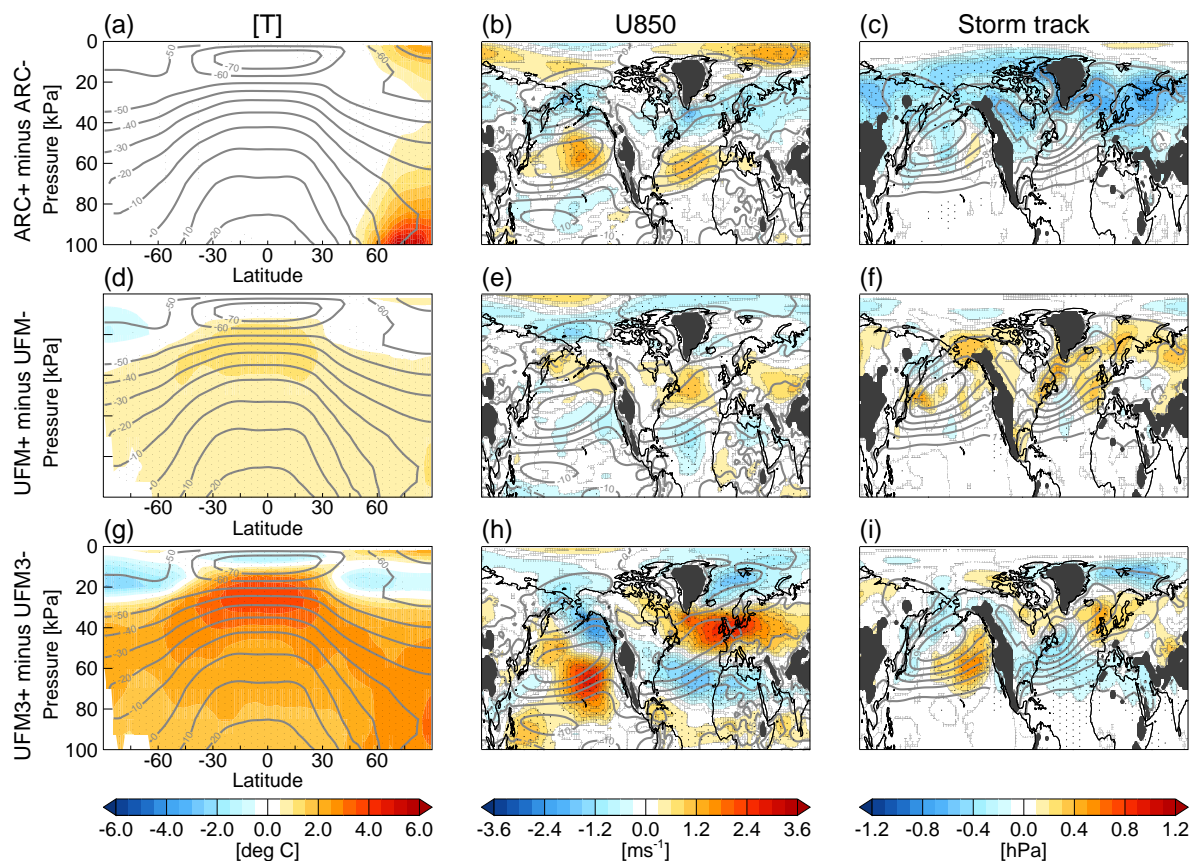


Fig. 5 DJF mean responses in the global perturbation experiments. The *columns* show, from left to right, the zonal mean temperature response, the U850 response, and the 6-2 day MSLP storm track response. The *rows* show, from top to bottom, the ARC+ minus ARC- difference, the UFM+ minus UFM- difference, and the UFM3+ minus UFM3- difference. The *grey contours* show the corresponding climatology fields from CON-A1B for reference, with *contour intervals* as in Figure 3, and the *stippling* indicates regions where the two experiments shown differ at the 95% confidence level, relative to the inter-annual variability, according to a Student's *t*-test.

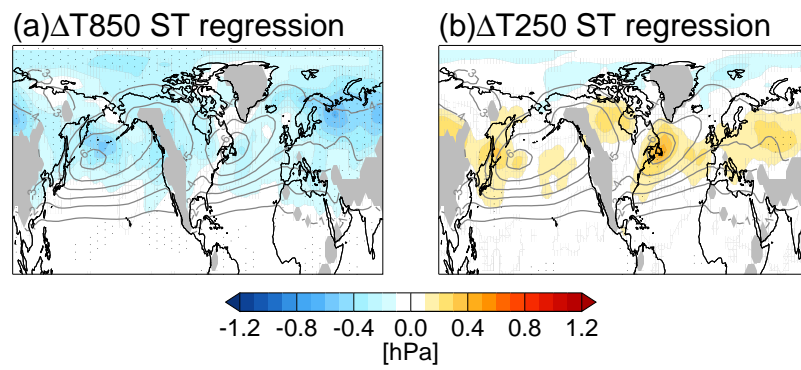


Fig. 6 Inter-model regression between the local CMIP3 DJF storm track responses and the DJF responses of **a** ΔT_{850} and **b** ΔT_{250} . The quantity shown is the regression slope multiplied by two standard deviations of the corresponding ΔT responses and the sign of the mean ΔT response. The stippling indicates a correlation greater than 0.47, which is the 95% confidence level for a two-tailed test.

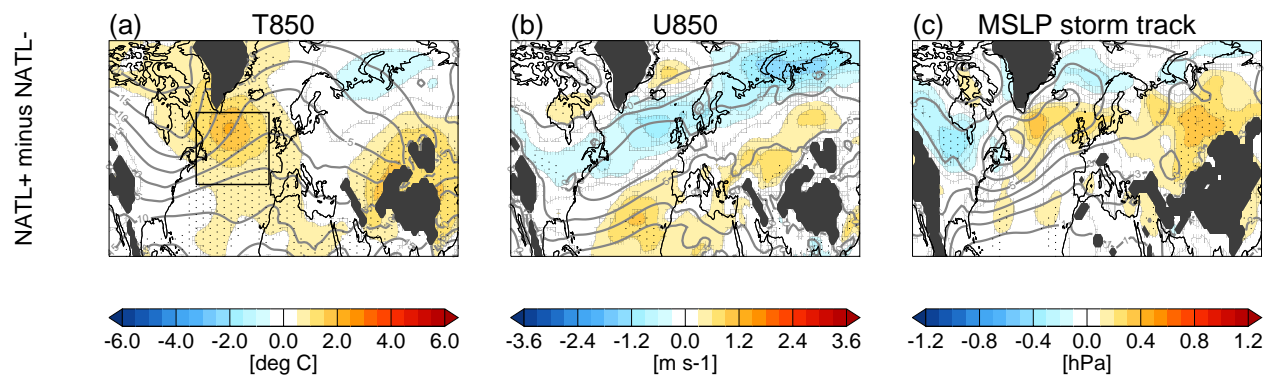


Fig. 7 DJF mean responses in the North Atlantic perturbation experiments. The *panels* show **a** the T850 response **b** the U850 response **c** and the 6-2 day MSLP storm track response. The *grey contours* show the corresponding climatology fields from CON-A1B for reference, with *contour intervals* as in Figure 3, and the *stippling* indicates regions where the two experiments shown differ at the 95% confidence level, relative to the inter-annual variability, according to a Student's t-test. The box in **a** indicates the region used to define the North Atlantic SST index in Figure 4 (10-60W, 40-60N).

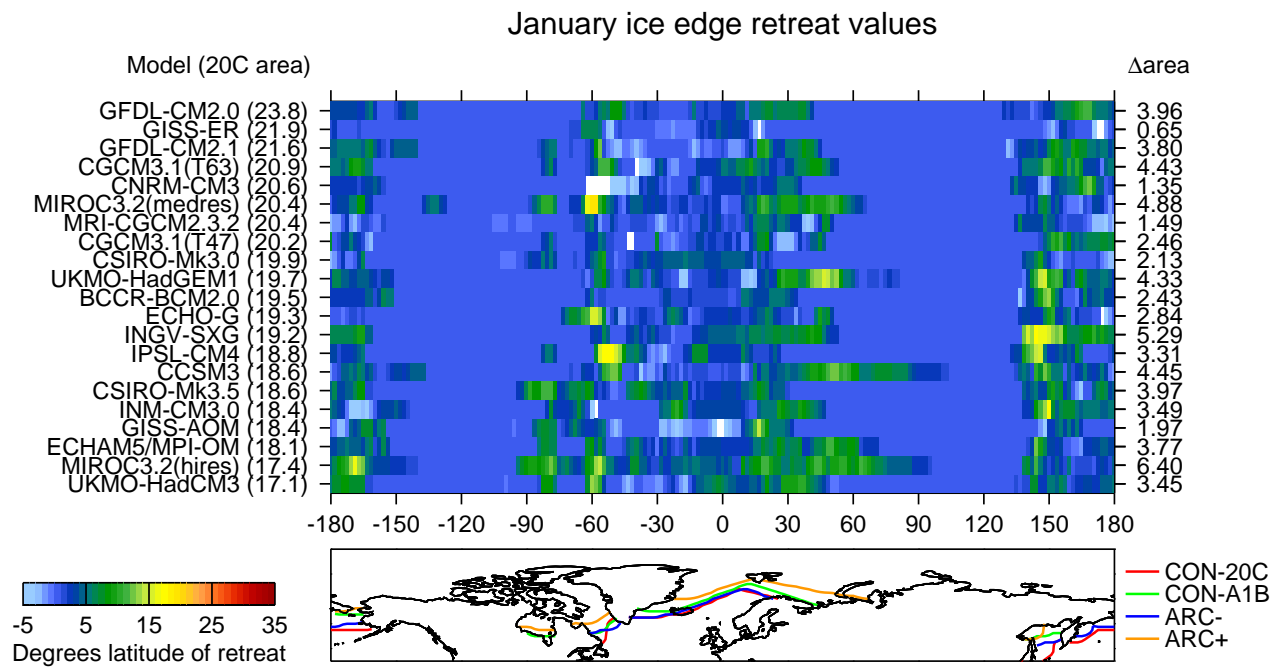


Fig. 8 The *large panel* shows the distance of ice edge retreat, as a function of longitude, for the monthly-mean January ice distributions of each CMIP3 model. The retreat is defined as the difference in ice edge position between years 1961-2000 from 20C3M and years 2061-2100 from SRESA1B and is measured in units of degrees of latitude. The models are ordered by the total area of monthly mean January ice in the 20C3M period, as indicated on the left-side axis; also indicated (right-side axis) is the change in January ice area between the 20C3M and SRESA1B periods. The *lower panel* indicates the constructed January ice edge positions for the CON-20C, CON-A1B, ARC- and ARC+ experiments, as indicated.

640 **List of Tables**641 1 The experiments presented in this paper. 39

Experiment name	SST forcing field	SIC forcing field
CON-20C	Multi-model mean 20C3M	Reconstructed 20C3M
CON-A1B	Multi-model mean SRESA1B	Reconstructed SRESA1B
ARC \pm	CON-A1B	CON-A1B \pm perturbation to Arctic sea-ice edge
UFM \pm	CON-A1B \pm global uniform perturbation	CON-A1B
UFM3 \pm	CON-A1B \pm 3 \times global uniform perturbation	CON-A1B
NATL \pm	CON-A1B \pm North Atlantic perturbation	CON-A1B

Table 1 The experiments presented in this paper.