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Southern Hemisphere Westerly Wind Changes during the Last Glacial Maximum: Model-Data Comparison

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Abstract

1 The Southern Hemisphere (SH) westerly winds are thought to be critical
2 to global ocean circulation, productivity, and carbon storage. For example,
3 an equatorward shift in the winds, though its affect on the Southern Ocean
4 circulation, has been suggested as the leading cause for the reduction in at-
5 mospheric CO₂ during the last glacial period. Despite the importance of
6 the winds, it is currently not clear, from observations or model results, how
7 they behave during the last glacial. Here, an atmospheric modelling study
8 is performed to help determine likely changes in the SH westerly winds dur-
9 ing the Last Glacial Maximum (LGM). Using LGM boundary conditions,
10 the maximum in SH westerlies is strengthened by $\sim +1 \text{ ms}^{-1}$ and moved
11 southward by $\sim 2^\circ$ at the 850 hPa pressure level. Boundary layer stabilisation
12 effects over equatorward extended LGM sea-ice can lead to a small appar-
13 ent equatorward shift in the wind band at the surface. Further sensitivity
14 analysis with individual boundary condition changes indicate that changes
15 in sea surface temperatures are the strongest factor behind the wind change.
16 The HadAM3 atmospheric simulations, along with published PMIP2 coupled
17 climate model simulations, are then assessed against the newly synthesised
18 database of moisture observations for the LGM. Although the moisture data
19 is the most commonly cited evidence in support of a large equatorward shift
20 in the SH winds during the LGM, none of the models that produce realistic
21 LGM precipitation changes show such a large equatorward shift. In fact,
22 the model which best simulates the moisture proxy data is the HadAM3
23 LGM simulation which shows a small poleward wind shift. While we cannot

24 prove here that a large equatorward shift would not be able to reproduce the
25 moisture data as well, we show that the moisture proxies do not provide an
26 observational evidence base for it.

Keywords: glacial-interglacial cycles, westerly winds, Southern Ocean,
LGM, atmospheric modelling, PMIP2, data-model comparison

27 **1. Introduction**

28 The location and the strength of Southern Hemisphere (SH) westerly
29 winds play a critical role in global climate, with glacial-interglacial change
30 in CO₂ strongly linked to ocean-atmosphere circulation feedbacks mediated
31 through the SH wind field (Sigman and Boyle, 2000; Toggweiler et al., 2006;
32 Levermann et al., 2007; Toggweiler and Russell, 2008; Toggweiler, 2009; Den-
33 ton et al., 2010; De Boer et al., 2010). For example, oceanic carbon seques-
34 tration depends strongly on Southern Ocean circulation, which is influenced
35 by this wind field (Toggweiler, 1999; Wunsch, 2003; Le Quéré et al., 2007;
36 Toggweiler and Russell, 2008). If glacial period equatorward positioned west-
37 erlies occurred, it is thought that they would curtail deep water ventilation
38 and lead to a more stratified ocean (Toggweiler and Samuels, 1995, 1998).
39 Isolated deep water traps carbon and leads to a reduction in atmospheric
40 CO₂ (Toggweiler, 1999; Watson and Naveira Garabato, 2006; Skinner et al.,
41 2010; Sigman et al., 2010).

42 In addition to this primary ocean-CO₂ modulation mechanism, other po-
43 tential oceanic impacts of wind changes have been identified. The impact of
44 SH winds on North Atlantic Deep Water (NADW) formation through dy-
45 namical processes has been established (Toggweiler and Samuels, 1993, 1995;
46 Rahmstorf and England, 1997; Nof and De Boer, 2004; De Boer and Nof,
47 2005; De Boer et al., 2008). Another hypothesis that has recently gained
48 prominence is the suggestion that SH westerly winds affect NADW forma-
49 tion through its modulation of Agulhas Leakage (Sijp and England, 2009; Beal
50 et al., 2011; Caley et al., 2012), the idea being that an equatorward shift in
51 the SH winds forces a similar shift in the Subtropical Front, in turn reduc-
52 ing the amount of high salinity Indian Ocean water that enters the Atlantic
53 Ocean. The decrease in Atlantic salt input then translates into a weaker
54 Atlantic meridional overturning.

55 Outwith the ocean, changes in the SH westerlies may also modify other
56 parts of the climate system that are critical to the ocean-atmospheric CO₂

57 balance. For example, iron rich dust borne by SH winds affects Southern
58 Ocean productivity, thereby influencing the ocean-atmosphere carbon budget
59 (Kohfeld et al., 2005).

60 Given this critical importance of SH westerly winds for the ocean circula-
61 tion and climate, it is thus important that we both understand what controls
62 their changes and how we can read wind changes from paleo-environmental
63 evidence.

64 Kohfeld et al. (submitted) discussed how the location and strength of
65 the westerly winds during the glacial period have been inferred using recon-
66 structed paleo-changes in moisture conditions and sometimes oceanic front
67 shifts (Bard and Rickaby, 2009; Kohfeld et al., submitted). Inferred changes
68 in precipitation reconstructed from vegetation regimes, lakes, glaciers, and
69 other precipitation sensitive paleo-environments have been interpreted as
70 equatorward shifts in the winds as large as 7-10° relative to their interglacial
71 position (Heusser, 1989; Lamy et al., 1999; Moreno et al., 1999; Toggweiler,
72 1999; Toggweiler et al., 2006; Toggweiler and Russell, 2008; Toggweiler, 2009).
73 We note here that this interpretation has arisen from the relationship between
74 present-day SH westerlies and precipitation (Fig. 1ac). Thus, in places such
75 as South America, an equatorward shift in westerly winds during the last
76 glacial period has been inferred from data that demonstrate enhanced pre-
77 cipitation to the north of the modern-day position of the SH westerlies, and
78 colder drier conditions in more southerly locations (Heusser, 1989; Lamy
79 et al., 1999; Moreno et al., 1999). This proposed shift has often been asso-
80 ciated with a proposed glacial weakening of the Hadley cell (Williams and
81 Bryan, 2006; Toggweiler and Russell, 2008; Toggweiler, 2009; Denton et al.,
82 2010). See Kohfeld et al. (submitted) for additional details and other possible
83 interpretations of this data.

84 Model studies do not agree on the position and strength of the westerly
85 winds during the Last Glacial Maximum (LGM - 19-23 ky) conditions. Stud-
86 ies using coupled Atmosphere-Ocean General Circulation Models (AOGCM)
87 have suggested that the westerlies move equatorward (Williams and Bryan,
88 2006) and weaken (Kim et al., 2002, 2003); move poleward and strengthen
89 (Shin et al., 2003); move poleward (Kitoh et al., 2001); or increase in strength
90 with no latitudinal shift (Otto-Bliesner et al., 2006). Analysis of more recent
91 AOGCM Paleoclimate Modelling Intercomparison Project 2 (PMIP2) wind
92 results still shows similar variation between different models (Meniel et al.,
93 2008; Rojas et al., 2009). Likewise, studies using atmosphere-only General
94 Circulation Models (AGCMs) have shown both equatorward (Drost et al.,

95 2007) and poleward shifts (Wyrwoll et al., 2000).

96 This apparent disconnect between modelling results and paleo-observations
97 represents a serious gap in our knowledge about the nature of changes in
98 westerly winds, and our ability to understand the impacts of these changes.
99 This paper represents the second of two research contributions to examine
100 how westerly winds behaved during the LGM. In our first paper (Kohfeld
101 et al., submitted), we synthesize and summarise interpretations of paleo-
102 environmental observations relating to the SH westerlies during the LGM. We
103 conclude that resolving the position and strength of the westerly winds based
104 on this data alone remains difficult because of several possible interpretations.
105 Kohfeld et al. (submitted) shows that although atmospheric evidence
106 has been frequently used to support the hypothesised large equatorward shift
107 or strengthening, the assumptions used have not been tested previously.

108 Here, our goal is to make progress through modelling, and by an atmo-
109 spheric data-model intercomparison. To this purpose we perform a set of
110 atmospheric modelling experiments to simulate the glacial winds using LGM
111 boundary conditions. We further test the sensitivity of the winds to each in-
112 dividual LGM boundary condition change. In section 2 we lay out the design
113 of the atmospheric modelling experiments, note the PMIP2 AOGCM coupled
114 simulations to which we compare our results, and explain how we quantify
115 the comparison of model output with observations. In section 3, we explore
116 how SH westerly winds change in response to individual LGM boundary con-
117 dition changes. In section 4, we explore the SH moisture changes, comparing
118 HadAM3 AGCM and PMIP2 AOGCM moisture changes with the main Ko-
119 hfeld et al. (submitted) moisture observation synthesis. Implications of the
120 results are discussed within the context of observed dust and oceanic front
121 changes in section 5.

122 **2. Experiment design**

123 *2.1. AGCM (HadAM3) model description*

124 To test the sensitivity of SH winds and moisture to changes in glacial-
125 interglacial atmospheric boundary conditions, experiments are set-up using
126 an atmospheric-only model HadAM3 on a regular latitude longitude grid
127 with a horizontal resolution of $2.5^\circ \times 3.75^\circ$, and 20 hybrid coordinate levels
128 in the vertical (Pope et al., 2000). HadAM3 features a good representation of
129 present day model climatology (Pope et al., 2000). Where the model is run in
130 its coupled state (HadCM3), it features a reasonable representation (Fig 1b)

131 of the mean annual SH westerlies (Russell et al., 2006) and produces realis-
132 tic simulations of coupled wind dependent low frequency variability (Gordon
133 et al., 2000; Collins et al., 2001; Sime et al., 2006). This suggests that the
134 atmospheric response to sea surface temperature and sea-ice change is re-
135 liable. A pre-industrial control simulation is performed along with 9 LGM
136 sensitivity experiments.

137 *2.2. Simulation boundary conditions*

138 The prescribed boundary conditions for each model integration includes
139 values for sea surface temperature, sea-ice, land-ice, solar insolation, and
140 atmospheric gas concentrations.

141 The sea surface temperature and sea-ice data for our pre-industrial (PI)
142 control simulation are based on a twenty year monthly average of HadISST
143 data (Rayner et al., 2003). Ice-sheet volume (and therefore also the land-sea
144 masking) and insolation conditions are taken from the present-day, but at-
145 mospheric gas composition is set to pre-industrial values (CO_2 is 280 ppmv;
146 CH_4 is 0.76 ppmv; and N_2O is 0.27 ppmv). SH westerly winds in this pre-
147 industrial simulation (Fig. 1a) are close to the early period European Centre
148 for Medium-Range Weather Forecasts (ECMWF) zonal winds (Fig. 1b);
149 early period data are used to validate the model simulation to minimise the
150 effect of warming or ozone changes on the comparison (Shindell and Schmidt,
151 2004). Similarly, the pre-industrial precipitation, is also in reasonable agree-
152 ment with ECMWF precipitation. The comparison between ECMWF and
153 the control simulation suggests that the SH westerlies and the precipitation
154 regime in the HadAM3 pre-industrial experiment are reasonable in compar-
155 ison with coupled model results (Russell et al., 2006).

156 For the LGM, the most recent compilation of sea surface temperature data
157 is from the MARGO project MARGO Project Members (2009). We treat it
158 here as the best observational estimate available of glacial sea surface condi-
159 tions. However, at present, the MARGO compilation has not been interpo-
160 lated to produce a globally gridded dataset. A global dataset is required to
161 drive a global atmospheric general circulation model. The most recent grid-
162 ded version of the GLAMAP LGM (19-22 ky) sea surface conditions (Paul
163 and Schäfer-Neth, 2003a,b) is therefore used to drive the HadAM3 experi-
164 ments (Fig. 2). Fig. 3a shows the difference between the gridded GLAMAP
165 sea surface temperatures and sea-ice values and the MARGO data. The
166 GLAMAP LGM sea surface temperatures are mainly within 1-2 K of the

167 nearest MARGO values. However, some larger regional discrepancies oc-
168 cur, particularly around New Zealand, GLAMAP values here are generally
169 2-4 K too warm. The GLAMAP dataset is also used for our LGM sea-ice
170 conditions.

171 LGM sea surface temperature data is rather sparse poleward of 40°S, lead-
172 ing to difficulty in establishing accurate meridional temperature gradients in
173 this region (MARGO Project Members, 2009). However, paleo-observations
174 of sea-ice extent allows estimation of the sea surface temperatures near the
175 geographical LGM sea-ice limit (at about 55°S), at roughly minus 2°C (Ger-
176 sonde et al., 2005). Thus we have reasonable confidence that there was a
177 steeper LGM extra-tropical temperature gradient (Paul and Schäfer-Neth,
178 2003b). Compared with the polar regions, there are more glacial sea surface
179 temperature observations from the tropical regions (Paul and Schäfer-Neth,
180 2003b; MARGO Project Members, 2009). These extra observations improve
181 confidence in the tropical boundary conditions.

182 Orbitally dependent insolation for the LGM (Laskar et al., 2004) and
183 the atmospheric greenhouse gas composition (CO₂ is 185 ppmv; CH₄ is 0.35
184 ppmv; and N₂O is 0.20 ppmv) is taken from the specified PMIP2 bound-
185 ary conditions (Braconnot et al., 2007)) and are known relatively well. Al-
186 though there is uncertainty about detailed aspects of the East Antarctic
187 glacial ice-sheet morphology, we have used the broad-scale glacial-interglacial
188 ICE-5G(VM2) model mass changes in which there is reasonable confidence
189 (Peltier, 2004).

190 *2.3. AGCM (HadAM3) sensitivity simulations*

191 In addition to the PI control simulation, nine atmospheric-only GCM sim-
192 ulations are used to assess the LGM SH wind and moisture changes. The
193 most realistic LGM experiment, named LGM, applies all the known boundary
194 conditions for that period as described in the previous section. The individ-
195 ual sensitivity experiments are designed to simulate the effect of the various
196 observed LGM changes in sea surface temperatures, sea-ice, ice-sheets, in-
197 coming solar radiation, and atmospheric gas composition. The experiments
198 are generated by varying one boundary condition at a time from the PI to
199 the LGM conditions (Table 1). The atmospheric simulations are run for
200 30 years, which is long enough to allow the atmosphere to reach an equi-
201 librium state with the specified boundary conditions (for atmospheric-only
202 simulations, an equilibrium state is reached after approximately two to four

203 weeks). Results are presented as differences between pre-industrial and the
204 LGM-based simulations.

205 The details of the sensitivity experiments are as follows. The SST exper-
206 iment is identical to PI except using LGM sea surface temperature values.
207 The SEAICE experiment is identical to PI, except that LGM sea-ice con-
208 ditions are applied, *i.e.* this experiment uses LGM sea-ice-conditions, but
209 pre-industrial sea surface temperatures. This means that in the experiment
210 SEAICE sea-ice is prescribed in some areas where the surface temperature
211 is above -1.8°C . The individual SST and SEAICE experiments are not in-
212 tended to represent realistic simulations of past conditions, but are useful
213 as they allow us to investigate separately the boundary condition effect of
214 sea surface temperature and sea-ice changes. SST SEAICE is more realistic
215 since it combines the sea surface temperature and sea-ice conditions from
216 the SST and SEAICE experiments. The TRO and EXT experiments are
217 variations of the SST SEAICE experiments. SST TRO is based on PI but
218 has the LGM GLAMAP reconstruction imposed equatorward of 20° with a
219 linear reduction in the anomaly imposed between 20° and 30° : poleward of
220 30° SST TRO is identical to PI. Similarly, SST EXT has the GLAMAP sea
221 surface reconstruction imposed between 90° and 30° , with a linear reduction
222 to zero between 30° and 20° , and PI conditions equatorward of 20° . LAND
223 ICE ice-sheet volume, ice-sheet topography, and land-sea masking is adjusted
224 to LGM conditions. ORBIT varies from the PI experiment in orbital param-
225 eters. GAS uses LGM atmospheric gas composition values. The HadAM3
226 LGM experiment combines all the different LGM changes.

227 To help examine the impact of uncertainties in the sea surface temper-
228 ature observations, twelve additional sensitivity simulations were also per-
229 formed. See Appendix A for these additional experiments.

230 *2.4. Description of AOGCM (PMIP2) simulations*

231 To assess the robustness of the results from the AGCM HadAM3 ex-
232 periments, the published AOGCM PMIP2 simulations are also analysed
233 where surface and 850 hPa wind and precipitation fields are available. The
234 AOGCM pre-industrial simulations are similar in specification to the full
235 LGM AGCM simulation, except that they are run with a full dynamic ocean
236 and sea-ice model, rather than with specified sea surface conditions (Table 1).
237 The simulations are run for long enough to allow the atmosphere and oceans
238 reach quasi-equilibrium state to the specified boundary conditions and for

239 trends to become small (Rojas et al., 2009). In these ocean-atmosphere sim-
240 ulations, the pre-industrial SH westerlies band is less well simulated compared
241 to our atmospheric-only simulation, with a general bias towards westerlies
242 that are too intense and located too far north (Fig. 1b). These pre-industrial
243 simulation errors hinder comparison with the glacial simulation winds. This
244 difficulty can cloud the picture of interglacial to glacial wind changes obtained
245 from AOGCM simulations (Rojas et al. 2009). However, their atmosphere,
246 ocean, and sea-ice components are dynamically consistent with each other
247 (Rojas et al., 2009) and analysis of these coupled simulations sheds light on
248 the sensitivity of our results to model choice.

249 Analysing coupled AOGCM LGM multi-model simulations, alongside
250 core AGCM simulations, ensures that a wide possible range of LGM model
251 results are considered. The comparison between MARGO sea surface tem-
252 perature data and the two PMIP2 simulations, for which we have the com-
253 parative sea surface temperature output, suggests that as we might expect
254 the LGM sea surface conditions are slightly better represented by the obser-
255 vationally based GLAMAP AGCM boundary condition data, compared to
256 the AOGCM simulations (Fig. 3). Some problems in the accuracy of sim-
257 ulated LGM sea surface conditions from coupled atmospheric-ocean models
258 may be due to difficulties in accurately parameterising glacial ocean mixing
259 terms (Wunsch, 2003; Watson and Naveira Garabato, 2006), and also partly a
260 problem of model resolution. Both of these problems affect the simulation of
261 the narrow Southern Ocean frontal jets which dominate the Southern Ocean
262 surface temperature patterns. Further improvements in ocean model param-
263 eterisation, and also in model resolution, may in future help to resolve these
264 glacial ocean simulation problems. These type of ocean modelling problems
265 also afflict coupled PMIP2 pre-industrial simulations (Russell et al., 2006),
266 thus errors in the pre-industrial ocean surface temperature simulation may
267 be the cause of errors in the simulated pre-industrial SH westerly winds (Fig.
268 1b).

269 *2.5. Data-model comparison metrics*

270 The Kohfeld et al. (submitted) paleo-observational study compiles evi-
271 dence of widespread glacial-interglacial moisture changes in the SH. Simula-
272 tion results are evaluated against 105 observations from 97 locations in this
273 new synthesized moisture dataset, by visual inspection and using two quan-
274 titative approaches. The first quantitative approach compares simulated and
275 observed changes at the ‘exact’ observation location, whilst the second checks

276 for a ‘local’ match. The local-match approach helps allow for model-inability
277 to represent small scale topographic features that are not represented in
278 GCMs but may affect observations. Both model evaluation techniques allow
279 the use of simple percentage statistics (Tables 1-3). This independent evalu-
280 ation of the goodness of model simulations helps in allowing investigation of
281 both glacial-interglacial SH circulation and moisture changes.

282 The assessment of the exact-type and local-type agreement between mod-
283 elled and observed moisture matches is made by classing simulated precip-
284 itation changes as positive or negative, if they are larger than $\pm 5\%$. This
285 5% threshold allows for a total 10% to be attributable to uncertainty in the
286 interpretation of proxies, or simulation of the changes. A simulated change
287 of less than 5% is classified as no change. This threshold is narrow compared
288 to discussion on pollen-proxy sensitivity (Bartlein et al., 2011), however re-
289 cent evaluations of model simulations against paleo-data have shown that
290 models tend to underestimate the magnitude of precipitation changes, par-
291 ticularly regional changes (Braconnot et al., 2012). Our result here support
292 the Braconnot et al. (2012) finding: the 5% threshold gives a better match
293 with paleo-data compared with using a wider $\pm 10\%$ threshold (not shown).

294 In the exact-type approach, simulation change values are linearly inter-
295 polated to the exact observation site before the threshold is applied. The
296 number of sites which match the observed changes can then be presented as
297 simple percentage statistic, *i.e.* if 20 of 100 observations match, then the sim-
298 ulation scores 20% (Table 1). For a local-type approach, simulation changes
299 are calculated as above (using the same threshold), but rather than interpo-
300 late directly to the observation site, the simulation results (on the original
301 model grid) are searched within a given radius for a locally correct match
302 (Table 1, last column). This local approach helps allow for model inability
303 to represent localised effects of gradients on moisture: small-scale features
304 affect observations but relatively coarse resolution models cannot simulate
305 them. Even given the local-type method, it is unlikely that any relatively
306 coarse resolution model study will achieve a perfect observation match.

307 The local-type percentage agreements are checked using a variety of spa-
308 tial search radiuses. The main local-type match results given in Table 1 use
309 a radius of 400 km. This radius is used because it matches the approximate
310 resolution of the model grid at the equator (3.75° distance in longitude at
311 0° latitude is 417 km). However, as one moves towards the poles this dis-
312 tances decreases (e.g. it is 341 km at 35°S , 295 km at 45°S , 268 km at 50°S ,
313 and so on). Largely because of this variation, it is not possible to choose

314 a search radius where the 'local' and 'exact' model-observation agreement
315 results are in perfect agreement. However, in general, reducing the radius
316 towards about 250 km brings these results into rough agreement, since that
317 means that usually about one model grid point is within the local search
318 radius. We also present metrics in Table 2 and 3 using these alternative
319 search radii. It can be argued that data-model metrics for the extra-tropical
320 region, from 35°S to 90 °S are the more relevant for understanding Southern
321 Ocean westerly wind changes. We therefore also provide results for separate
322 tropical (0-35°S) and extra-tropical (35-90°S) sub-domains in Tables 2 and
323 3, and compare these results to those for the whole SH domain.

324 Due to the three-value nature of our results (wetter, drier, or no change)
325 we cannot apply common statistical tests to the datasets. However, by ran-
326 domising the relationship between the given observations against simulation
327 results, we can calculate null agreement values for the simulations. Assuming
328 the simulation of equal areas or wetter, drier, and no change in precipitation,
329 the exact-type null value is 33%, and the local-type null value is 57% for a
330 400 km radius, and 44% for 250 km. Thus percentage values higher than
331 these numbers indicate skill in the model simulation. Note whilst local-type
332 null values vary slightly between simulations (due to the variable geograph-
333 ical pattern of changes), using the ECHAM5, MIROC3.2.2, and HadCM3
334 AOGCM LGM simulation, null results are within two percentage points of
335 each other.

336 Moisture observations tend to reflect moisture flux (*i.e.* P-E). However,
337 for HadAM3 simulations, the precipitation and moisture flux observation-
338 model local-type match comparisons are quite similar: within 10% of each
339 other for simulations that do not involve ICE-5G; and with a maximum dif-
340 ference of 12% for simulations using ICE-5G. Since we do not have PMIP2
341 moisture flux fields available, observations are tested against simulation pre-
342 cipitation changes for both HadAM3 and coupled PMIP2 simulations (Table
343 1, right columns).

344 **3. Simulated changes in LGM winds**

345 In our atmospheric LGM simulation full glacial boundary conditions are
346 imposed and the results compared to the pre-industrial control. For this
347 simulation, the 850 hPa westerlies show a small increase poleward of 50°S
348 (+1.5 ms⁻¹ at 60°S) and a small decrease northward of 50°S (-1 ms⁻¹ at
349 40°S) leading to a maximum that is shifted poleward by about 2° (Fig. 4).

350 At 2° the latitudinal shift of the 850 hPa wind band obtained here is rather
351 small and in the opposite direction from what has often been suggested (Tog-
352 gweiler et al., 2006; Toggweiler and Russell, 2008; Toggweiler, 2009). In the
353 core region of the westerlies, at about 52°S , the maximum increases by ap-
354 proximately $+1 \text{ ms}^{-1}$. We concentrate here on wind changes at 850 hPa
355 level, which are similar to those at the surface but do not include boundary
356 layer effects due to sea-ice cover. Surface wind effects and moisture changes
357 are considered in the later sections. Our sensitivity experiments (Table 1)
358 allow the effect of various LGM boundary conditions to be compared to their
359 individual impact on the atmospheric circulation (Fig. 5). The experiments
360 are used to examine LGM changes in the westerlies due to: [1] extra-tropical
361 and tropical sea surface conditions; and [2] orbitally dependent insolation,
362 atmospheric gases, and the morphology of the Antarctic ice-sheet.

363 *3.1. The role of sea surface changes*

364 In the SST EXT simulation where tropical sea surface temperatures are
365 held at pre-industrial levels and glacial sea surface temperatures are imposed
366 poleward of 20° , a small $\sim 1^\circ$ poleward shift in the location of the winds
367 maximum occurs. Pronounced cooling near the edge of the LGM Southern
368 Ocean sea-ice (Fig. 5ab) and extended sea-ice coverage (Fig. 5c) drives a
369 small intensification of the westerlies. The wind intensification is largest be-
370 tween $56\text{-}58^\circ\text{S}$; approximately 5° poleward of the pre-industrial maximum in
371 the winds location (Fig. 6a). This pattern of change drives the small pole-
372 ward shift in the location of the winds maximum. Observational evidence for
373 extended Antarctic sea-ice (Gersonde et al., 2005), supports an increase in
374 the glacial meridional temperature gradient around 55°S . This steeper South-
375 ern Ocean surface meridional temperature gradient (Paul and Schäfer-Neth,
376 2003b; Gersonde et al., 2005), is associated with increased available poten-
377 tial energy (Wyrwoll et al., 2000; Wunsch, 2003), and increased potential
378 atmospheric baroclinicity. Similar intensifications in the westerlies, due to
379 southern cooling, are visible in the poleward intensifications of westerlies in
380 some previous studies (Wyrwoll et al., 2000; Yin and Battisti, 2001; Kitoh
381 et al., 2001).

382 In the SST TRO tropical simulations, sea surface temperatures equa-
383 torward of 20° are set to glacial values, while Southern Ocean sea surface
384 temperatures are set at pre-industrial values. In contrast to local temper-
385 ature gradient effects discussed above, the tropical sea surface temperature
386 changes affect atmospheric temperature gradients in regions distant from

387 the band of SH westerlies. Reduced tropical sea surface temperatures cause
388 pressure changes through modifications of the Rossby wave pattern (Lachlan-
389 Cope et al., 2007), particularly in the SH wavenumber-3 pattern (Marshall
390 and Connolley, 2006). This leads to geographical variability in wind changes
391 around the Southern Ocean (Yin and Battisti, 2001). The Indian Ocean
392 sector experiences the largest changes, but every ocean basin sector sees an
393 intensification of the glacial westerlies south of 50°S and a reduction to the
394 north (Fig. 5de). This leads to a small $\sim 1^\circ$ poleward shift in the location of
395 the winds maximum (Fig. 6a).

396 The detail of the glacial tropical sea surface temperatures is still a matter
397 of ongoing research (MARGO Project Members, 2009), however analysis of
398 additional sensitivity experiments (Appendix A) confirms that this small
399 poleward shift in the SH westerlies appears to be a robust response to tropical
400 cooling. Note a much larger uniform -4K tropical cooling causes a very
401 similar, but intensified, pattern of wind changes (not shown).

402 *3.2. The roles of orbit, greenhouse gases, and ice volume*

403 In the orbital forcing and greenhouse gas simulations, these boundary
404 conditions are changed to glacial values, whilst all other conditions are kept
405 pre-industrial. The direct forcing of glacial-interglacial changes in orbital
406 parameters and atmospheric gas composition on atmospheric circulation is
407 small. Our results confirm that orbital changes do not lead to substantial
408 changes between the pre-industrial and the LGM westerlies, whilst green-
409 house gas changes lead to small reductions in the strength of the average
410 westerlies (of $0.25\text{-}0.5\text{ ms}^{-1}$). These glacial boundary conditions only cause
411 larger changes in the westerlies in ocean-atmosphere coupled GCMs where
412 sea surface temperature and sea-ice feedbacks may amplify initial forcings
413 (Rojas et al., 2009).

414 In the glacial ice volume simulation, glacial ice-sheets are imposed (Peltier,
415 2004), whilst other boundary conditions are kept pre-industrial. The changes,
416 in continents other than Antarctica, have little effect on hemispheric scale
417 wind patterns. However, the intensity of the westerlies south of 40°S de-
418 creases according to the prescribed size of the Antarctic ice-sheet (Fig. 5f
419 and 6ab, yellow line). Katabatic winds drain out from the Antarctic con-
420 tinental ice-sheet, causing northerly and weak easterly surface flows around
421 the edge of the Antarctica (Fig. 1a, vectors). The higher ICE-5G (Peltier,
422 2004) LGM Antarctic ice-sheet results in steeper ice-sheet margin slopes,
423 with resultant increases in the strength of the katabatic winds (Heinemann,

424 2003). These more intense katabatic winds reduce the southward penetration
425 of cyclones (Parish et al., 1994), which reduces the average intensity of the
426 westerlies (Fig. 5f).

427 The Antarctic glacial ice volume increase (Peltier, 2004), together with
428 the extra-tropical surface temperature effects, reduce the simulated East
429 Antarctic Plateau LGM precipitation by approximately 70% (Fig. 7). This
430 reduction is larger than the observed decrease of about 50% (Parrenin et al.,
431 2007), and supports independent gas pressure evidence that the East Antarc-
432 tic ICE-5G ice-sheet elevation is probably too high (Masson-Delmotte et al.,
433 2006). In this case, our simulations using the ICE-5G ice reconstruction
434 may produce poleward westerly winds that are too weak; our Fig 6ab results
435 suggest somewhere from 0.2-0.5 ms^{-1} too weak.

436 *3.3. The impact of wind prognostic choice*

437 One factor which has affected comparison of previous glacial-interglacial
438 modelling studies is the choice of wind prognostic. Studies of SH circu-
439 lation change have used: surface winds (Kim et al., 2003); above-surface
440 winds; or surface shear stress (Otto-Bliesner et al., 2006). Where sea-ice re-
441 places pre-industrial open water, each of these prognostics shows a different
442 response. The intensity of surface winds is sensitive to the sea-ice extent
443 because expanded sea-ice cover leads to weak turbulent fluxes, strong sta-
444 ble stratification of the air layer above, and weaker surface winds (Fig. 5c
445 and 6c). Conversely, turbulent fluxes over the open ocean, or water with a
446 high fraction of leads, acts to intensify the surface winds (Heinemann, 2003).
447 Boundary layer sea-ice effects are also visible in some other model studies
448 which show a substantial weakening of the surface westerlies at high latitudes
449 (Kim et al., 2003). Outside the sea-ice zone, glacial-interglacial changes in
450 all atmospheric boundary conditions have a very similar impact on surface
451 and 850 hPa winds (Fig. 6ab). Whilst prognostic choice will depend on sci-
452 entific purpose, the above surface (around 850-hPa) winds may be the most
453 consistent prognostic between models for glacial-interglacial studies of the
454 SH westerlies because they do not suffer from boundary layer effects.

455 *3.4. The impact of errors in the simulation of Southern Ocean sea surface 456 temperatures*

457 A second factor, affecting coupled glacial-interglacial modelling studies is
458 errors in simulated pre-industrial Southern Ocean sea surface temperature
459 gradients. This leads to difficulty in interpreting some PMIP2 changes. For

460 example, the PMIP2-HadCM3M2 atmosphere-ocean simulation does not fea-
461 ture an enhanced glacial meridional temperature gradient and wind increases
462 in this region (Rojas et al., 2009; Drost et al., 2007). However, this is largely
463 due to overly strong temperature gradients and SH westerlies in the control
464 pre-industrial simulation (Fig. 1b). Some caution is therefore required in the
465 interpretation of Southern Ocean temperature gradient driven wind changes
466 in these coupled AOGCM simulations.

467 **4. Simulated and observed moisture changes**

468 Evaluating model results against the moisture observations offer a unique
469 means to assess the veracity of the simulated glacial-interglacial SH circu-
470 lation changes (Fig. 7-10). Examination of the the wind and precipita-
471 tion changes shows that the proposed glacial weakening of the Hadley cell
472 (Williams and Bryan, 2006; Toggweiler and Russell, 2008) occurs in several
473 of the LGM simulations, including the HadAM3 LGM simulation. Glacial
474 easterly winds weaken around 20°S in each ocean basin (Fig. 4), and LGM
475 precipitation increases on the west side of South America and South Africa
476 south of 20°S (Fig. 7). The reduced Hadley cell strength leads to reduced
477 subsidence related drying around 30°S. However, contrary to what has been
478 suggested, the weakening of the Hadley cell does not lead to a significant
479 equatorward shift (Toggweiler, 2009; Denton et al., 2010) or to a weakening
480 (Williams and Bryan, 2006; Toggweiler and Russell, 2008) of the westerlies.

481 *4.1. Interpreting Western Seaboard and wider SH LGM moisture changes*

482 The local-type match between the annual mean atmospheric-only LGM
483 simulation and observed moisture changes is 100% on the west of all SH con-
484 tinents. This is important because most paleo-observational interpretations
485 which support a northward shift in the SH westerlies come from western sea-
486 board observations, between these latitudes (Kohfeld et al., submitted). If
487 we more simply restrict the spatial domain to between 35 and 90°S (Table 2),
488 the match is 97%. An excellent model-observation match in these regions,
489 without a significant change in winds, provides a strong counter-argument
490 to the hypothesis that an equatorward shift in winds is required to explain
491 these observations.

492 The general pattern of moisture changes can be described thus. Due to
493 the extended sea-ice and Southern Ocean cooling, the extra-tropical atmo-
494 sphere cools and reduces the moisture holding capacity of atmosphere above

495 the Southern ocean (Fig. 9ac). This extra-tropical drying poleward of 50°S
496 occurs in the AGCM and AOGCM LGM simulations (Fig. 7 and 10). Equa-
497 torwards, distinct regional patterns emerge from both the observational and
498 model analysis: sites between 30°S and 45°S in western South America all
499 experienced wetter conditions at the LGM compared to pre-industrial con-
500 ditions, due to reduced dry air subsidence (Fig. 9d). Sites south of 45°S are
501 drier than pre-industrial conditions due to general atmospheric cooling and a
502 reduced moisture retaining capacity (Fig. 9ac 10). In addition to some dry-
503 ing on the central to east regions of the tip South America and Africa, there
504 is also some moistening on the south-west tips of Africa and New Zealand, in
505 both the simulations and in the paleo observations, some of which is related
506 to glacial-ice volume changes (Fig. 9b).

507 The exact-local type match analysis shows that the atmospheric-only
508 LGM simulation captures 57-83% of the moisture change observations. The
509 remaining unmatched 17% of moisture change paleo-observations include ge-
510 ographically closely spaced observations (of various signs) at about 5°S in the
511 east of Africa, at 20°S on the east coast of South America, and a few from
512 central southern region of Australia. These unmatched observations, from
513 the east or central tropical continental regions, may relate to uncertainties in
514 the tropical zonal gradient changes in our imposed sea surface temperatures
515 (MARGO Project Members, 2009). However, since simply imposing a uni-
516 form tropical cooling gives a relatively similar SH westerlies change pattern
517 (Appendix A), it is unlikely that plausible gradient errors could induce a
518 completely different simulated pattern of LGM SH westerlies.

519 *4.2. Seasonal moisture and precipitation changes*

520 Some moisture proxies, such as paleo-vegetation reconstructed from pollen,
521 are likely to reflect seasonal conditions (*e.g.* Heusser, 1990; Pickett et al.,
522 2004; Williams et al., 2009; Bartlein et al., 2011). Precipitation patterns
523 in parts of the Southern Hemisphere are distinctly seasonal in nature, and
524 the season reflected in the vegetation depends on the region examined (*e.g.*
525 Williams et al., 2009). Summer precipitation is influenced by southeast trade
526 winds (Gasse and Williamson, 2008; Zech et al., 2009) or the migration of
527 the Intertropical Convergence Zone near the equator (*e.g.* Barker and Gasse,
528 2003; Williams et al., 2009), and winter precipitation can be influenced by
529 winter migrations of the modern westerly winds (*e.g.* Gasse and Williamson,
530 2008; Lamy et al., 1999; Williams et al., 2009). Thus, it is valid to ask

531 whether the LGM moisture patterns inferred from paleo-data could there-
532 fore be mainly driven by one particular season, such as wintertime changes
533 influenced by the westerly wind band.

534 The HadAM3 LGM simulation provides a means to explore which sea-
535 sonal changes in precipitation patterns contribute to the observed changes in
536 our paleo-moisture reconstruction.

537 There are distinct changes to the precipitation patterns that persist through-
538 out the year (Fig. 8). However, there are also some seasonal changes. In
539 zones influenced by the modern-day westerlies band, all seasons show en-
540 hanced precipitation along the west coasts of South America and Africa dur-
541 ing the LGM, but the latitudes of enhanced precipitation can depend on
542 season. The overall effect is that the mean annual conditions show higher
543 precipitation all along the west coasts of these continents during the LGM,
544 and thus provide a better match to the data than individual seasons. In con-
545 trast, the simulated LGM increases in precipitation off southern Australia
546 and southern New Zealand are more seasonal, occurring predominantly in
547 austral autumn and winter. In this case, the increased winter precipitation
548 is not enough to offset the strength of the summer drying in the model. This
549 could decrease the coherence between the simulated mean annual precipita-
550 tion and observations.

551 Overall we find that the simulated mean annual conditions provide a
552 better match to the paleo-proxy data than simulated conditions for any par-
553 ticular season. It is possible that this is because the spatial extent of our
554 comparison covers regions in which different seasonal precipitation regimes
555 dominate, and the simulated mean annual precipitation is best able to in-
556 tegrate the net changes seen in all of these regions. When the exact match
557 model-data agreement metrics are calculated for these individual seasons we
558 find that the match is: 57% for the annual mean; 57% for the austral summer
559 (DJF); 55% for the austral autumn (MAM); 30% for austral winter (JJA);
560 and 47% for austral spring (SON) (see also Fig. 7). This suggests that it will
561 usually be more robust to use mean annual, rather than any individual sin-
562 gle season, results when testing LGM model simulations against the Kohfeld
563 et al. (submitted) moisture database. Finally, we note that given the ocean-
564 wind carbon hypothesis is constructed in terms of mean annual wind changes,
565 it is perhaps reassuring that the model-data match to the glacial-interglacial
566 moisture reconstruction is highest for mean annual conditions.

567 *4.3. Hadley Cell induced changes*

568 The sub-tropical moistening across each continent is related to the tropi-
569 cal sea surface changes (Fig. 9d), and an associated weakening of the Hadley
570 cell strength. Fig. 9-10, alongside Table 1 and Appendices A, show that
571 simulations which feature the weakening of the Hadley cell strength, tend to
572 have the strongest model-observation agreements. None of the simulations
573 which feature this change in the Hadley Cell strength feature a significant
574 equatorwards shift in the LGM SH westerlies.

575 Results from coupled ocean-atmosphere glacial PMIP2 simulations (Table
576 1) indicate that two AOGCMs, HadCM3M2 and MIROC3.2.2, also have
577 good matches of at least 53-70% (exact-local) to observed glacial-interglacial
578 moisture change. In common with the atmospheric only simulations, the
579 Hadley cell weakens in these simulations. Two other AOGCM simulations
580 feature lower model-observational scores of 44-63% (exact-local) and show
581 little or no reduction in Hadley cell strength. These less good matches to
582 observations suggest that these atmospheric simulations are less likely correct
583 in the SH. This PMIP2 data-model comparison supports the idea that that
584 the glacial Hadley cell was reduced in strength, but that the reduction was
585 not associated with a large (greater than 2°) latitudinal shift in the position
586 of the westerly wind band.

587 **5. Discussion of relation of results to a wider set of paleo-environmental**
588 **evidence**

589 The section above indicates that LGM moisture changes, which comprise
590 the bulk of the available LGM Southern Hemisphere paleo-environmental at-
591 mospheric change evidence (Kohfeld et al., submitted), can be explained
592 without a strengthening of westerly winds, except over the southerly areas
593 of the Southern Ocean. However, Kohfeld et al. (submitted) also discusses
594 and compiles other types of data which indicate LGM paleo-environmental
595 changes. Whilst many of these observed changes were found to be too diffi-
596 cult to interpret as evidence which might support any wind change hypoth-
597 esis, here we nevertheless briefly discuss the implications of dust and ocean
598 front evidence in the context of our simulated LGM Southern Hemisphere
599 wind and moisture changes.

600 *5.1. Observed LGM dust changes*

601 One piece of paleo-environmental evidence which has previously inter-
602 preted in terms of wind changes, is the enhanced dust fluxes to the ocean
603 surface and Antarctic continent during the LGM. Several factors could con-
604 tribute to greater dust: expanded source areas, reduced entrainment thresh-
605 olds (due to moisture reductions), stronger winds over these source regions,
606 greater residence times of dust, and increased transport lengths (due to in-
607 creased wind strengths). Although using a dust cycle model is beyond the
608 scope of this work, the LGM simulation shows drying in the areas we as-
609 sociate with LGM dust entrainment (Patagonia and shelf areas; Australia,
610 southern Africa; Fig. 7). Furthermore, stronger winds in the modern-day
611 Antarctic Circumpolar Current region and a drier atmosphere over much of
612 the Antarctic region (Fig. 4c and Fig. 7) suggest that the transport lengths,
613 and atmospheric residence time for dust, may have increased facilitating
614 greater amounts of dust transport to the Southern Ocean and Antarctica.
615 In addition, a lower LGM sea-level exposed more continental shelf, which
616 would have increased the potential dust source region size. Thus it is possi-
617 ble that the LGM atmospheric changes simulated here are in agreement with
618 observed LGM dust flux changes.

619 *5.2. Observed ocean front changes: The Southern Ocean and the Agulhus*
620 *Current*

621 On the position of Southern Ocean fronts, previous studies have argued
622 that equatorward shifts in these fronts may indicate an equatorward shift
623 of the westerly winds at the LGM (Howard and Prell, 1992; Kohfeld et al.,
624 submitted). This question cannot be addressed using our atmospheric sim-
625 ulations since the HadAM3 model use ocean surface conditions as an input.
626 The PMIP2 coupled simulations are also incapable of addressing this ques-
627 tion because their oceanic model components have a resolution too coarse
628 to resolve individual oceanic fronts. However, as a general comment on this,
629 Kohfeld et al. (submitted) both shows that the position of these fronts at the
630 LGM is presently under constrained in most sectors of the Southern Ocean,
631 and suggests that the nature of the relationship between winds and fronts is
632 also quite poorly understood.

633 Available paleo-evidence for the strength of the Southern Ocean Antarctic
634 Circumpolar Current (ACC) suggest that during the LGM it could either be
635 stronger than today (Noble et al., 2012) or of similar intensity (McCave
636 and Kuhn, 2012). If the ACC was a wind-driven current the observations

637 also indicate similar or stronger SH westerlies. However, this interpretation
638 is too simplistic because the ACC is now thought to be driven by both
639 buoyancy and wind forcing (Hogg, 2010; Allison et al., 2010; Munday et al.,
640 2011) . Recent oceanic modelling has also shown that significant changes in
641 ACC transport occur when diapycnal mixing changes in far-field sites ocean
642 basin (Munday et al., 2011), and such mixing changes have been proposed
643 for the LGM in response to sea level lowering (Wunsch, 2003). Thus, it is
644 probably not currently possible to draw any useful conclusion about whether
645 our simulated winds are in agreement with the LGM ACC strength.

646 We also note that LGM changes in Southern Ocean water masses and
647 ocean productivity cannot easily be interpreted in terms of wind changes
648 (Kohfeld et al., submitted).

649 Outside the Southern Ocean, a number of studies have indicated that
650 the volume of Agulhas Leakage was probably reduced during glacial periods
651 (Flores et al., 1999, 2003). It has been hypothesized that this is due to an
652 equatorward shift of the westerly wind belt and Subtropical Front (Bard and
653 Rickaby, 2009). However, a reduction in Agulhas Leakage may also be caused
654 by increased wind stress at the southern tip of South Africa, increasing the
655 wind stress curl over the subtropical gyre circulation in the Indian Basin
656 (Beal et al., 2011). Consistent with this particular hypothesis, our LGM
657 simulation shows increased wind speeds over this region.

658 **6. Conclusions and implications for glacial winds**

659 Whilst Southern Hemisphere (SH) westerly winds are thought to be crit-
660 ical to global ocean circulation, productivity, and carbon storage, it is cur-
661 rently not clear, from observations or model results, how they behave during
662 the last glacial (Kohfeld et al., submitted). Here we performed a set of at-
663 mospheric model simulations, including sensitivity simulations to examine
664 the impact of individual LGM boundary condition changes. We examined
665 the Southern Hemisphere westerly wind changes which occur as a result of
666 these boundary changes. Additionally, we compare both these simulations,
667 and PMIP2 coupled AOGCM simulations, with the new Kohfeld et al. (sub-
668 mitted) synthesised moisture database.

669 In our main atmospheric-only LGM simulation, the SH westerlies are
670 strengthened by $\sim +1 \text{ ms}^{-1}$ and moved southward by $\sim 2^\circ$ at the 850 hPa
671 pressure level. However, boundary layer stabilisation effects over equator-
672 ward extended LGM sea-ice can lead to a small apparent equatorward shift

673 in the wind band at the surface. It is likely that this boundary layer effect,
674 due to extended LGM sea-ice, is one reason why some previous model studies
675 have suggested an equatorward shift in SH westerly winds. Thus the impact
676 of the choice of wind prognostic can lead to a divergence in the the apparent
677 SH westerly wind changes.

678 Interestingly, we find here that a reduction in Hadley Cell strength does
679 not equate to an equatorward shift of the westerlies. However, the weakening
680 does result in a wetting of the usually dry subsidence regions, which we
681 find can explain most of the observed LGM SH moisture changes. It also
682 appears to be more robust to use mean annual, rather than individual single
683 season, results when testing LGM model simulations against the Kohfeld
684 et al. (submitted) moisture database.

685 Our data-model comparison suggests that glacial-interglacial changes in
686 atmospheric circulation are being simulated with a relatively good accu-
687 racy, according to the available atmospheric observational constraints. The
688 HadAM3 LGM simulation, which shows small poleward wind shift, produces
689 the best fit with the moisture proxy data. This implies that current models
690 perform quite well at capturing these paleo-environmental changes. As a
691 result, whilst this does not fully exclude the possibility that an equatorward
692 shift of the westerlies could result in moister conditions around 40°S, we
693 conclude there is no direct evidence in moisture change observations for any
694 equatorward shift in the SH wind band. Our model experiments are carried
695 out at a resolution of $2.5^\circ \times 3.75^\circ$. Finer model resolutions can improve the
696 representation of Southern Hemisphere winds (*e.g.* Matsueda et al., 2010),
697 thus it is possible that a better simulation of the observed moisture and wind
698 changes may be possible using models run at higher resolutions.

699 Whilst the wind changes simulated here seem to be in good agreement
700 with observed LGM moisture changes, and may also be in agreement with ob-
701 served LGM-dust trends, it has not been possible here for us to test the wind
702 changes against observed LGM ocean front data. Recent ocean modelling
703 work (*e.g.* Section 5.2) has again highlighted the difficulty in interpreting
704 the LGM ocean changes from sparse observations (Wunsch, 2003). Thus it
705 is probably not presently possible to discern if our simulated atmospheric
706 changes may be in agreement with oceanic observations. A detailed high
707 resolution ocean and biogeochemical modelling study might be one possible
708 avenue to attempt to explore this intriguing question.

709 In terms of the implications for the longer Quaternary record of atmo-
710 sphere CO₂ changes, cold glacial periods other than the LGM have insuffi-

711 cient paleoclimate observations (sea surface conditions and moisture changes)
712 with which to make a similar SH westerly wind assessment. However, ad-
713 ditional sensitivity experiments, in which sea surface temperature patterns
714 and gradients are changed (temperature reductions of between 1 and 4°C),
715 suggest that the maximum equatorward shift in the wind band that can be
716 induced is about 3°(Appendix A). This is much smaller than the 7-10 degree
717 shift of wind hypothesised for the LGM which was posited as the cause of
718 the observed rise in atmospheric CO₂ into the Holocene (Toggweiler et al.,
719 2006).

720 In summary, although Kohfeld et al. (submitted) shows that observational
721 evidence has most often been interpreted as indicating a 3 to 15° LGM equa-
722 torward shift in SH westerly winds, the synthesis of LGM paleo-environmental
723 change evidence and discussion of it's interpretation suggests that there is no
724 unambiguous observation evidence which supports the shift idea. The broad
725 analysis of GCM behaviour presented here suggests that an equatorward SH
726 westerly wind shift of more than 3° may be unlikely. Although we cannot
727 prove here that a large equatorward shift would not be able to reproduce the
728 moisture data as well, we have shown here that the moisture proxies do not
729 provide an observational evidence base for it.

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735 **References**

- 736 Allison, L.C., Johnson, H.L., Marshall, D.P., Munday, D.R., 2010. Where
737 do winds drive the Antarctic Circumpolar Current? *Geophysical Research*
738 *Letters* 37, L12605+.
- 739 Bard, E., Rickaby, R.E.M., 2009. Migration of the subtropical front as a
740 modulator of glacial climate. *Nature* 460, 380–383.
- 741 Barker, P., Gasse, F., 2003. New evidence for a reduced water balance in
742 east africa during the last glacial maximum: implication for model-data
743 comparison. *Quaternary Science Reviews* 22, 823–837.

- 744 Bartlein, P.J., Harrison, S.P., Brewer, S., Connor, S., Davis, B.A.S., Gajewski,
745 ski, K., Guiot, J., Harrison-Prentice, T.I., Henderson, A., Peyron, O.,
746 Prentice, I.C., Scholze, M., Seppä, H., Shuman, B., Sugita, S., Thompson,
747 R.S., Vial, A.E., Williams, J., Wu, H., 2011. Pollen-based continental climate
748 reconstructions at 6 and 21ka: a global synthesis. *Climate Dynamics*
749 *37*, 775–802.
- 750 Beal, L.M., De Ruijter, W.P.M., Biastoch, A., Zahn, R., 2011. On the role of
751 the Agulhas system in ocean circulation and climate. *Nature* *472*, 429–436.
- 752 Braconnot, P., Harrison, S.P., Kageyama, M., Bartlein, P.J., Masson-
753 Delmotte, V., Abe-Ouchi, A., Otto-Bliesner, B., Zhao, Y., 2012. Evaluation
754 of climate models using palaeoclimatic data. *Nature Clim. Change*
755 *2*, 417–424.
- 756 Braconnot, P., Otto-Bliesner, B., Harrison, S., Joussaume, S., Peterchmitt,
757 J.Y., Abe-Ouchi, A., Crucifix, M., Driesschaert, E., Fichefet, T., Hewitt,
758 C.D., Kageyama, M., Kitoh, A.; Laine, A., Loutre, M.F., Marti,
759 O., Merkel, U., Ramstein, G., Valdes, P., Weber, S.L., Yu, Y., Zhao, Y.,
760 2007. Results of PMIP2 coupled simulations of the Mid-Holocene and Last
761 Glacial Maximum - Part 1: Experiments and large-scale features. *Climate*
762 *of the Past* *3*, 261–277.
- 763 Caley, T., Giraudeau, J., Malaizé, B., Rossignol, L., Pierre, C., 2012. Agulhas
764 leakage as a key process in the modes of Quaternary climate changes.
765 *Proceedings of the National Academy of Sciences* .
- 766 Collins, M., Tett, S.F.B., Cooper, C., 2001. The internal climate variability
767 of HadCM3, a version of the Hadley Centre coupled model without flux
768 adjustments. *Climate Dynamics* *17*, 61–81.
- 769 De Boer, A.M., Nof, D., 2005. The island wind-buoyancy connection. *Tellus*
770 *A* *57*, 783–797.
- 771 De Boer, A.M., Toggweiler, J.R., Sigman, D.M., 2008. Atlantic Dominance of
772 the Meridional Overturning Circulation. *J. Phys. Oceanogr.* *38*, 435–450.
- 773 De Boer, A.M., Watson, A.J., Edwards, N.R., Oliver, K.I.C., 2010. A com-
774 prehensive, multi-process box-model approach to glacial-interglacial carbon
775 cycling. *Climate of the Past Discussions* *6*, 867–903.

- 776 Denton, G.H., Anderson, R.F., Toggweiler, J.R., Edwards, R.L., Schaefer,
777 J.M., Putnam, A.E., 2010. The last glacial termination. *Science* 328,
778 1652–1656.
- 779 Drost, F., Renwick, J.A., Bhaskaran, B., Oliver, H., McGregor, J., 2007.
780 Features of the zonal mean circulation in the Southern Hemisphere during
781 the Last Glacial Maximum. *J. Geophys. Res.* 112.
- 782 Flores, J.A., Gersonde, R., Sierro, 1999. Pleistocene fluctuations in the agul-
783 has current retroflexion based on the calcareous plankton record. *Marine*
784 *Micropaleontology* 37, 1–22.
- 785 Flores, J.A., Marino, M., Sierro, F.J., Hodell, D.A., Charles, C.D., 2003.
786 Calcareous plankton dissolution pattern and coccolithophore assemblages
787 during the last 600 kyr at ODP site 1089 (cape basin, south atlantic): pale-
788 oceanographic implications. *Palaeogeography, Palaeoclimatology, Palaeoe-*
789 *cology* 196, 409–426.
- 790 Gasse, F., C.F.V.A.W.M., Williamson, D., 2008. Climatic patterns in equa-
791 torial and southern africa from 30,000 to 10,000 years ago reconstructed
792 from terrestrial and near-shore proxy data. *Quaternary Science Reviews*
793 27, 2316–2340.
- 794 Gersonde, R., Crosta, X., Abelmann, A., Armand, L., 2005. Sea surface
795 temperature and sea ice distribution of the last glacial. *Southern Ocean -*
796 *A circum-Antarctic view based on siliceous microfossil records. Quaternary*
797 *Science Reviews* 24, 869–896.
- 798 Gordon, C., Cooper, C., Senior, C.A., Banks, H., Gregory, J.M., Johns, T.C.,
799 Mitchell, J.F.B., Wood, R.A., 2000. The simulation of SST, sea ice extents
800 and ocean heat transports in a version of the Hadley Centre coupled model
801 without flux adjustments. *Climate Dynamics* 16, 147–168.
- 802 Heinemann, G., 2003. Forcing and feedback mechanisms between the kata-
803 batic wind and sea ice in the coastal areas of polar ice sheets. *The Global*
804 *Atmosphere-Ocean System* 9, 169–201.
- 805 Heusser, C., 1989. Southern westerlies during the last glacial maximum.
806 *Journal of Quaternary Science* 31, 423–425.

- 807 Heusser, C., 1990. Ice-age vegetation and climate of subtropical chile. *Palaeo-*
808 *geography Palaeoclimatology Palaeoecology* 80, 107–127.
- 809 Hogg, A.M., 2010. An Antarctic Circumpolar Current driven by surface
810 buoyancy forcing. *Geophysical Research Letters* 37, L23601+.
- 811 Howard, W.R., Prell, W.L., 1992. Late Quaternary Surface Circulation of
812 the Southern Indian Ocean and its Relationship to Orbital Variations.
813 *Paleoceanography* 7.
- 814 Kim, S., Flato, G., Boer, G., 2003. A coupled climate model simulation
815 of the Last Glacial Maximum, Part 2: approach to equilibrium. *Climate*
816 *Dynamics* 20, 635–661.
- 817 Kim, S., Flato, G., Boer, G., McFarlane, N., 2002. A coupled climate model
818 simulation of the Last Glacial Maximum, Part 1: transient multi-decadal
819 response. *Climate Dynamics* 19, 515–537.
- 820 Kitoh, A., Murakami, S., Koide, H., 2001. A simulation of the Last Glacial
821 Maximum with a coupled atmosphere-ocean GCM. *Geophysical Research*
822 *Letters* 28, 2221–2224.
- 823 Kohfeld, K., Graham, R., de Boer, A., Sime, L., Wolff, E., Le Quéré, C.,
824 Bopp, L., submitted. Southern hemisphere westerly wind changes dur-
825 ing the last glacial maximum: Paleo-data synthesis. *Quaternary Science*
826 *Reviews* , –.
- 827 Kohfeld, K.E., Quere, C.L., Harrison, S.P., Anderson, R.F., 2005. Role of
828 Marine Biology in Glacial-Interglacial CO₂ Cycles. *Science* 308, 74–78.
829 <http://www.sciencemag.org/cgi/reprint/308/5718/74.pdf>.
- 830 Lachlan-Cope, T.A., Connolley, W.M., Turner, J., 2007. Effects of trop-
831 ical sea surface temperature (SST) errors on the Antarctic atmospheric
832 circulation of HadCM3. *Geophysical Research Letters* 34.
- 833 Lamy, F., Hebbeln, D., Wefer, G., 1999. High-resolution marine record of
834 climatic change in mid-latitude chile during the last 28,000 years based on
835 terrigenous sediment parameters. *Quaternary Research* 51, 83–93.
- 836 Laskar, J., Robutel, P., Joutel, F., Gastineau, M., Correia, A., Levrard, B.,
837 2004. A long term numerical solution for the insolation quantities of the
838 Earth. *Astronomy and Astrophysics* 428, 261–285.

- 839 Le Quéré, C., Rödenbeck, C., Buitenhuis, E.T., Conway, T.J., R., L., Gomez,
840 A., Labuschagne, C., Ramonet, M., Nakazawa, T., Metzl, N., Gillett, N.,
841 Heimann, M., 2007. Saturation of the Southern Ocean CO₂ sink due to
842 recent climate change. *Science* 316, 1735–1738.
- 843 Levermann, A., Schewe, J., Montoya, M., 2007. Lack of bipolar see-saw in
844 response to Southern Ocean wind reduction. *Geophysical Research letters*
845 34.
- 846 MARGO Project Members, 2009. Constraints on the magnitude and patterns
847 of ocean cooling at the Last Glacial Maximum. *Nature Geoscience* 2, 127–
848 132.
- 849 Marshall, G.J., Connolley, W.M., 2006. Effect of changing Southern Hemi-
850 sphere winter sea surface temperature on Southern Annular Mode strength.
851 *Geophysical Research Letters* 33.
- 852 Masson-Delmotte, V., Kageyama, M., Braconnot, P., Charbit, S., Krinner,
853 G., Ritz, C., Guilyardi, E., Jouzel, J., Abe-Ouchi, A., Crucifix, M., Glad-
854 stone, R., Hewitt, C., Kitoh, A., LeGrande, A., Marti, O., Merkel, U.,
855 Motoi, T., Ohgaito, R., Otto-Bliesner, B., Peltier, W., Ross, I., Valdes,
856 P., Vettoretti, G., Weber, S., Wolk, F., Yu, Y., 2006. Past and future po-
857 lar amplification of climate change: climate model intercomparisons and
858 ice-core constraints. *Climate Dynamics* 26, 513–529.
- 859 Matsueda, M., Endo, H., Mizuta, R., 2010. Future change in southern hemi-
860 sphere summertime and wintertime atmospheric blockings simulated using
861 a 20-km-mesh AGCM. *Geophysical Research Letters* 37, L02803+.
- 862 McCave, I.N., C.S.H.C.D., Kuhn, G., 2012. Constant flow speed of the ACC
863 through Drake Passage between glacial maximum and Holocene. Euro-
864 pean Geophysical Union, publisher=Vienna, Austria, doi=EGU2012-9842,
865 note=EGU2012-9842, .
- 866 Menviel, L., Timmermann, A., Mouchet, A., Timm, O., 2008. Climate and
867 marine carbon cycle response to changes in the strength of the Southern
868 Hemispheric westerlies. *Paleoceanography* 23, PA4201.
- 869 Moreno, P.I., Lowell, T.V., Jacobson Jr, G.L., Denton, G.H., 1999. Abrupt
870 vegetation and climate changes during the Last Glacial Maximum and last

- 871 termination in the Chilean Lake District: A case study from Canal de la
872 Puntilla (41s). *Geografiska annaler*, Series A 81, 285–311.
- 873 Munday, D.R., Allison, L.C., Johnson, H.L., Marshall, D.P., 2011. Remote
874 forcing of the Antarctic Circumpolar Current by diapycnal mixing. *Geo-*
875 *physical Research Letters* 38, L08609+.
- 876 Noble, T.L., Piotrowski, A.M., Robinson, L.F., McManus, J.F., Hillenbrand,
877 C.D., Bory, A.J.M., 2012. Greater supply of Patagonian-sourced detritus
878 and transport by the ACC to the Atlantic sector of the Southern Ocean
879 during the last glacial period. *Earth and Planetary Science Letters* 317-
880 318, 374–385.
- 881 Nof, D., De Boer, A.M., 2004. From the Southern Ocean to the North
882 Atlantic in the Ekman Layer? *Bull. Amer. Meteor. Soc.* 85, 79–87.
- 883 Otto-Bliesner, B.L., Brady, E., Clauzet, G., Thomas, R., Levis, S.,
884 Kothavala, Z., 2006. Last glacial maximum and Holocene climate in
885 CCSM3. *Journal of Climate* 19, 2526–2544.
- 886 Parish, T.R., Bromwich, D.H., Tzeng, R.Y., 1994. On the Role of the Antarc-
887 tic Continent in Forcing Large-Scale Circulations in the High Southern
888 Latitudes. *Journal of the Atmospheric Sciences* 51, 3566–3579.
- 889 Parrenin, F., Dreyfus, G., Durand, G., Fujita, S., Gagliardini, O., Gillet, F.,
890 Jouzel, J., Kawamura, K., Lhomme, N., Masson-Delmotte, V., Ritz, C.,
891 Schwander, J., Shoji, H., Uemura, R., Watanabe, O., Yoshida, N., 2007.
892 1-D-ice flow modelling at EPICA Dome C and Dome Fuji, East Antarctica.
893 *Climate of the Past* 3, 243–259.
- 894 Paul, A., Schäfer-Neth, C., 2003a. Gridded Global LGM SST and Salinity
895 Reconstruction, IGBP PAGES/World Data Center for Paleoclimatology,
896 Boulder Data Contribution Series 2003-046 NOAA/NGDC Paleoclimatol-
897 ogy Program, Boulder CO, USA. , –.
- 898 Paul, A., Schäfer-Neth, C., 2003b. Modeling the water masses of the Atlantic
899 Ocean at the Last Glacial Maximum. *Paleoceanography* 18.
- 900 Peltier, W., 2004. Global Glacial Isostasy and the Surface of the Ice-Age
901 Earth: The ICE-5G(VM2) model and GRACE. *Ann Rev Earth Planet*
902 *Sci* 32, 111–149.

- 903 Pickett, E.J., Harrison, S.P., Hope, G., Harle, K., Dodson, J.R., Peter Ker-
904 shaw, A., Colin Prentice, I., Backhouse, J., Colhoun, E.A., D'Costa, D.,
905 Flenley, J., Grindrod, J., Haberle, S., Hassell, C., Kenyon, C., Macphail,
906 M., Martin, H., Martin, A.H., McKenzie, M., Newsome, J.C., Penny, D.,
907 Powell, J., Ian Raine, J., Southern, W., Stevenson, J., Sutra, J.P., Thomas,
908 I., Kaars, S., Ward, J., 2004. Pollen-based reconstructions of biome distri-
909 butions for australia, southeast asia and the pacific (seapac region) at 0,
910 6000 and 18,000 14c yr bp. *Journal of Biogeography* , 1381–1444.
- 911 Pope, V.D., Gallani, M.L., Rowntree, P.R., Stratton, R.A., 2000. The im-
912 pact of new physical parametrizations in the Hadley Centre climate model:
913 HadAM3. *Climate Dynamics* 16, 123–146.
- 914 Rahmstorf, S., England, M.H., 1997. Influence of Southern Hemisphere
915 Winds on North Atlantic Deep Water Flow. *J. Phys. Oceanogr.* 27, 2040–
916 2054.
- 917 Rayner, N.A., Parker, D.E., Horton, E.B., Folland, C.K., Alexander, L.V.,
918 P., R.D., 2003. Global analyses of sea surface temperature, sea ice, and
919 night marine air temperature since the late nineteenth century. *Journal of*
920 *Geophysical Research* 108.
- 921 Rojas, M., Moreno, P., Kageyama, M., Crucifix, M., Hewitt, C., Abe-Ouchi,
922 A., Ohgaito, R., Brady, E., Hope, P., 2009. The Southern Westerlies during
923 the last glacial maximum in PMIP2 simulations. *Climate Dynamics* 32,
924 525–548.
- 925 Russell, J.L., Stouffer, R.J., Dixon, K.W., 2006. Intercomparison of the
926 Southern Ocean circulations in the IPCC coupled model control simula-
927 tions. *Journal of Climate* 19, 4560–4575.
- 928 Shin, S.I., Liu, Z., Otto-Bliesner, B., Brady, E., Kutzbach, J., Harrison,
929 S., 2003. A Simulation of the Last Glacial Maximum climate using the
930 NCAR-CCSM. *Climate Dynamics* 20, 127–151.
- 931 Shindell, D., Schmidt, G., 2004. Southern hemisphere climate response to
932 ozone changes and greenhouse gas increases. *Geophysical Research Letters*
933 L18209.
- 934 Sigman, D.M., Boyle, E.A., 2000. Glacial/interglacial variations in atmo-
935 spheric carbon dioxide. *Nature* , 859–869.

- 936 Sigman, D.M., Hain, M.P., Haug, G.H., 2010. The polar ocean and glacial
937 cycles in atmospheric co2 concentration. *Nature* 466, 47–55.
- 938 Sijp, W.P., England, M.H., 2009. Southern Hemisphere Westerly Wind Con-
939 trol over the Ocean’s Thermohaline Circulation. *J. Climate* 22, 1277–1286.
- 940 Sime, L., Stevens, D., Heywood, K., Oliver, K., 2006. A Decomposition of
941 the Atlantic Meridional Overturning. *Journal of Physical Oceanography*
942 36, 2253–2270.
- 943 Skinner, L.C., Fallon, S., Waelbroeck, C., Michel, E., Barker, S., 2010. Ven-
944 tilation of the deep southern ocean and deglacial co2 rise. *Science* 328,
945 1147–1151.
- 946 Toggweiler, J.R., 1999. Variation of atmospheric CO₂ by ventilation of the
947 ocean’s deepest water. *Paleoceanography* 14, 571–588.
- 948 Toggweiler, J.R., 2009. Shifting westerlies. *Science* 323, 1434–1435.
- 949 Toggweiler, J.R., Russell, J., 2008. Ocean circulation in a warming climate.
950 *Nature* 451, 286–288.
- 951 Toggweiler, J.R., Russell, J., Carson, S.R., 2006. Midlatitude westerlies, at-
952 mospheric CO₂, and climate change during the ice ages. *Paleoceanography*
953 21.
- 954 Toggweiler, J.R., Samuels, B., 1993. *The Global Carbon Cycle*. Springer-
955 Verlag. chapter Is the magnitude of the deep outflow from the Atlantic
956 Ocean actually governed by Southern Hemisphere winds? pp. 303–331.
- 957 Toggweiler, J.R., Samuels, B., 1995. Effect of Drake Passage on the global
958 thermohaline circulation. *Deep-Sea Research I* 42, 477–500.
- 959 Toggweiler, J.R., Samuels, B., 1998. On the ocean’s large-scale circulation
960 near the limit of no vertical mixing. *Journal of Physical Oceanography* 28,
961 1832–1852.
- 962 Watson, A.J., Naveira Garabato, A.C., 2006. The role of Southern Ocean
963 mixing and upwelling in glacial-interglacial atmospheric CO₂ change. *Tel-
964 lus* 58, 73–87.

- 965 Williams, G., Bryan, K., 2006. Ice Age Winds: An Aquaplanet Model.
966 Journal of Climate 19, 1706-1715.
- 967 Williams, M., Cook, E., van der Kaars, S., Barrows, T., Shulmeister, J.,
968 Kershaw, P., 2009. Glacial and deglacial climatic patterns in Australia
969 and surrounding regions from 35000 to 10000 years ago reconstructed from
970 terrestrial and near-shore proxy data. Quaternary Science Reviews 28,
971 2398-2419.
- 972 Wunsch, C., 2003. Determining paleoceanographic circulations, with em-
973 phasis on the Last Glacial Maximum. Quaternary Science Reviews 22,
974 371-385.
- 975 Wyrwoll, K.H., Dong, B.W., Valdes, P., 2000. On the position of southern
976 hemisphere westerlies at the Last Glacial Maximum: an outline of AGCM
977 simulation results and evaluation of their implications. Quaternary Science
978 Reviews 19.
- 979 Yin, J.H., Battisti, D.S., 2001. The importance of tropical sea surface tem-
980 perature patterns in simulations of last glacial maximum climate. Journal
981 of Climate 14, 565-581.
- 982 Zech, M., Zech, R., Morrás, H., Moretti, L., Glaser, B., Zech, W., 2009. Late
983 quaternary environmental changes in misiones, subtropical NE argentina,
984 deduced from multi-proxy geochemical analyses in a palaeosol-sediment
985 sequence. Quaternary International 196, 121-136.

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Table 1: Experiments and 0-90°S spatial domain (whole Southern Hemisphere) model-data agreement.

Experiment	sst	sea-ice	ice-sheets	insolation	gases	obs-model agreement % exact-type ¹ local-type ²
<i>atmospheric-only HadAM3 pre-industrial simulation</i>						
pre-industrial	present	present	present	present	pre-industrial	
<i>atmospheric-only HadAM3 LGM simulation</i>						
SST	GLAMAP					59
SEAICE		GLAMAP				8
SST SEAICE	GLAMAP	GLAMAP				50
SST EXT	Extra-tropical	Extra-tropical				21
SST TRO	Tropical	-				40
LAND ICE			ICE-5G			17
ORBIT				21 ky		16
GAS					21 ky	39
LGM	GLAMAP	GLAMAP	ICE-5G	21 ky	21 ky	57
<i>PMIP2 coupled LGM simulation</i>						
CNRM-CM33	coupled	coupled	ICE-5G	21 ky	21 ky	35
ECHAM53	coupled	coupled	ICE-5G	21 ky	21 ky	44
HadCM3M2	coupled	coupled	ICE-5G	21 ky	21 ky	53
MIROC3.2.2	coupled	coupled	ICE-5G	21 ky	21 ky	59

Foot note 1. Exact position match. 5% threshold between PI and LGM model simulation precipitation used to specify a wetter/drier change. See Section 2.5 for details.

Foot note 2. Local match within 400 km . 5% threshold, as above. See Section 2.5 for details.

Table 2: Model-data agreement. Sensitivity check on model-observation agreement depending on local radius and 35-90°S spatial domain (29 observations).

Experiment	400 km radius		342 km radius		295 km radius		250(200) km radius	
	0-90°S	35-90°S	0-90°S	35-90°S	0-90°S	35-90°S	0-90°S	35-90°S
SST	75	79	71	76	68	72	62	69
SEAICE	28	24	26	24	25	21	17	17
SST SEAICE	75	76	71	72	69	69	63	62
SST EXT	43	45	38	45	31	31	27	21
SST TRO	65	59	62	55	61	55	50	45
LAND ICE	66	93	55	83	51	76	43	55 (45)
ORBIT	41	41	36	38	34	34	27	24
GAS	64	69	60	62	56	59	51	45
LGM	83	97	78	90	74	86	69	79 (72)

Table 3: Model-data agreement. Sensitivity check on model-observation agreement depending on local radius and 0-35°S spatial domain (74 observations).

Experiment	400 km radius		342 km radius		295 km radius		250(200) km radius	
	0-90°S	0-35°S	0-90°S	0-35°S	0-90°S	0-35°S	0-90°S	0-35°S
SST	75	73	71	69	68	66	62	59
SEAICE	28	30	26	27	25	27	17	18
SST SEAICE	75	74	71	70	69	69	63	64
SST EXT	43	42	38	35	31	31	27	30
SST TRO	65	68	62	65	61	64	50	53
LAND ICE	66	55	55	45	51	43	43	38 (32)
ORBIT	41	41	36	35	34	34	27	28
GAS	64	62	60	59	56	55	51	54
LGM	83	77	78	73	74	69	69	65 (54)

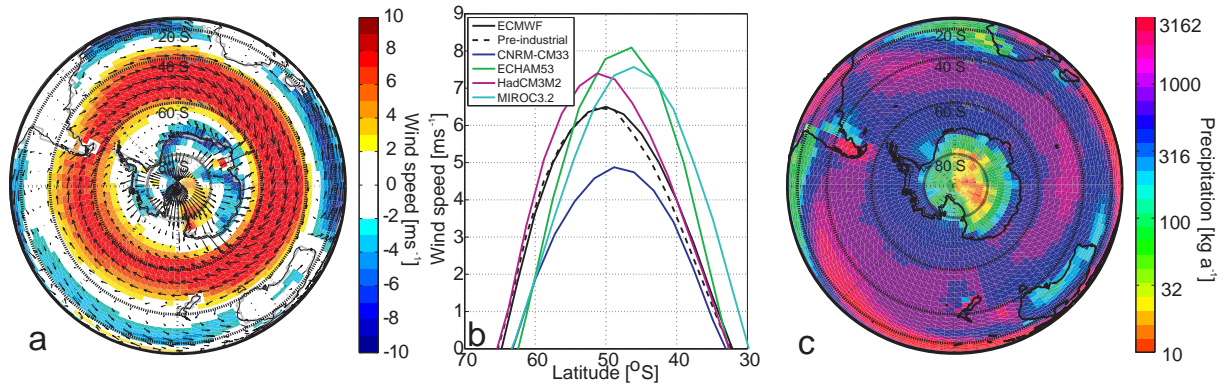


Figure 1: (a) Annual mean surface (10 m) westerly wind speed (m s^{-1}) for the pre-industrial simulation. Arrows show wind velocity. (b) The zonal mean annual mean surface (10 m) westerly wind speed from the HadAM3 (pre-industrial) and ECMWF-ERA40 reanalysis; alongside results from each PMIP2 pre-industrial (control) experiment (as noted on legend). (c) Annual mean precipitation (mm yr^{-1}) for the pre-industrial simulation on a logarithmic colour-scale.

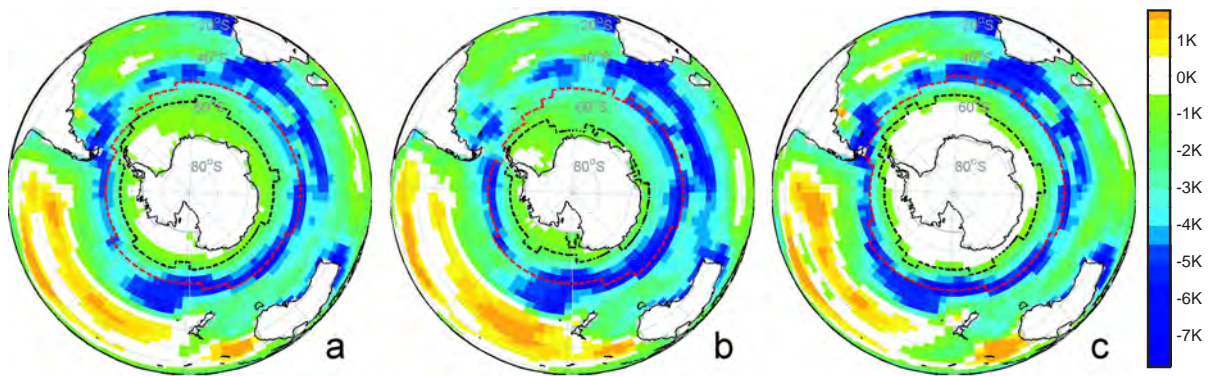


Figure 2: Shading shows the sea surface temperature anomaly (K) obtained by differencing GLAMAP LGM from a twenty year mean of present day HadISST temperatures (1980-1999) for (a) annual, (b) austral summer, and (c) austral winter conditions. Dashed lines show the 50 % sea-ice concentration boundaries for GLAMAP in red and present day in black.

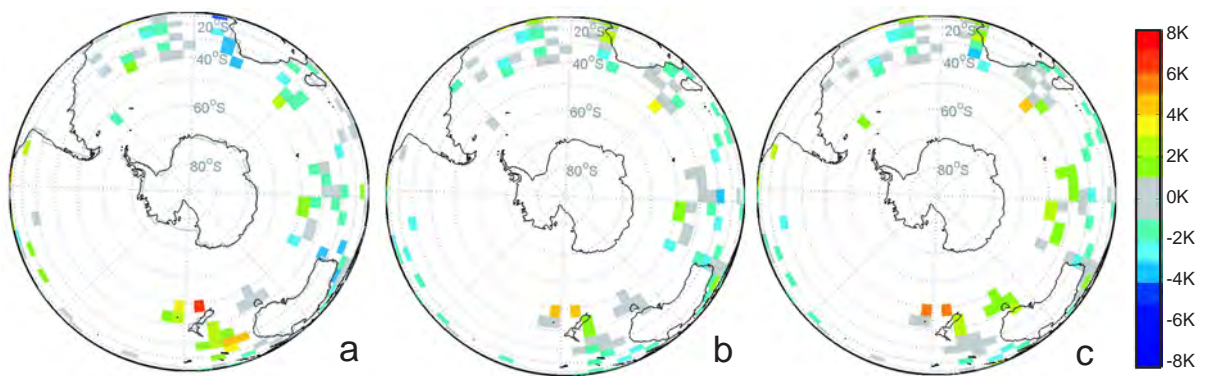


Figure 3: Shading shows the (a) annual mean GLAMAP; (b) PMIP2 HadCM3; and (c) PMIP2 MIROC3.2; each minus the MARGO LGM sea surface temperatures observations. Results are plotted for the 5 x 5 deg grid positions where MARGO observations are available. Grey shows a difference of less than 1 K, white shows regions where no MARGO data exists.

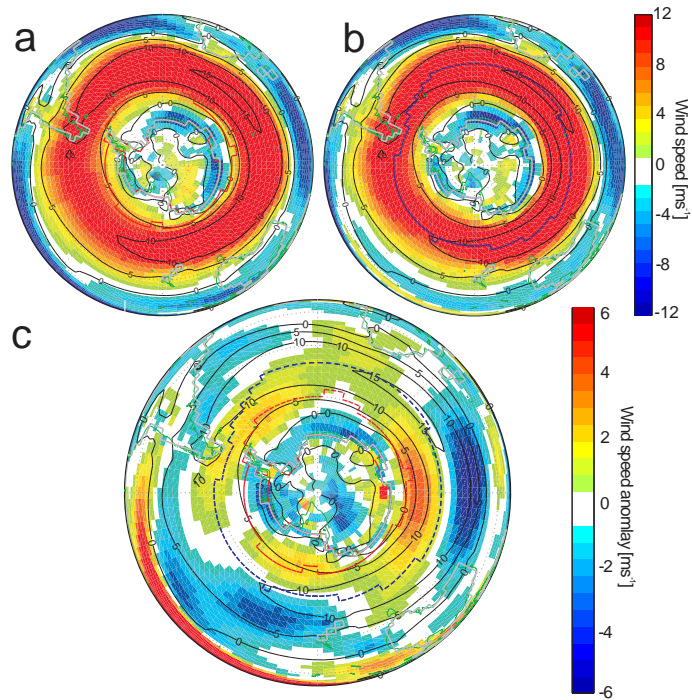


Figure 4: Changes between simulated pre-industrial and glacial Southern Hemisphere westerlies. Shading on panel (a) shows pre-industrial and (b) LGM 850 hPa mean annual westerly wind speed. Shading on panel (c) shows LGM minus pre-industrial 850 hPa wind speed. Red (blue) dashes on each panel show the pre-industrial (LGM) 50% sea-ice edge. To facilitate inter-simulation wind comparison, black contours on each panel indicate simulated pre-industrial westerly wind speed.

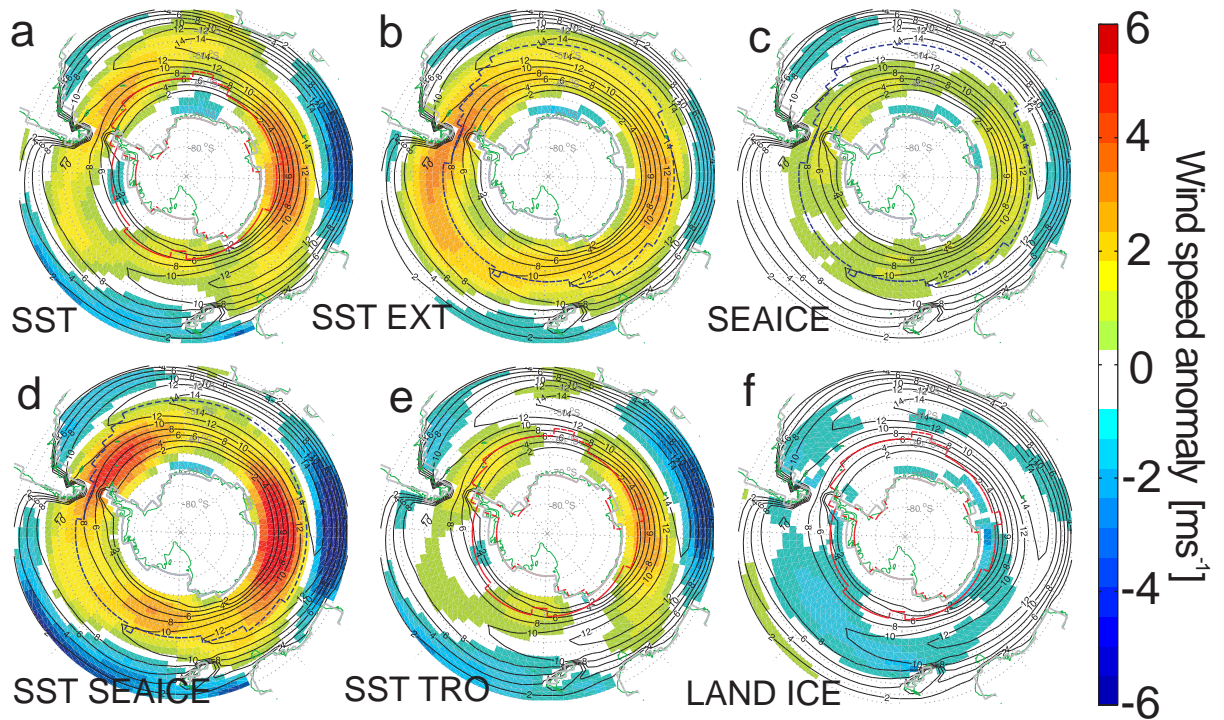


Figure 5: Shading shows the anomalous annual mean 850 hPa westerly wind speed (m s^{-1}). In each case the anomaly shown is glacial simulation minus pre-industrial. The simulations are: (a) sea surface temperature (SST); (b) extra-tropical sea surface conditions (SST EXT); (c) sea-ice (SEAICE); (d) sea surface temperature and sea-ice (SST SEAICE); (e) tropical sea surface temperature (SST TRO); (f) and ice-sheet morphology (LAND ICE). See Table 1 for simulation details. Black contours show the pre-industrial westerly 850 hPa wind speed (at 2 m s^{-1} intervals). Red (blue) dashes show the pre-industrial (LGM) 50% sea-ice edge.

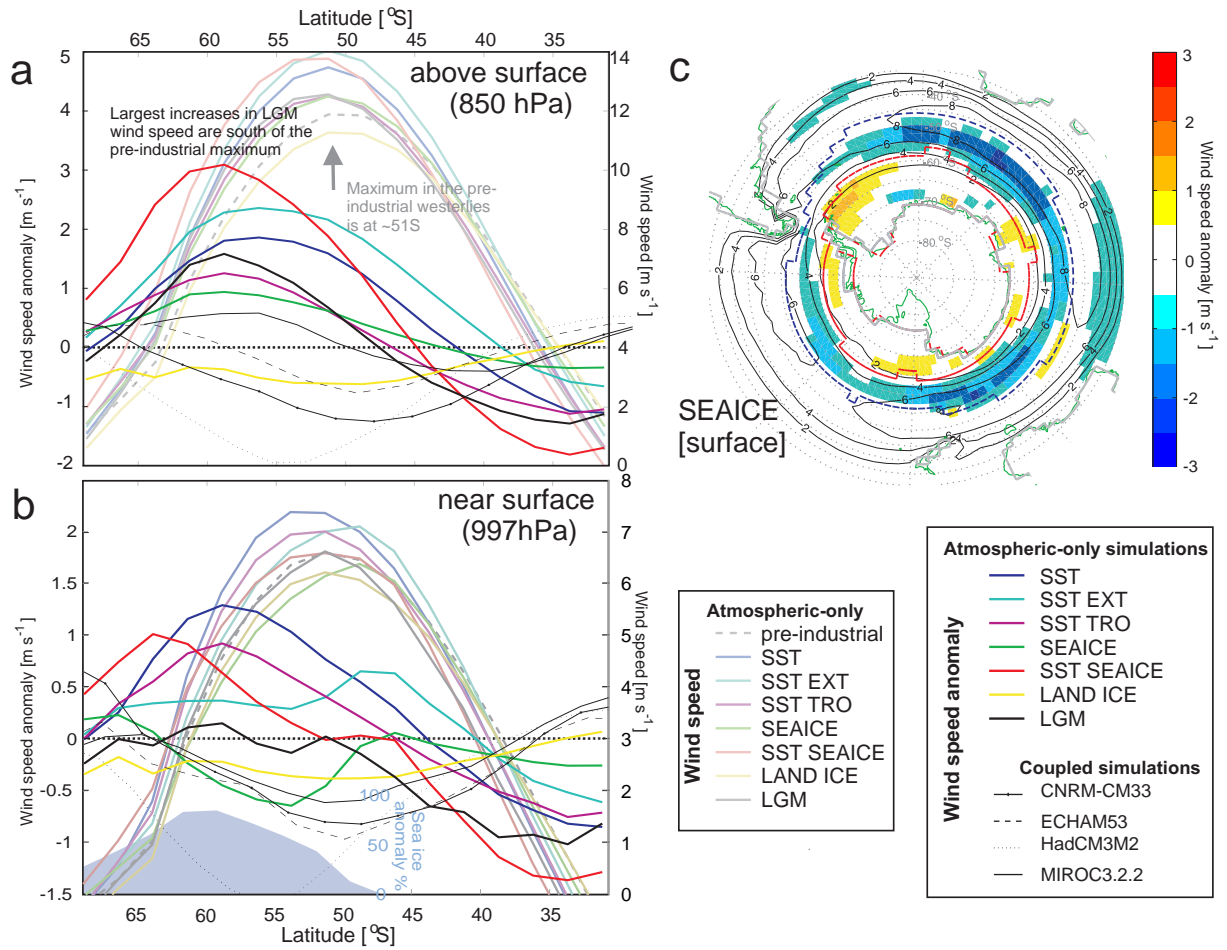


Figure 6: The anomalous annual mean speed (m s^{-1}) on the left axis. For each model this is calculated as glacial simulation minus pre-industrial. Zonal annual mean speed (m s^{-1}) values from each simulation are on right axis. (a) 850 hPa, and (b) lowest model level (30 m) westerly winds. Anomalous PMIP2 annual mean speed results are shown using thin lines. (c) Shading shows anomalous annual mean surface westerly wind speed (SEAICE minus pre-industrial). Contours show the pre-industrial surface wind velocity. Red (blue) dashes show the pre-industrial (LGM) 50% sea-ice edge.

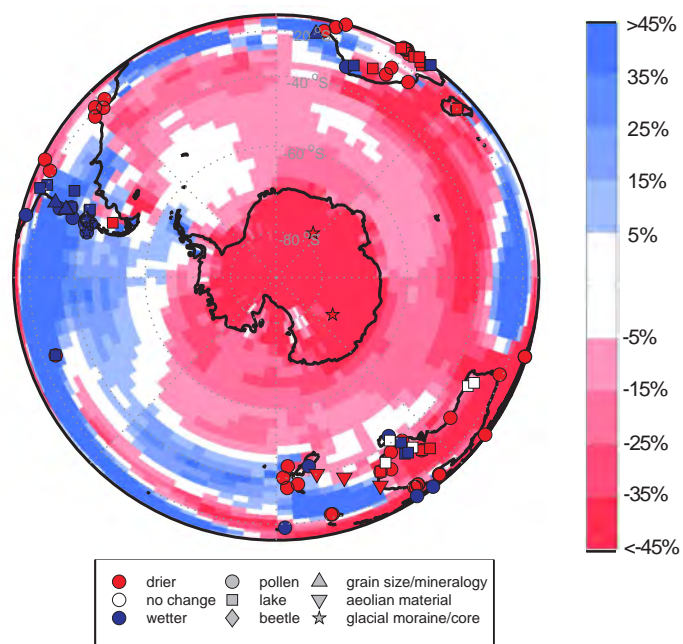


Figure 7: Changes between simulated pre-industrial and glacial precipitation. Kohfeld et al. (submitted) compilation changes in paleo observations of moisture (symbols) and percentage change in precipitation between the LGM and the pre-industrial simulation (LGM minus pre-industrial divided by pre-industrial).

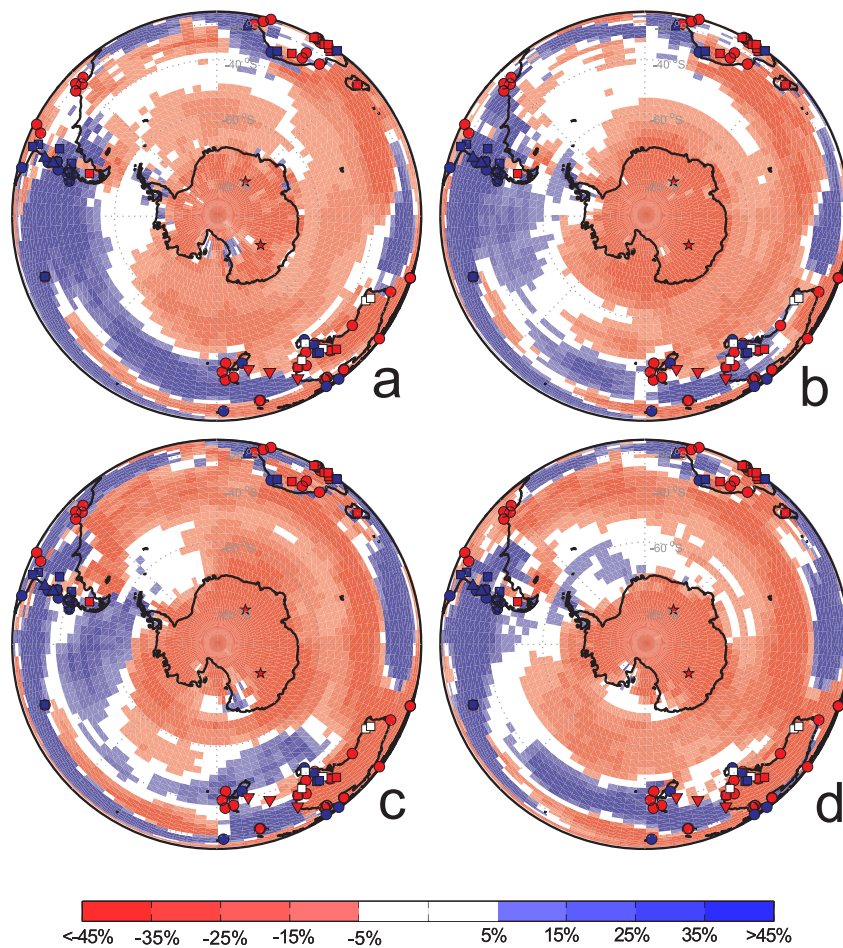


Figure 8: Changes in the moisture fluxes in published studies of paleo-data (symbols) and percentage change in precipitation between the LGM and the pre-industrial simulation, for individual seasons. (a) Austral summer (DJF); (b) austral autumn (MAM); (c) austral winter (JJA); (d) austral spring (SON).

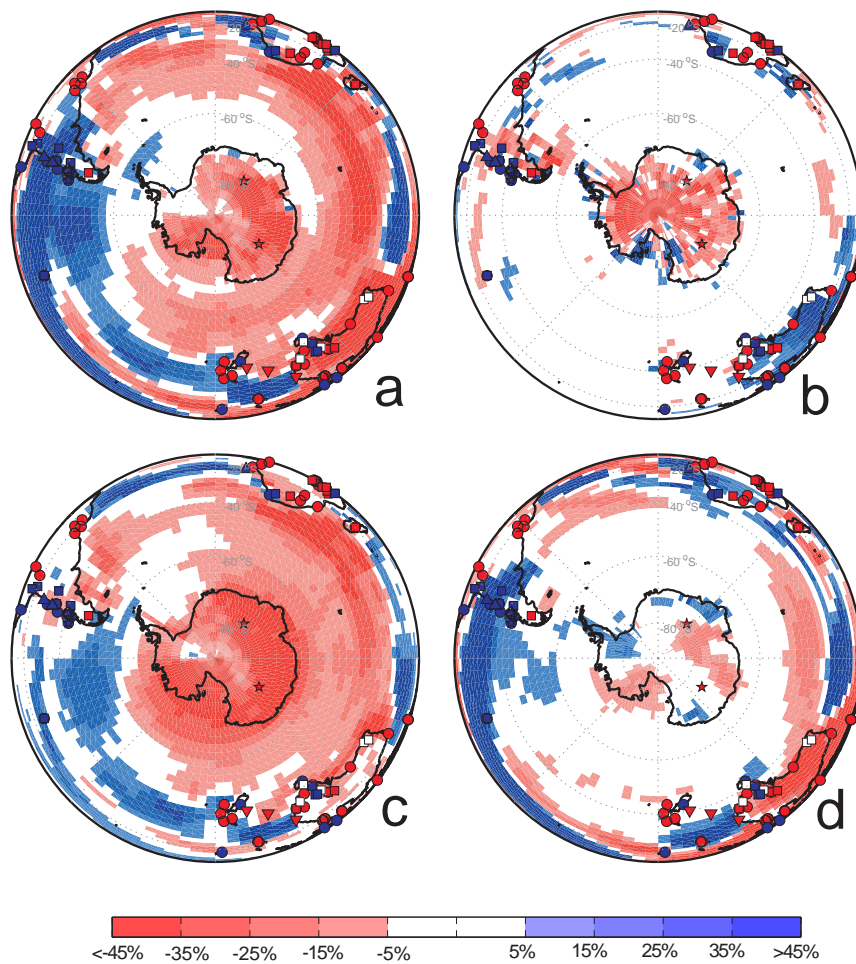


Figure 9: Changes in the moisture fluxes in published studies of compiled paleo-observations (key on Fig. 7) and simulated change in precipitation between the LGM and the pre-industrial simulation (shading). Red is drier, blue is wetter in 10% shading steps, $\pm 5\%$ is white. Simulations are (a) SST; (b) LAND ICE; (c) SST TRO; and (d) SST EXT.

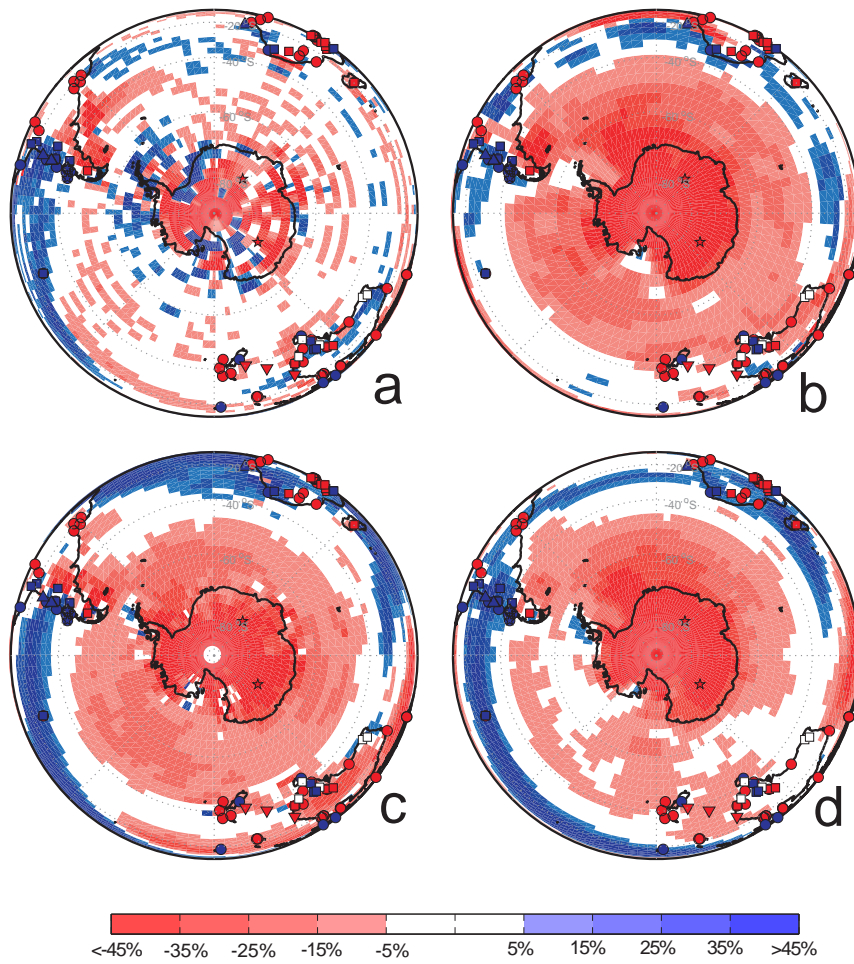


Figure 10: As Fig. 9, but for AOGCM PMIP2. PMIP2 model experiments are (a) CNRM-CM33; (b) ECHAM5.3; (c) HadCM3M2; and (d) MIROC3.2.2.

989 **Appendix A. Additional AGCM sensitivity experiments: Assess-**
990 **ing the impacts of sea surface condition uncertain-**
991 **ties**

992 In addition to the set of core HadAM3 AGCM simulations, and PMIP2
993 AOGCM multi-model analysis presented (Table 1), additional model simu-
994 lations were performed to check the sensitivity of modelled wind changes to
995 a wide range of colder sea surface temperatures changes. In each case the
996 sensitivity experiment is identical to the control pre-industrial experiment
997 except with the application of some idealised sst or sea-ice changes (Fig.
998 A.11). The simulations are run for long enough to allow the atmosphere to
999 reach an equilibrium state with the specified boundary conditions. Results
1000 are presented as differences between the pre-industrial and LGM simulation
1001 for each individual simulation.

1002 The main TRO and EXT experiments (Table 1) are variations of the
1003 SST SEAICE simulation. Here we provide an extended sensitivity set based
1004 directly on these core simulations. Different fractions of the cold TRO and
1005 EXT sst anomaly fields are applied to generate a suite of sensitivity simu-
1006 lations (Fig. A.12 and A.13). The results indicate that wind changes tend
1007 to scale with the size of the temperature anomaly applied. This implies that
1008 errors in the overall magnitude of the GLAMAP sea surface temperature
1009 reconstruction will not affect our main findings.

1010 We also go outside this range and impose very idealised sea surface tem-
1011 perature anomalies. Fig. A.14 shows the results from three such uncon-
1012 strained simulations. These extra simulations help check the sensitivity of
1013 our results to very large GLAMAP sea surface temperature errors. These
1014 simulations were obtained by imposing: global; tropical (within 20° of the
1015 equator, and linearly scaled to zero by 30°); and zonally uniform, but with a
1016 simply linear gradient between the equator and the poles, sea surface temper-
1017 ature anomalies (Fig. A.12). The anomalies applied are zonally uniform in
1018 each case. Additionally, a sensitivity simulation is run to check the impact
1019 on winds of sea-ice extent, and surrounding sea surface temperature (Fig.
1020 8d). This experiment is like the SEA ICE set-up, however here sea-ice is
1021 extended by specifying sea-ice where sea surface temperatures are below 0°C
1022 (rather than the standard -1.8°C).

1023 Results indicate that wind changes due to tropical cooling effects are not
1024 controlled by the more uncertain details of the cooling pattern. In each case
1025 where the tropics are cooled, the Southern Hemisphere westerly winds moves

1026 a little poleward (strengths decrease on the equatorward side of the band and
1027 increase on the poleward side of the band). Thus present LGM sst uncer-
1028 tainties in tropical areas should have a rather limited impact on results. The
1029 results from the uniform global cooling simulation show that uniform cool-
1030 ing tends to decrease wind strengths. This agrees with the hypothesis that
1031 the overall atmospheric temperature generally controls the overall strength
1032 of the atmospheric circulation. Since the small wind strength decreases have
1033 little geographical structure, this implies that errors in the overall cooling
1034 will also have a limited affect on the simulated wind changes. The idealised
1035 gradient change experiment results show more geographical structure in wind
1036 changes. This confirms that modelled glacial-interglacial wind changes are
1037 sensitive to large errors in meridional temperature gradients (Fig. A.14c).
1038 This implies that remaining uncertainties in the position and strength of the
1039 glacial Southern Hemisphere westerlies are mainly due to a lack of LGM
1040 observations of sea surface conditions over the Southern Ocean. The final
1041 sea-ice sensitivity experiment result, showing increases in 850 hPa wind speed
1042 in this region where sea-ice is allowed to persist over warmer waters, confirms
1043 the conclusion that extended sea-ice coverage drives a small intensification
1044 of the westerlies which is largest close to the extended sea-ice region.

1045 **Appendix B. An alternative view of mean annual moisture changes**
1046 **(Fig. 7)**

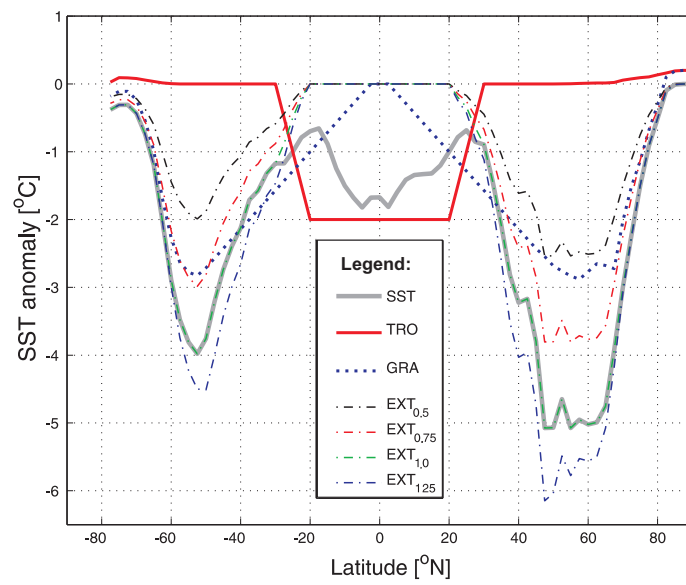


Figure A.11: Mean annual zonal SST anomalies applied to generate SST sensitivity experiments.

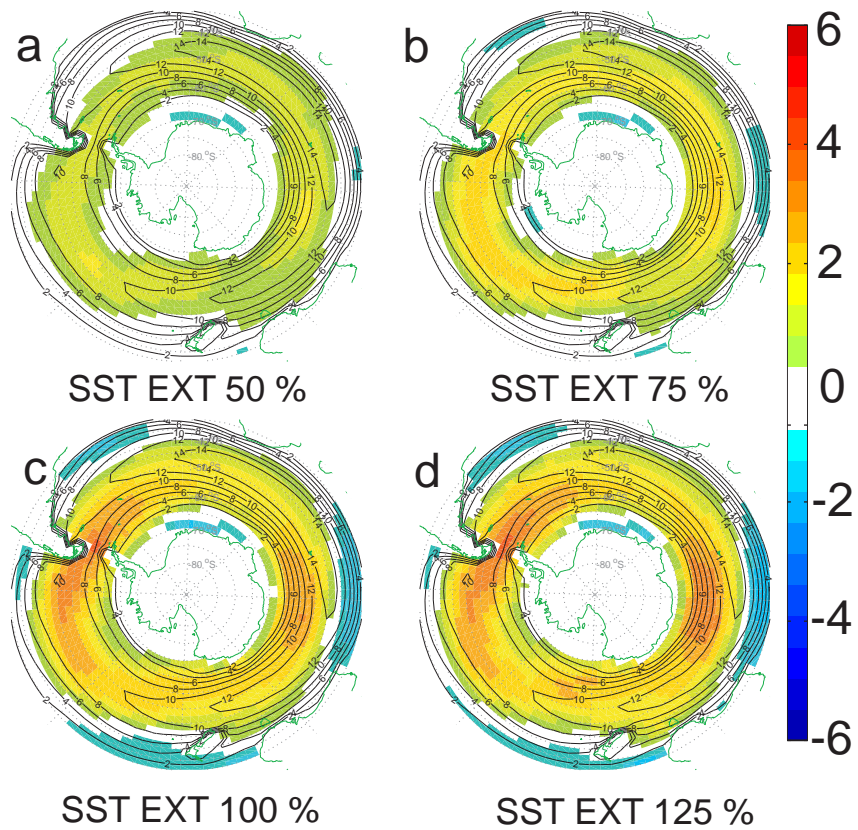


Figure A.12: Differences between 850 hPa pre-industrial and sensitivity study Southern Hemisphere westerlies. Different percentages of the GLAMAP tropical SST anomaly are applied in each experiment (a) 50%, (b) 75%, (c) 100%, and (d) 125%. (These sensitivity experiments are based on the GLAMAP compilation of observations.)

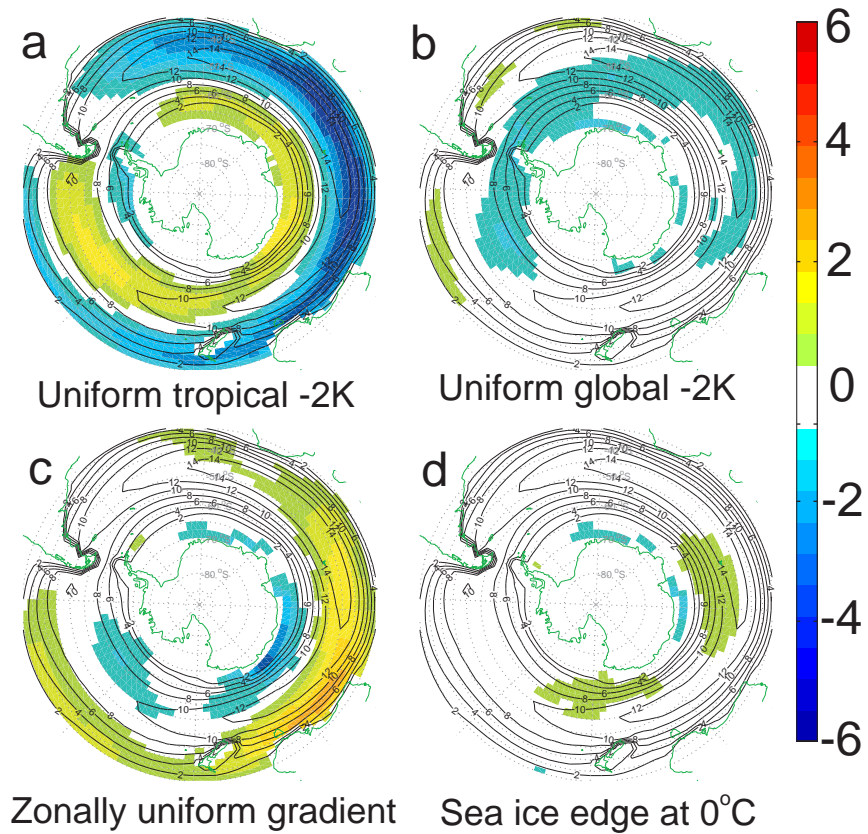


Figure A.13: Differences between 850 hPa pre-industrial and sensitivity study Southern Hemisphere westerlies. Different percentages of the GLAMAP extra-tropic SST anomaly are applied in each experiment (a) 50%, (b) 75%, (c) 100%, and (d) 125%. (These sensitivity experiments are based on the GLAMAP compilation of observations.)

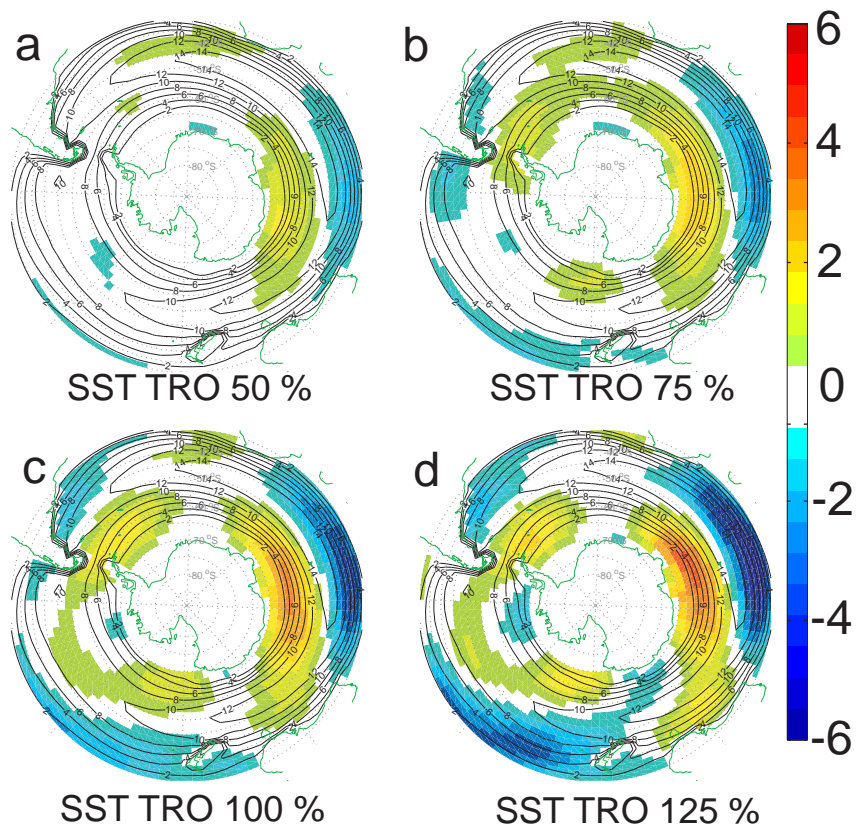


Figure A.14: Differences between 850 hPa pre-industrial and sensitivity study Southern Hemisphere westerlies. Panel (a) a uniform 2 K tropical cooling; (b) a uniform 2 K global cooling; (c) a zonally uniform, but meridionally graded, cooling (see Fig. S5); (d) specifying sea-ice for all regions with SSTs less than 0°C . (These curiosity driven experiments are not based on observations.)

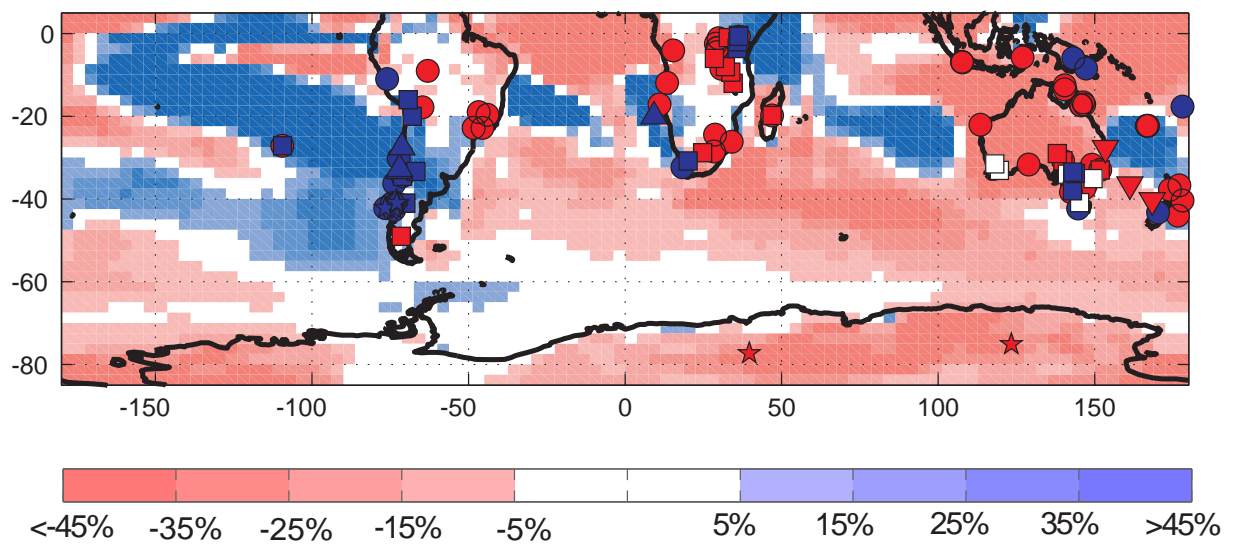


Figure B.15: Changes in the moisture fluxes in published studies of paleo-data (symbols) and percentage change in precipitation between the LGM and the pre-industrial simulation. This is an alternative view of the main Fig. 7 map (using a different projection).