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Marianne Bügelmayer-Blaschek, Didier M. Roche, Hans Renssen, John T. Andrews

1 Internal ice - sheet variability as source for the
2 multi-century and millennial-scale iceberg events
3 during the Holocene? A model study

4
5 February 16, 2016

6 **Abstract**

7 The climate of the Holocene, the current interglacial covering the past
8 11,700 years, has been relatively stable compared to previous periods.
9 Nevertheless, repeating occurrence of rapid natural climate changes that
10 challenged human society are seen in proxy reconstructions. Ocean sed-
11 iment cores for example display prominent peaks of enhanced ice rafted
12 debris (IRD) during the Holocene with a multi-decadal to millennial scale
13 periodicity. Different mechanisms were proposed that caused these en-
14 hanced IRD events, for example variations in the incoming total solar
15 irradiance (TSI), volcanic eruptions and the combination of internal cli-
16 mate variability and external forcings. We investigate the probable mech-
17 anisms causing the occurrence of IRD-events over the past 6,000 years
18 using a fully coupled climate - ice-sheet - iceberg (*i*LOVECLIM) model.
19 We performed 18 experiments that differ in the applied forcings (TSI, vol-
20 canic) and the initial atmospheric conditions. To explore internal ice sheet
21 variability one further experiment was done with fixed climate conditions.
22 All the model runs displayed prominent peaks of enhanced iceberg melt
23 flux (IMF), independent of the chosen experimental set-up. The spectral
24 analysis of the experiments with the ice-sheet - climate model coupled
25 displays significant peaks at 2,000, 1,000 years in all the experiments and
26 at 500 years in most runs. The experiment with fixed climate conditions
27 displays one significant peak of about 1,500 years related to internal ice
28 sheet variability. This frequency is modulated to 2,000 and 1,000 years
29 in all the experiments with a coupled climate - ice sheet due to interac-
30 tions between the climate components. We further investigate the impact
31 of minimum TSI events on the timing and occurrence of enhanced IMF.
32 In the experimental set-up that was forced with idealized sinusoidal TSI
33 variations ($\pm 4 \text{ W m}^{-2}$), we find a significant occurrence of an increased
34 iceberg melt flux about 60 years after the minimum TSI value. Yet, we also
35 see a significant time lag of 80 years between reconstructed TSI minima
36 and the simulated enhanced iceberg melt flux in some of the experiments
37 without TSI forcing. The fact that also model runs that are not forced with
38 TSI variations display an 80 year time lag indicates that the relationship
39 between TSI and IMF is due to internal dynamics of the coupled system.

40 From our experiments we conclude that internal ice sheet variability
41 seems to be the source of the multi-century and millennial-scale iceberg
42 events during the Holocene.

43 **1 introduction**

44 The climate of the past 11,700 years, the Holocene, did not experience strong
45 changes compared to previous periods. This provided the possibility for hu-
46 mans to establish our current society (Wanner et al., 2011) that is now facing
47 unprecedented challenges due to a fast changing climate. But even during the
48 relatively stable Holocene, there have been rapid natural climate changes (e.g.
49 Denton and Karlén, 1973; Wanner et al., 2008, 2011, 2015; Walker et al., 2012)
50 that coincide with disruptions in human society, illustrating the impact of the
51 prevailing conditions and changes therein on humans (Mayewski et al., 2004).
52 Evidence of changing climate conditions, such as temperature and precipita-
53 tion, during the past is found in proxy reconstruction made from tree rings, ice
54 cores as well as ocean sediments (e.g. Bond et al., 1997, 1999, 2001; Bianchi and
55 McCave, 1999; Andersen et al., 2004).

56 In ocean sediment cores the occurrence of Ice Rafted Debris (IRD) indicate
57 that icebergs floated over a specific site and melting in the deposition of sedi-
58 ment. The increase in IRD of a certain size fraction (Andrews et al., 2000) in
59 a sediment core is used as a proxy of enhanced iceberg melt flux. Moreover,
60 by analyzing the mineralogy of the IRD the most probable calving location can
61 be identified (Andrews et al., 2014). The quantity of IRD found in one sedi-
62 ment core depends first on the amount of icebergs calved from the ice sheet,
63 second on the prevailing ocean and wind currents that transport the bergs and
64 thirdly on the melting history (Ruddiman, 1977). Cores taken south and east
65 of Greenland display periods of enhanced IRD during the last glacial as well
66 as the Holocene with a multi-century to millennial scale periodicity (e.g. An-
67 drews et al., 1997; Andrews, 2009; Andrews et al., 2014; Jennings et al., 2002;
68 Bendle and Rosell-Melé, 2007). Moreover, Bond et al. (1997, 2001) presented
69 IRD records from cores located in the North Atlantic, which show periodicities
70 of 1,500 years. However, there is some debate whether these records display
71 changes in IRD due to changed iceberg transport or rather reflect for example
72 pulses in overflow water (Andrews et al., 2014). In addition, Obrochta et al.
73 (2012) questioned if the 1,500 year cycle reported by Bond et al. (1997, 2001)
74 may be a transient signal instead of a physically caused cycle.

75 Further evidence of colder and drier conditions as well as periods of advanced
76 glaciers has been found in global proxy data for the Northern and Southern
77 Hemispheres (Mayewski et al., 2004; Wanner et al., 2008, 2011). These changes
78 are thought to be driven by a globally active mechanism, or a chain of mecha-
79 nisms that affect both hemispheres synchronously and coincide with enhanced
80 IRD fluxes in the Northern Hemisphere (Wanner et al., 2011).

81 A possible mechanism was proposed by Bond et al. (2001), who related
82 periods of increased IRD flux, based on hematite-stained grains recorded in

83 four ocean cores around Greenland, to periods of decreased total solar irradiance (TSI), using the production rate of cosmogenic nuclides Carbon-14 and
84 Beryllium-10 as proxies of TSI. They found a significant correlation between
85 the external TSI forcing and the increased IRD flux recorded in ocean sediment
86 cores close to the Greenland ice sheet (GrIS) and in the North Atlantic.
87

88 Due to feedback mechanisms within the climate system, a weak external
89 forcing such as variations in the incoming total solar irradiance can have sub-
90 stantial climatic impact. Renssen et al. (2006) conducted a study using a cou-
91 pled ocean-atmosphere-vegetation model and revealed that the probability of a
92 colder ocean state increases with decreasing TSI. Therefore, they concluded that
93 the ocean further enhances the weak TSI forcing. Moreover, modeling studies
94 addressing the dynamical effects of TSI variations on stratospheric ozone indi-
95 cated that negative TSI anomalies also cause colder stratospheric temperatures
96 due to decreased stratospheric ozone formation (Haigh, 1994, 1996). The colder
97 stratospheric temperatures can propagate downward and thereby even change
98 the atmospheric circulation (Haigh, 1994; Shindell et al., 1999, 2001).

99 Another possible mechanism for the occurrence of the Holocene IRD events
100 are strong volcanic eruptions. Large-scale eruptions can cause an average global
101 cooling of 0.1-0.2°C because they decrease the incoming solar radiation at the
102 surface (Robock, 2000). The Northern Hemisphere (NH) is more sensitive to
103 the related radiative cooling due to its larger land surface. During winter,
104 stronger mid-latitude westerlies occur to counteract the enhanced meridional
105 heat gradient that favor positive North-Atlantic-Oscillation (NAO) and Arctic-
106 Oscillation (AO) indices. Therefore, a related warming over the western NH
107 and a cooling over the arctic region and eastern NH is observed (Shindell et al.,
108 1999, 2004). Moreover, (Miller et al., 2012) argued that the cooling effect of
109 volcanic eruptions can be maintained for years after the eruptions due to sea-ice
110 - ocean feedbacks.

111 Recently, (Jungclaus et al., 2010) concluded that multi-centennial climate
112 variation like the Medieval Warm Period might represent internal variability of
113 the climate system rather than a response to TSI variations. The authors used
114 a comprehensive Earth System model to investigate the climate variability over
115 the past millennium using TSI and volcanic reconstructions as forcing fields.
116 Jungclaus et al. (2010) find a clear cooling impact of volcanic eruptions on the
117 global temperatures, but the climate response to variations in TSI is within the
118 same range as internal climate variability. Also Mignot et al. (2011) displayed a
119 stronger model response to volcanic eruptions than to changes in the incoming
120 solar radiation. Moreover, Khider et al. (2014) argue that the millennial scale
121 variations in sea surface temperatures seen in a core from the western tropical
122 Pacific are due to variations in the deep ocean circulation, independent of exter-
123 nal forcings. The authors came to this conclusion by investigating planktonic
124 and benthic foraminifera found in the same core from the western tropical Pa-
125 cific. The planktonic record displays millennial-scale SST oscillations whereas
126 the benthic record shows millennial-scale changes of Upper Circumpolar Deep
127 Water.

128 Other studies proposed that there might not be one sole cause for the various

129 cold events and periods of increased IRD, but rather a combination of different
130 mechanisms, such as freshwater pulses, variations in total solar irradiance as
131 well as the combination of different factors for each event (Mayewski et al.,
132 2004; Wanner et al., 2008, 2011, 2015).

133 Climate models are valuable tools to investigate the impact of various forc-
134 ings, such as TSI or volcanic eruptions on climate as well as to analyze the inter-
135 actions between the different climate components. Only recently, the Past4Future
136 project (<http://www.past4future.eu>) supported the compilation of TSI and
137 volcanic emission reconstructions covering the whole Holocene. Additionally,
138 the PMIP project (<https://pmip.lsce.ipsl.fr/>) compiled the greenhouse
139 gas (GHG) forcing, thus providing the possibility to perform transient experi-
140 ments over the past thousands of years with all the external forcings included.

141 Therefore, we use the earth system model of intermediate complexity *i*LOVECLIM,
142 a fully coupled climate - cryosphere model (Roche et al., 2014; Bügelmayer et al.,
143 2015) that dynamically computes iceberg calving and transport, to perform long
144 - term simulations to address the following questions:

145 **(1)** Can we reproduce the periods of increased iceberg flux observed in ocean
146 sediment cores close to Greenland and in the North Atlantic? **(2)** Can we
147 determine the mechanism of these increased iceberg events? To answer these
148 questions, we concentrated on the last 6,000 years and performed 1 experiment
149 where we prescribed periodic variations in TSI and volcanic forcing, 15 ensemble
150 experiments that varied in the strength of the TSI forcing prescribed, as well as
151 3 additional experiments without any volcanic forcing.

152 In the presented paper, we will first explain in detail the climate model and
153 forcings used as well as the experiments performed, continue with our results
154 and finish with the discussion and conclusions.

155 **2 Methods**

156 **2.1 Climate Model *i*LOVECLIM**

157 The *i*LOVECLIM climate model includes the atmospheric model ECBilt (Op-
158 steegh et al., 1998), a quasi-geostrophic, spectral model calculated on a horizon-
159 tal T21 truncation (5.6° in longitude/latitude) and three vertical pressure levels
160 (800, 500, 200 hPa). The atmospheric variables, e.g. precipitation and tem-
161 perature, are computed every four hours. Precipitation is only incorporated in
162 the lowermost level depending on the available humidity. The vegetation model
163 VECODE (Brovkin et al., 1997) is computed on the same grid as ECBilt, but
164 fractional use of one grid cell is allowed due to the small spatial changes in vege-
165 tation. The vegetation (tree, grass or bare soil) depends on the temperature and
166 precipitation as provided by ECBilt. Further, the ocean model CLIO consists of
167 a dynamic - thermodynamic sea-ice model (Fichefet and Maqueda, 1997, 1999)
168 coupled to a 3D ocean general circulation model (Deleersnijder and Campin,
169 1995; Deleersnijder et al., 1997; Campin and Goosse, 1999). The oceanic vari-
170 ables, sea temperature and salinity, are computed daily and on a 3x3° latitude

171 longitude grid. CLIO consists of 20 unevenly spaced vertical layers. The com-
172 putation of the albedo of sea ice depends on its state (frozen or melting) and
173 the thickness of the snow and ice cover (Goosse et al., 2010). The free surface
174 of the ocean model allows the use of real freshwater fluxes.

175 The ice-sheet model GRISLI (Ritz et al., 1997, 2001) is a three-dimensional
176 thermomechanical ice sheet model first developed for Antarctica and then fur-
177 ther expanded to include the Northern Hemisphere (Peyaud et al., 2007). In this
178 study we concentrate on the Greenland ice sheet, thus only the Northern Hemi-
179 sphere grid is used and the Antarctic ice sheet is prescribed and fixed to present
180 day values. The resolution of GRISLI is 40x40 km on a Lambert azimuthal grid.
181 It predicts the changes in geometry (thickness and extension) of the ice sheet
182 according to the surface mass balance (accumulation minus ablation) and ice
183 flow. The surface mass balance of GRISLI depends on the monthly tempera-
184 tures and the yearly snowfall as computed in ECBilt (Roche et al., 2014). Three
185 different glaciological conditions are taken into account, namely inland ice, ice
186 streams and ice shelves. The inland ice is computed using the 0-order shallow
187 ice approximation (Hutter, 1983; Morland, 1984) whereas the calculation of fast
188 flowing ice streams and ice shelves is based on the shallow shelf approximation
189 (MacAyeal, 1989). Calving occurs whenever the ice sheet thickness at the bor-
190 der of the ice sheet is below 150 m and there is not enough ice coming from the
191 points upstream to maintain the height above the threshold. The total amount
192 of calved ice is accumulated over one GRISLI model year and then given to the
193 iceberg model and can thus add up to more than 150 m. The surface runoff and
194 basal melt are computed at the end of one model year by taking the difference
195 between the ice sheet thickness at the beginning and at the end of the year and
196 considering the mass lost due to calving. After one GRISLI model year, the
197 runoff (surface and basal melt) is incorporated in the land routing scheme of
198 ECBilt and the calving flux is given to the iceberg model (Bügelmayer et al.,
199 2015).

200 In the iceberg model (Bigg et al., 1996, 1997; Gladstone et al., 2001; Jongma
201 et al., 2009, 2013; Wiersma and Jongma, 2010) the yearly calving flux is used
202 to daily generate icebergs according to a monthly distribution (Reid, 2005).
203 Every day icebergs of 10 size classes, as defined by Bigg et al. (1996) based on
204 present day Greenland observations (Dowdeswell et al., 1992), are produced if
205 enough ice mass is available. The icebergs are then moved according to the
206 prevailing oceanic (sea ice drag, water drag and horizontal pressure gradient)
207 and atmospheric (air drag and wave radiation force) conditions. If the icebergs
208 melt, their length to height ratio changes and they are allowed to roll over. The
209 latent heat needed to melt the icebergs is taken from the surrounding ocean
210 and the melt water is applied to the ocean surface (Bügelmayer et al., 2015).
211 In the present model, splitting up of icebergs, refreezing of melt water and the
212 sediment load transported by icebergs are not included.

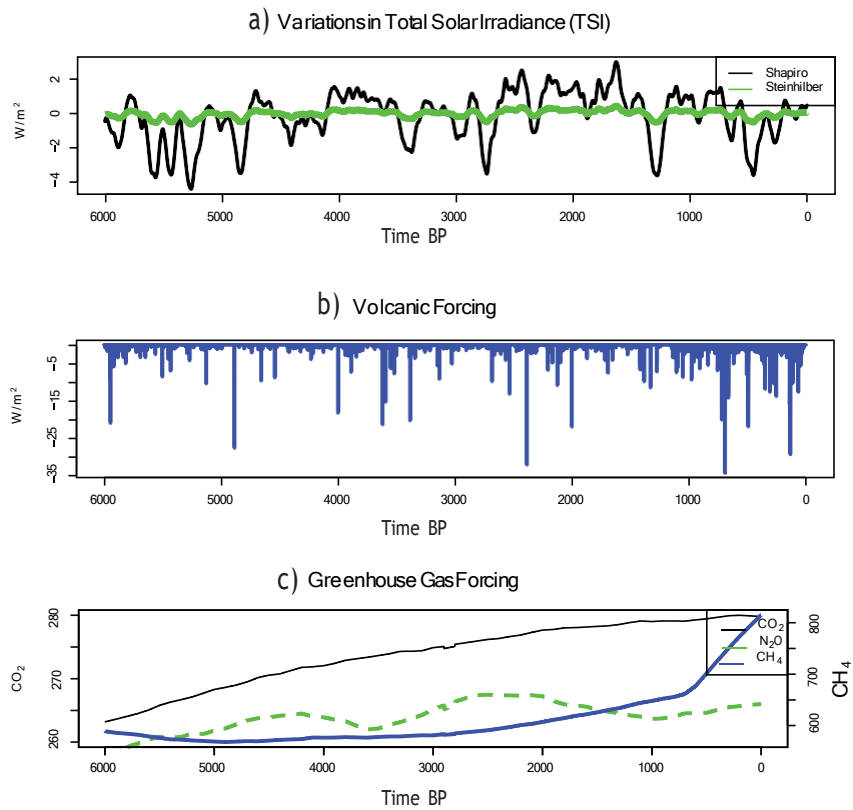


Figure 1: **(a)** 100 year running mean of variations in Total Solar Irradiance (TSI) as reconstructed by Steinhilber et al. (2009), green line and Shapiro et al. (2011), black line; **(b)** Volcanic forcing; **(c)** CO₂, N₂O and CH₄ (=greenhouse gas (GHG) forcing)

213 **2.2 Forcing Fields**

214 We used two TSI reconstructions, which differ in amplitude, but not in the
215 timing of the minima (maxima), to investigate their impact on the modeled
216 Greenland calving flux and icebergs. First, we applied the TSI variations as
217 presented by Steinhilber et al. (2009) who obtained a time series covering the
218 past 9300 years (Figure 1a, green line). In their study, they used the cosmogenic
219 radionuclide ^{10}Be obtained from ice cores to compute the open solar magnetic
220 field that is needed to reconstruct TSI. They find that the TSI varied by ± 1
221 W m^{-2} over the considered time period, thus the mean value of 1364 W m^{-2}
222 was globally increased and decreased by up to 1 W m^{-2} , respectively.

223 Second, we used the TSI reconstruction published by Shapiro et al. (2011)(Fig-
224 ure 1a, black line), who use the same cosmogenic radionuclide ^{10}Be proxy, but
225 compute the solar irradiance by taking into account the active and quiet regions
226 of the sun based on observations of the last years. To reconstruct TSI variations
227 over the past, the ratio between active and quiet sun is scaled with proxies for
228 solar activity. Due to their definition of the quiet sun, their amplitude varies
229 by up to $\pm 5 \text{ W m}^{-2}$ over the past 7000 years. Thus, using the reconstruction
230 published by Shapiro et al. (2011), we alter the mean value of 1364 W m^{-2}
231 globally by up to 5 W m^{-2} .

232 We further prescribed the radiative effect of volcanic eruptions following
233 the reconstructions of Crowley et al. (2008) and Bo Vinther (personal commu-
234 nication, Figure 1b). Large volcanic eruptions pump sulfates into the strato-
235 sphere thereby decreasing the incoming solar radiation. In *i*LOVECLIM vol-
236 canic eruptions are represented by decreased incoming solar radiation, which is
237 latitudinally dependent according to the location of the volcano. This is in con-
238 trast to the globally uniform TSI variations. If both forcings (TSI and volcanic
239 eruptions) are applied the mean value of incoming radiation is altered globally
240 according to variations in TSI and also latitudinally dependent according to
241 volcanic eruptions.

242 Reconstructed CO_2 , N_2O and Methane fluxes (greenhouse gases, GHG) are
243 taken from the PMIP project (Figure 1c). Moreover, the orbital forcing is
244 computed following Berger (1978) inducing a long-term cooling trend over the
245 past 6000 years caused by the decreasing summer insolation over the Northern
246 Hemisphere.

247 **2.3 Experimental set-up**

248 Starting from an equilibrated pre-industrial ice sheet (Bügelmayer et al., 2015),
249 we performed a 3,000 year spin-up period where we applied constant 6kyr BP
250 (6,000 years Before Present (BP)) orbital and GHG values. The pre-industrial
251 ice sheet is a valid initial condition because the Northern Hemisphere ice sheet
252 configuration at 6kyr BP was similar to present day (Vinther et al., 2009) and
253 the ice sheet margin was near or behind the present day margin (Ten Brink and
254 Weidick, 1974; Funder et al., 2011).

255 We then conducted three sets of experiments that differ in the applied TSI

256 forcings. First, 5 ensemble runs of the control simulation without any TSI
257 variations, but with volcanic eruptions (CTRLv) were performed (Table 1).
258 Second, we ran 5 experiments where we applied the TSI variations reconstructed
259 by Steinhilber et al. (2009) (LOWv) and third, we computed 5 runs altered by
260 the TSI variations as defined by Shapiro et al. (2011) (HIGHv, Table 1). The
261 ensemble runs were started from the same ice-sheet and ocean state, but with
262 different atmospheric conditions, which were obtained by performing additional
263 1000 model years after the spin up period and saving the atmospheric conditions
264 every 200 years.

265 We further performed three experiments without including the cooling ef-
266 fect of volcanic eruptions, but with the same TSI forcing as in the ensemble
267 experiments (CTRL, LOW and HIGH, Table 1). This was done in order to
268 investigate the impact of the radiative forcing related to volcanic eruptions on
269 the Greenland ice discharge and the icebergs' distribution.

270 In addition, we tested the impact of an idealized periodic TSI forcing (sinTSI,
271 Table 1). We conducted one run where we applied sinusoidal variations of up to
272 $\pm 4 \text{ W m}^{-2}$ and a periodicity of 400 years. Finally, an experiment was performed
273 where the ice sheet model GRISLI was fed once with a 50 year model climatology
274 and then run for 5000 years without interacting with ECBilt or CLIO to test
275 internal ice sheet variability (fixCLIM, Table 1).

Table 1: Performed experiments

| | | | TSI (Stein- hilber et al., 2009) | TSI (Shapiro et al., 2011) | 400 year Peri- odic TSI forcing | Volcanic Forcing | Initial atmosph. condi- tions | |
|-----------------------|----|---------|--|-------------------------------------|--|---------------------|--|-------------------------|
| S T A N D | 1 | CTRLv-1 | | | | X | - | |
| | 2 | CTRLv-2 | | | | X | 200 | |
| | 3 | CTRLv-3 | | | | X | 400 | |
| | 4 | CTRLv-4 | | | | X | 600 | |
| | 5 | CTRLv-5 | | | | X | 800 | |
| A R D S | 6 | LOWv-1 | X | | | X | - | |
| | 7 | LOWv-2 | X | | | X | 200 | |
| | 8 | LOWv-3 | X | | | X | 400 | |
| | 9 | LOWv-4 | X | | | X | 600 | |
| | 10 | LOWv-5 | X | | | X | 800 | |
| E T - U P | 11 | HIGHv-1 | | X | | X | - | |
| | 12 | HIGHv-2 | | X | | X | 200 | |
| | 13 | HIGHv-3 | | X | | X | 400 | |
| | 14 | HIGHv-4 | | X | | X | 600 | |
| | 15 | HIGHv-5 | | X | | X | 800 | |
| N o V | 16 | CTRL | | | | | - | |
| | 17 | LOW | X | | | | - | |
| | 18 | HIGH | | X | | | - | |
| | 19 | sinTSI | | | X | X | - | |
| | 20 | fixCLIM | | | | X | - | 50yr mean climate |

276 **3 Results**

277 Before explicitly answering the research questions posed in the introduction, we
278 present the modeled climate over the past 6kyr to evaluate the model perfor-
279 mance.

280 **3.1 Modeled Holocene climate**

281 The simulated Holocene climate is characterized by the long-term, orbitally in-
282 duced insolation changes with a decrease in summer insolation in the NH. This
283 causes an overall cooling trend in both air (TAIR) and sea surface tempera-
284 tures (SST) in all the experiments performed (Figure 2a,b). Superimposed on
285 the long-term trend are short-term variations in SST and TAIR that represent
286 internal variability as well as the effect of the applied forcings. Especially in
287 HIGH(v) at 5.5 - 5kyrBP the variations in TSI have a strong response. The
288 effect of the volcanic forcing is for example visible in the comparison of CTRLv
289 and CTRL at 2.3 to 2 kyrBP, where CTRLv displays decreasing air temper-
290 atures whereas TAIR in CTRL is at first not changing and then increasing.
291 In the HIGHv and LOWv set-ups the negative volcanic effect is partly coun-
292 teracted by positive TSI values, thereby decreasing the difference between the
293 experiments. It is interesting to notice that the TAIR in the standard set-up
294 is about 0.2°C lower than in the experiments without volcanic forcing, clearly
295 displaying the response of the model to the decreased radiative forcing caused
296 by volcanic eruptions.

297 **3.2 The simulated iceberg melt flux**

298 In the following section we present the iceberg melt flux instead of the Greenland
299 calving flux because it depends on both the ice discharge and the transport of the
300 icebergs, as does the IRD data from the ocean sediment cores. We thus expect
301 the IMF to be more comparable to IRD than the calving flux and will therefore
302 compare the IMF to the IRD records, since we do not explicitly simulate the
303 (unknown) sediment load of icebergs. Moreover, note that the modeled IMF
304 and ice discharge from the GrIS strongly correlate ($r > 0.96$).

305 **3.2.1 Mechanisms of enhanced iceberg discharge**

306 The computed iceberg melt fluxes (IMF) of all the actively coupled (climate -
307 ice-sheet) experiments display up to 5 distinct phases of increased values during
308 the last 6000 years (Figure 3). The ensemble mean of CTRLv/LOWv/HIGHv
309 displays three major events that peak at around 5000, 3000 and 1250 years
310 BP that are also prominent in the NoV set-up. The bold black line represents
311 the ocean stacked record of four cores published by Bond et al. (2001) (cf.
312 their Figure 2) that is plotted here for reference. Moreover, the detrended
313 magnitude of released freshwater (up to $100 \text{ m}^3 \text{ s}^{-1}$, Figure 3) is similar in all
314 the experiments. The ensemble mean averages over the five ensemble members,

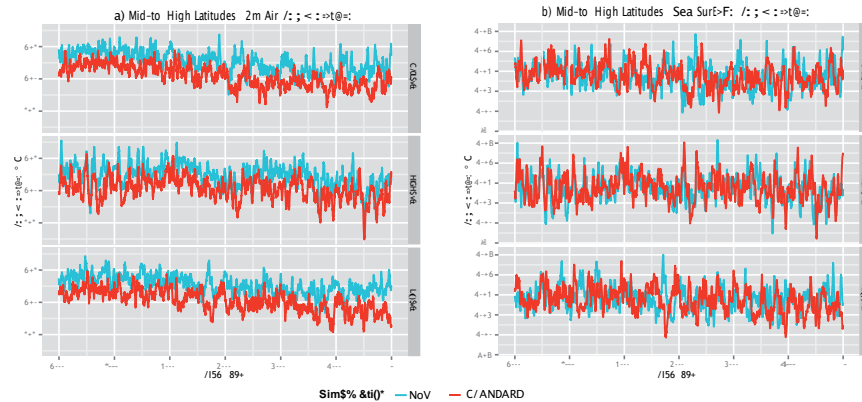


Figure 2: Mid- to High Latitude Area Mean (80° W-15° E, 40° -90° N) of **(a)** 2m Air Temperatures; **(b)** Sea Surface Temperature. The red line corresponds to CTRLv/HIGHv/LOWv-1 and the blue line to CTRL/HIGH/LOW; only the first ensemble member is shown because they share the same starting conditions as the NoV runs.

315 thus smooths the internal variability and displays lower maximum values than
 316 the single runs. To summarize, we don't detect a clear impact of the TSI
 317 or the volcanic forcing on the iceberg melt fluxes. Instead, Figure 3 displays
 318 a coincidence in the timing of the major events of enhanced IMF within the
 319 different experiments.

320 The results of the spectral analysis display significant frequencies at 2,000
 321 and 1,000 years in almost all the experiments (Figure 5a,b,c,d), independent of
 322 the initial atmospheric conditions. This analysis was done in order to gain better
 323 insight on the effect of the forcings on the occurrence of enhanced IMF events.
 324 We used the PAST software (Hammer et al., 2001) to analyze the model data as
 325 well as ocean sediment cores and the model results were averaged over 70 years
 326 to fit to the proxy resolution of Bond et al. (2001). The PAST software includes
 327 the REDFIT spectral analysis based on the REDFIT procedure of Schulz and
 328 Mudelsee (2002). It is a more advanced version of the Lomb periodogram and
 329 includes different windowing techniques as well as an AR(1) red noise model.

330 In addition to the 2,000 and 1,000 years peak, the HIGHv ensemble mem-
 331 bers, NoV and sinTSI experiments display a further peak at 500 years, which is
 332 also seen in most of the CTRLv and some LOWv members (Figure 5). Since the
 333 2,000 year peak can only occur 3 times within 6,000 years, its robustness was
 334 further tested and confirmed by cutting one peak of the IMF time series and re-
 335 peating it 5 times (Thomas Laepple, personal communication).The applied TSI
 336 forcing (Figure 5e) enhances the shorter frequencies at around 300 to 100 years,
 337 especially in the HIGHv runs, but also in the LOWv set-up (Figure 5b,c). The

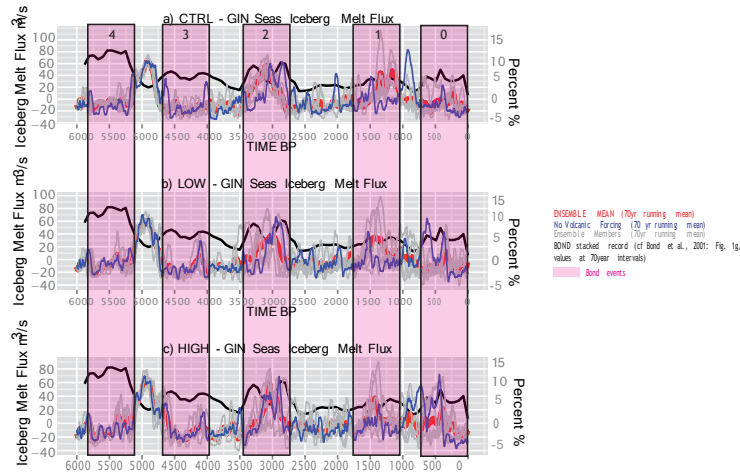


Figure 3: Bold black line is the stacked proxy data (%) as presented by Bond et al., 2001 (cf. Fig. 2g, online available see caption Figure 4); the stacked proxy data can be interpreted as an ensemble mean of the 5 sediment cores used by Bond et al. (2001). The resolution of the proxy data is 70 years, therefore, we computed the 70 year running mean of the modeled iceberg melt flux, taken over the area of 45° W-15° E, 50° -85° N to capture the core locations. The thick red line is the ensemble mean of the 5 ensemble members, the thin grey lines correspond to the individual members, the blue line corresponds to the experiment performed without incorporating volcanic eruptions. The pink bars are the Bond events number 4 to 0; **(a) CTRL**; **(b) LOW**; **(c) HIGH**

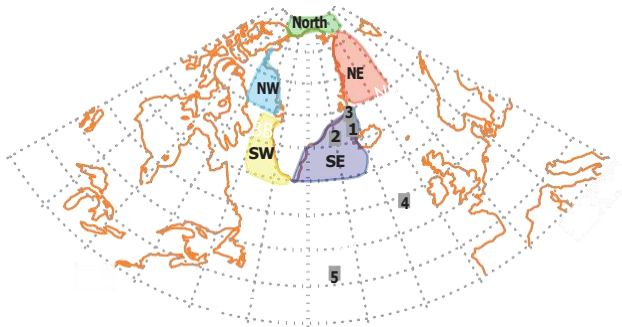


Figure 4: map showing core locations and areas used in figures 5 and 7: North = North Greenland: Lon: 70° W - 0° E, Lat: 80° -85° N ; NE = North-East Greenland: Lon: 25° W - 0° E, Lat: 70° - 80° N; NW = North-West Greenland: Lon: 70° -50° W, Lat: 70° -80° N; SE = South-East Greenland: Lon: 45° -20° W, Lat: 60° -70° N; SW = South-West Greenland: Lon: 60° -45° W, Lat: 60° -70° N; Numbers correspond to ocean cores: 1) JR51GC35; 2) MD99-2269; 3) JM96-1205 at the same site as MD99-2317; 4) VM29-191; 5) KN158-4 GGC22

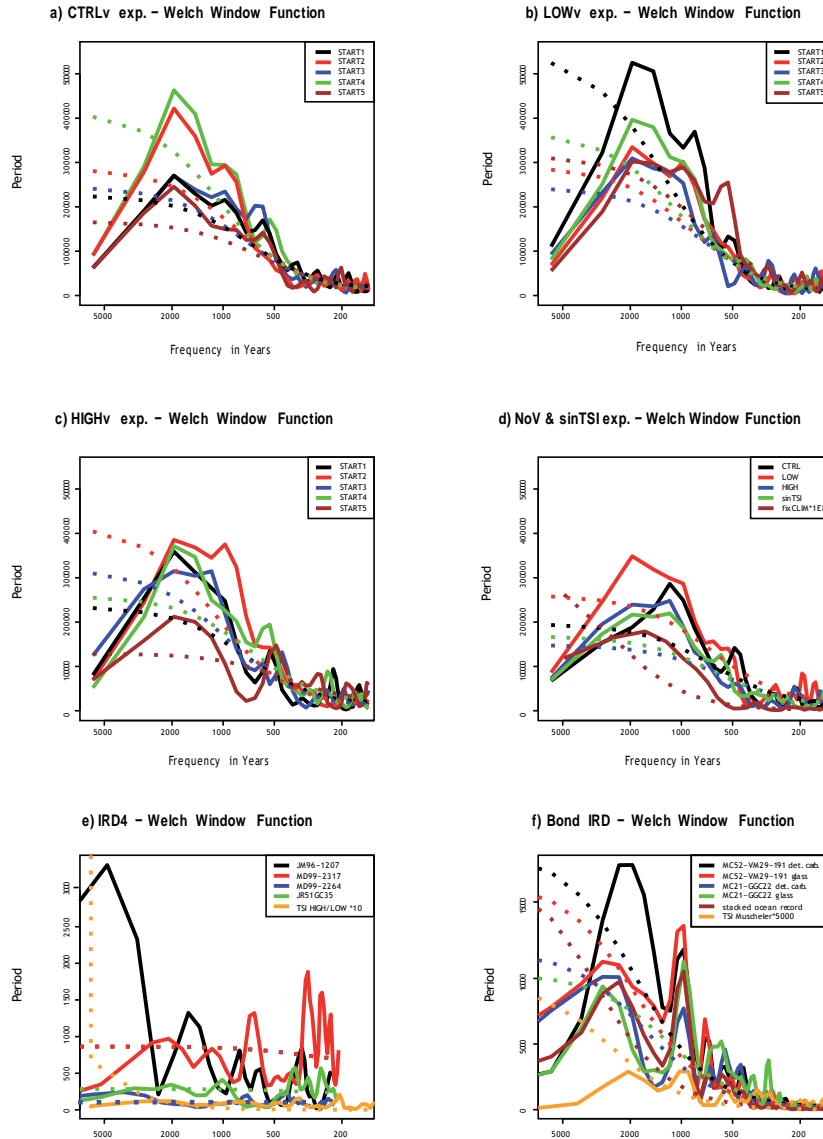


Figure 5: Frequency analysis of model results (70 year mean values because the resolution of the Bond data is 70 years) and IRD data using the PAST software; **(a)** CTRLv members; **(b)** LOWv members; **(c)** HIGHv members; **(d)** CTRL/LOW/HIGH, sinTSI and fixCLIM experiments; **(e)** IRD4: cores JR51GC35, MD99-2269, JM96-1205, MD99-2317 and Past4Future TSI reconstruction; **(f)** Bond IRD: IRD records presented in Figure 2 of Bond et al. (2001) and TSI reconstruction presented in Bond et al. (2001, Fig. 3b); all Bond (2001) data is from ftp://ftp.ncdc.noaa.gov/pub/data/paleo/contributions_by_author/bond2001/bond2001.txt; logarithmic x-axis

Table 2: Significance Tests (95%), the T-Test was only performed for the cases with equal variances (negative F-Test); x=significant; SST = sea surface temperature; T2M = air temperature at 2 m; IMF = iceberg melt flux

| F-TEST (95%) | SST | T2M | IMF |
|--------------|-----|-----|-----|
| CTRLv-LOWv | - | X | X |
| CTRLv-HIGHv | X | X | - |
| LOWv-HIGHv | X | X | - |
| T-TEST (95%) | SST | T2M | IMF |
| CTRLv-CTRL | - | X | X |
| LOWv-LOW | X | X | X |
| HIGHv-HIGH | - | X | - |
| CTRLv1-v2 | - | X | X |
| LOWv1-v2 | X | X | X |
| HIGHv1-v2 | - | - | - |
| CTRLv1-v3 | X | X | - |
| LOWv1-v3 | - | - | X |
| HIGHv1-v3 | - | - | X |
| CTRLv1-v4 | X | - | - |
| LOWv1-v4 | X | X | X |
| HIGHv1-v4 | - | - | - |
| CTRLv1-v5 | - | X | X |
| LOWv1-v5 | - | X | X |
| HIGHv1-v5 | - | - | X |

338 accordance between the experiments is striking because they differ significantly
339 in iceberg melt flux, ocean- and air temperature (Table 2). It is important to
340 notice that the experiment conducted with fixed climate conditions (fixCLIM)
341 reveals one significant spectral peak at about 1,500 years (Figure 5d), which cor-
342 responds to internal ice sheet variability. In the CTRL(v)/ LOW(v)/ HIGH(v)
343 experiments this peak is split into two significant peaks at around 2,000 and
344 1,000 years due to the climate - ice-sheet interactions.

345 The high frequency cycles correspond well to the proxy data (IRD-4, Fig-
346 ure 5e) that is data collected at cores JR51GC35 (Bendle and Rosell-Melé, 2007),
347 MD99-2269 (Moros et al., 2006; Stoner et al., 2007), JM96-1205 (Smith, 2001;
348 Andrews et al., 2010) and MD99-2317 (Jennings et al., 2006, 2011). These cores
349 have been chosen because they cover the Holocene at a reasonable resolution
350 (100 to 25 years). Since the PAST software can handle unevenly spaced data
351 and automatically detrends it, further treatment of the IRD-4 data was not
352 undertaken. Most cores display prominent frequencies at around 400 and 200
353 years, as well as 1,000 years. Moreover, cores JM96-1207 and MD99-2317 dis-
354 play frequencies around 2,000 years, which are prominent in all the model results
355 (Figure 5). Using different software (Andrews et al., 2014) analyzed IRD data
356 from cores MD99-2322, MD99-2264, and JR51GC35 and also noted significant
357 millennial and multi-century scale periodicities in the detrended time-series.

358 The data published by Bond et al. (2001), and online available, also displays
359 significant frequencies at 2,000 and 1,000, as well as 500 years (Figure 5f). The
360 1,000 year spectral peak of Bond et al. (2001) coincides with the TSI reconstruc-
361 tion they used in their study (Figure 5f). This frequency is not prominent in the
362 TSI reconstruction applied in the present manuscript, which exhibits significant
363 spectral peaks at 2,000, 400 and 200 years (Figure 5e). The same analysis was
364 performed using a Hanning window function to test the impact of the chosen
365 window function on the occurring frequencies. The difference between the Han-
366 ning and the Welch window function is that the Hanning window function has
367 a steeper slope towards the end values, thus incorporates less low frequencies
368 (not shown). The results confirm the peaks seen in Figure 5.

369 An analysis of the timing of minimum TSI values and the occurrence of
370 enhanced IMF revealed a significant time lag of about 60 years when applying
371 a strong ($-4W m^{-2}$) idealized sinusoidal shaped forcing (sinTSI experiment,
372 Figure 6a). Using this set-up first, we analyzed the occurrence and exact time
373 of the enhanced IMF within 200 years after a minimum TSI event. In 10 out of
374 14 events the IMF increases strongly that is more than the mean plus two times
375 its standard deviation (Figure 6a, please note that some arrows overlap). Yet,
376 a similar time lag of about 80 years is found in the CTRLv, HIGHv and LOWv
377 experiments when using periods of strongly decreased TSI forcing as starting
378 time (Figure 6b,c,d). This shows that the cyclic forcing alters the exact timing
379 of the increased IMF, but we cannot confirm that the periods of increased IRD
380 recorded in the ocean sediment cores are related to decreased TSI. Especially
381 because a significant time lag of 80 years between periods of minimum TSI and
382 enhanced IMF is found in CTRLv and LOWv, but not in the HIGHv set-up.

383 To summarize, we cannot detect a clear impact of the variations in total solar

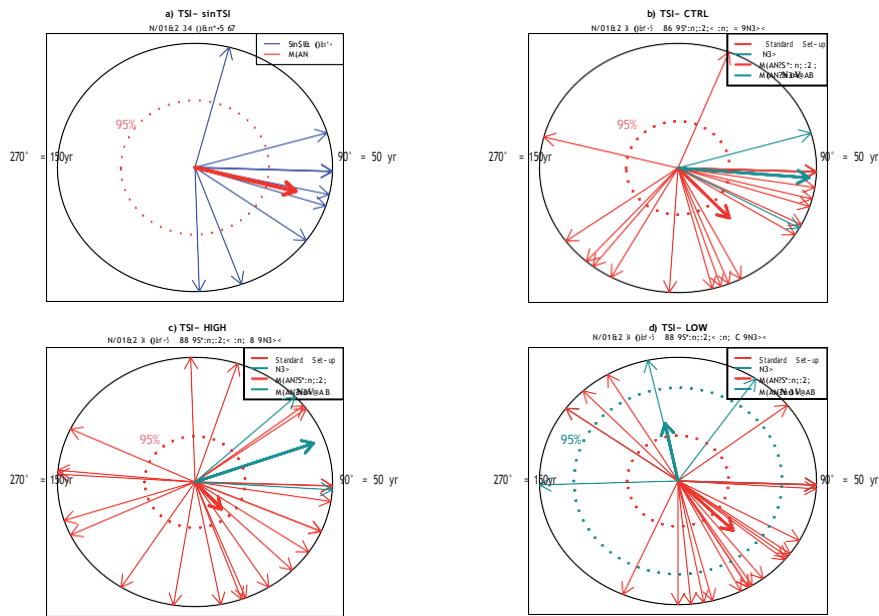


Figure 6: Statistical significance of the phase locking between the applied TSI forcing and the enhanced IMF; the timing of the occurrence of enhanced iceberg melt flux ($IMF(t) > IMF_{mean}(t=1,6000)+2*stdev$) relative to defined years of strongly decreased TSI can be displayed as vectors in a unit circle; the period considered following a minimum TSI forcing are 200 years. The bold vectors represent the averaged vector and display a 95% significant time lag if they exceed the respective dashed circle (color code; note that minimum 4 vectors have to be present to compute a significant averaged direction); **(a)** is the computed phase locking period of the sinTSI experiment, from year 300 up to 5500 every 400 years were taken as starting year, as described above the following 200 years were; **(b)** CTRL; **(c)** LOW; **(d)** HIGH; for b-d: red corresponds to the ensemble members, darkgreen to the experiment without volcanic forcing; the years of minimum TSI forcing considered are: 650, 1100, 2580, 3300, 4700, 5580; please note that some arrows overlap.

384 irradiance, independent of the chosen reconstruction (low or high amplitude),
385 nor can we find an evident impact of volcanic eruptions on the timing of the
386 enhanced ice discharge. All model experiments exhibit significant frequencies
387 of 500, 1,000 and 2,000 years of enhanced IMF that correspond well to IRD
388 data. The 2,000 and 1,000 year frequencies resemble internal ice sheet variability
389 (~1,500 years), modulated by ice-sheet - climate interactions. The impact of
390 variations in total solar irradiance on the IMF is seen in the shorter frequencies,
391 100-400 years, especially when applying the high amplitude forcing.

392 **3.2.2 The Geographic Origin and Spatial Pattern of Enhanced Ice-** 393 **berg Discharge**

394 In the following section we will only present the results of the CTRL set-up
395 because, even though the modeled IMF differs significantly between the experi-
396 ments (Table 2), the overall patterns strongly resemble each other. The plots of
397 the other experiments can be found online in the supplementary information.

398 All the calving sites around Greenland experience periods of enhanced ice
399 discharge over the past 6,000 years, except South - East Greenland (Figure 7).
400 In this region we do not find a recurrent enhanced calving flux, instead the ice
401 discharge of CTRL(v) is relatively constant over time with only one member
402 of CTRLv and the CTRL experiment exhibiting one and two peaks at 4,000
403 and 2,000 years BP, respectively (Figure 7 d). The calving sites situated at the
404 northern and western site of the Greenland ice sheet (Figure 7a,b,c,e) display
405 three strong peaks at 5,000, 3,000 and 1,250 years BP and 'quiet' periods
406 in between. The highest variability is found at the western calving sites of
407 Greenland, where the calving fluxes vary strongly especially over the past 3000
408 years (Figure 7c,e).

409 Even though the calving sites South-East of Greenland do not experience
410 periods of increased ice discharge, the composite maps of periods of enhanced
411 iceberg melt flux display intensification all around the ice sheet, especially east
412 and west of it (Figure 8a). In the Arctic and Greenland Sea the increase is
413 directly related to a strongly increased calving flux at the respective calving
414 sites, yet in the Iceland Sea it is purely due to the transport of the icebergs by
415 the East Greenland current. Also in Baffin Bay the IMF is further strengthened
416 because the East Greenland and the Labrador Current transport the icebergs
417 into this region (Figure 8a). In agreement with Bond et al. (1997), there are
418 more icebergs reaching the core sites of Bond et al. (2001) in the North Atlantic
419 during periods of enhanced ice discharge (Figure 8b), even though there is only
420 a small signal ($1 \text{ m}^3 \text{ s}^{-1}$) in the freshwater flux (Figure 8a).

421 **4 Discussion**

422 We used the global climate model *i*LOVECLIM that includes a fully coupled ice-
423 sheet and iceberg module to perform experiments covering the past 6000 years.
424 This was done in order to investigate whether we can simulate multi centennial

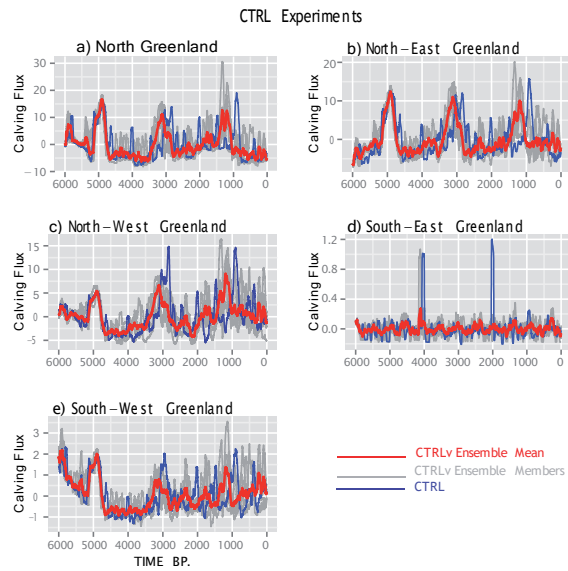


Figure 7: 70 year running mean of the modeled calving flux ($\text{m}^3 \text{s}^{-1}$) from the calving sites at the respective regions; **(a)** North Greenland: Lon: $70^\circ \text{W} - 0^\circ \text{E}$, Lat: $80^\circ - 85^\circ \text{N}$; **(b)** North-East Greenland: Lon: $25^\circ \text{W} - 0^\circ \text{E}$, Lat: $70^\circ - 80^\circ \text{N}$; **(c)** North-West Greenland: Lon: $70^\circ - 50^\circ \text{W}$, Lat: $70^\circ - 80^\circ \text{N}$; **(d)** South-East Greenland: Lon: $45^\circ - 20^\circ \text{W}$, Lat: $60^\circ - 70^\circ \text{N}$; **(e)** South-West Greenland: Lon: $60 - 45^\circ \text{W}$, Lat: $60^\circ - 70^\circ \text{N}$; The thick red line displays the ensemble mean of the 5 ensemble members, the thin grey lines correspond to the individual members and the blue line corresponds to the experiment performed without incorporating volcanic eruptions. Only the CTRL set-up is shown (LOW(v)/HIGH(v) results can be found in the supplementary information).

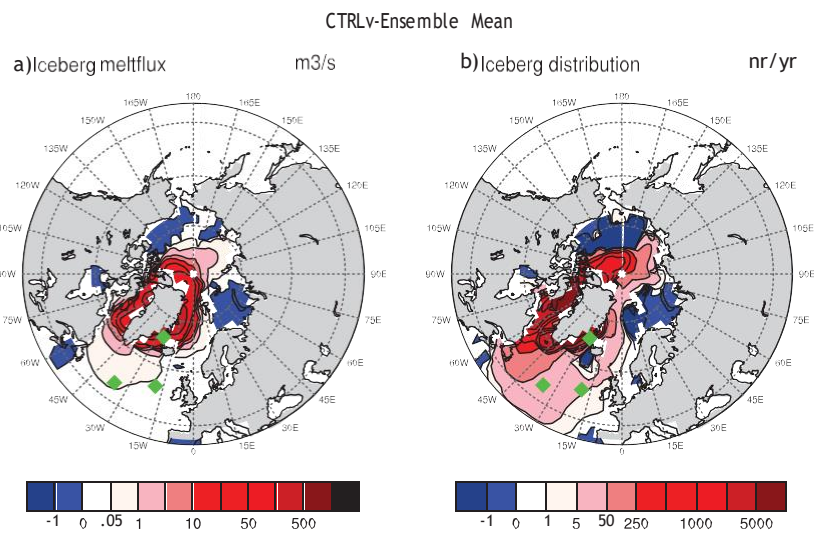


Figure 8: CTRLv-ensemble mean: Composite maps of difference between years of enhanced iceberg melt flux ($IMF(t) > IMF_mean(t=1, 6000) + 2*stdev$) and "quiet" periods ($IMF(t) < IMF_mean(t=1, 6000) + 1*stdev$); **(a)** Iceberg Melt Flux ($m^3 s^{-1}$): **(b)** number of icebergs moving within one grid cell ($nr yr^{-1} cell^{-1}$); non-linear color scheme! Green rectangles correspond to core locations of Bond et al. (1997, 2001). Only the CTRL set-up is shown. LOW(v)/HIGH(v) results can be found in the supplementary information.

425 to millennial scale variations in IRD that occurred during the Holocene and to
426 analyze the underlying mechanism. The modeled calving and iceberg melt flux
427 clearly display periods of increased ice discharge, but these appear independent
428 of the chosen external forcings (TSI, volcanic eruptions).

429 Bond et al. (1997, 2001) claimed variations in incoming TSI responsible for
430 observed variations in IRD thereby causing a lot of debate whether such a small
431 scale forcing could be responsible. Different model studies that investigated the
432 impact of the TSI on climate, which suggested that its effect is further amplified
433 by a stratospheric response (e.g. Haigh, 1994, 1996, 2000; Shindell et al., 2004).
434 The decreased incoming solar radiation causes colder stratospheric conditions
435 due to less ozone formation that consequently alters the atmospheric circulation,
436 resulting in drier conditions in the tropics and a cooling in the mid latitudes.
437 The stratospheric interaction is not taken into account in the model used and
438 might cause an underestimation of the impact of the TSI forcing on climate.

439 Renssen et al. (2006) proposed that the impact of variations in incoming solar
440 radiance is amplified by the ocean's response. They found a higher probability of
441 colder ocean conditions during times of decreased total solar irradiance due to a
442 weakening of the deep convection in the Nordic Seas using the same atmospheric-
443 ocean-vegetation model as incorporated in iLOVECLIM. We did not find a
444 similar response in deep convection, independent of the chosen TSI forcing.
445 This might be due to the fact that we have included an ice-sheet and an iceberg
446 model that stabilize the modeled climate, which was not incorporated in the
447 study of Renssen et al. (2006).

448 Recently, various modeling studies focused on the impact of variations in TSI
449 and volcanic eruptions on the Holocene climate and Mignot et al. (2011) showed
450 a stronger model response to volcanic forcing than to variations in incoming
451 TSI. Further, their results suggest general colder temperatures when including
452 volcanic eruptions, which corresponds well to our findings. Miller et al. (2012)
453 displayed that in the used climate model the cooling effect of volcanic eruptions
454 is maintained even for years after the event.

455 Khider et al. (2014) presented sea surface temperature reconstructions from
456 the western tropical Pacific that exhibit millennial scale variability. Yet, the
457 authors did not find a direct relationship to variations in total solar irradiance
458 or volcanic eruptions and propose that it might be internal variability rather
459 than external forcing factors. This is consistent with our results that the events
460 of increased ice discharge with a periodicity of about 2,000 years occur indepen-
461 dently of the use of TSI variations or volcanic eruptions. Instead, the internal
462 ice sheet variability that experiences significant spectral peaks at about 1,500
463 years is thought to be responsible. Moreover, Wanner et al. (2011) defined six
464 specific cold events covering the Holocene based on global time series of temper-
465 ature and precipitation/humidity. Their cold events only partly coincide with
466 the occurrence of Bond events, suggesting different mechanisms causing the one
467 and the other.

468 The spectral analysis on the modeled iceberg melt flux reveals a clear pref-
469 erence towards frequencies at 2,000, 1,000 and 500 years in all the experiments.
470 The experiment with fixed climate conditions that displays internal ice sheet

471 variability revealed just one peak at approximately 1,500 years, indicating that
472 the long-term frequencies are internally driven. Also other modelling studies
473 indicated the occurrence of internal ice sheet variability that appears to depend
474 on the surface temperatures, accumulation rates and sliding parameters (Payne,
475 1995; Calov et al., 2010). For the ice sheet model used, (Ritz et al., 1997) anal-
476 ysed the impact of various model parameters, such as the sliding coefficient or
477 ablation, on the ice volume, ice extent and maximum altitude. The authors
478 found that the ice volume and maximum altitude are strongly affected by the
479 sliding coefficient, whereas the ice extent mainly reacts to the ablation of snow
480 and ice, under present day climate conditions. Implementing the variations in
481 TSI causes the appearance of shorter frequencies in the range of 100-400 years
482 that might correspond to the de Vries and Gleissberg cycles (~ 210 and ~ 87
483 years, respectively) that were prominent throughout the Holocene (Muscheler
484 et al., 2003; Wagner et al., 2001). The short frequencies are also seen in the
485 IRD data of four ocean sediment cores, as well as peaks at around 1,500, 1,000
486 and 600 years. These are confirmed by Andrews et al. (2014), who noted fre-
487 quencies of 1040, 660 and 370 years at cores taken close to the ones used in
488 this manuscript. We did not incorporate core VM28-14 of Bond et al. (2001) in
489 the study because Andrews et al. (2014) noted that its 1,500 year spectral peak
490 might reflect pulses in overflow waters rather than in ice discharge. Obrochta
491 et al. (2012) revised the data of Bond et al. (1997, 2001) and presented promi-
492 nent peaks at 2,000, 1,000 and 500 years, which support our findings of the
493 model results as well as of the online available data of Bond et al. (2001).

494 The spatial distribution of icebergs and their freshwater flux during periods
495 of increased ice discharge clearly shows that icebergs are able to spread further
496 into the North Atlantic than during 'quiet' periods. Yet, there are only a
497 few icebergs reaching that far and more importantly, we have not incorporated
498 sediment loads in our iceberg model, thus we cannot predict how much sediments
499 would still be in the icebergs at the time of their arrival at the locations of the
500 ocean sediment cores. It is also important to note that during cold periods,
501 leading to the prolonged existence of fast ice and sikkusaqs, that it can take
502 years for icebergs to exit onto the shelf, hence may already have undergone
503 significant melting and loss of sediment (Dwyer, 1995; Reeh et al., 1999, 2001;
504 Reeh, 2004; Syvitski et al., 1996).

505 **5 Conclusions**

506 We simulated the climate of the last 6,000 years using the global climate model
507 iLOVECLIM that incorporates fully coupled ice-sheet and iceberg modules and
508 thus allows for the interactive computation of icebergs. This set-up was used
509 to investigate whether we can model periods of enhanced iceberg melt flux as
510 recorded in ocean sediment cores and to analyze the underlying mechanisms.

511 We have performed 19 experiments that differ in the applied forcings (vari-
512 ations in total solar irradiance and cooling due to volcanic emissions) and the
513 starting conditions (atmosphere). All the experiments display periods of in-

514 creased ice discharge from the Greenland ice sheet, independent of the chosen
515 TSI forcing and the implementation of volcanic eruptions. Moreover, the shifted
516 atmospheric starting conditions do not impact the timing of these events that
517 occur at similar model years in all experiments.

518 The spectral analysis of the iceberg melt flux displays significant peaks of
519 enhanced values in the range of 2,000, 1,000 and 500 years in all the experi-
520 ments. Shorter frequencies between 50 and 400 years are especially prominent
521 in the HIGHv set-up, but also appear in the LOWv runs, indicating the role of
522 variations in TSI on that time scales. Moreover, the ice sheet's internal vari-
523 ability displays events of enhanced ice discharge with a significant periodicity
524 of 1500 years. This spectral frequency is modulated to two significant peaks at
525 2,000 and 1,000 years when actively coupling the climate components to the ice
526 sheet.

527 Moreover, we applied sinusoidal shaped variations in TSI to investigate its
528 impact on the timing of the events of increased ice discharge. In this experiment
529 we find a significant time lag of 60 years between the minimum TSI and the
530 occurrence of the event. Yet, a common time lag of about 80 years between
531 the minimum incoming solar radiation and the occurrence of increased iceberg
532 melt flux is seen in all the CTRL(v)/HIGH(v) and LOW(v) experiments. We
533 therefore conclude that the variations in TSI do not pace the enhanced iceberg
534 flux at the longer time-scales.

535 Overall, we provide strong support that the enhanced IRD fluxes found in
536 ocean sediment cores reflect internal ice sheet variability rather than a response
537 to external forcing. This would also explain the different timing of Holocene
538 cold events as defined by Wanner et al. (2011) and Bond events Bond et al.
539 (2001).

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