



A New Mechanism for Dansgaard-Oeschger Cycles

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A new mechanism for Dansgaard-Oeschger cycles

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14 **Abstract**

15 We present a new hypothesis to explain the millennial-scale temperature variability
16 recorded in ice cores known as Dansgaard-Oeschger (DO) cycles. We propose that an ice
17 shelf acted in concert with sea ice to set the slow and fast timescales of the DO cycle,
18 respectively. The abrupt warming at the onset of a cycle is caused by the rapid retreat of
19 sea ice after the collapse of an ice shelf. The gradual cooling during the subsequent
20 interstadial phase is determined by the timescale of ice-shelf regrowth. Once the ice shelf
21 reaches a critical size, sea ice expands, driving the climate rapidly back into stadial
22 conditions. The stadial phase ends when warm subsurface waters penetrate beneath the
23 ice shelf and cause it to collapse. This hypothesis explains the full shape of the DO cycle,
24 the duration of the different phases, and the transitions between them and is supported by
25 proxy records in the North Atlantic and Nordic Seas.

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31 **1. Introduction**

32 During the last glacial period, the North Atlantic basin experienced a number of
33 large and abrupt millennial-scale fluctuations in climate referred to as Dansgaard-
34 Oeschger (DO) cycles. Ice cores from Greenland reveal that each cycle began with an
35 abrupt warming from stadial to interstadial conditions [*Johnsen et al.*, 1992; *Dansgaard*
36 *et al.*, 1993; *Groote et al.*, 1993; *Huber et al.*, 2006]. The effects of this warming
37 extended across much of the northern hemisphere [*Voelker et al.*, 2002; *Overpeck and*
38 *Cole*, 2006; *Pisias et al.*, 2010], while a near-simultaneous cooling occurred in Antarctica
39 [*EPICA Members*, 2006; *Wolff et al.*, 2010]. Greenland ice core records then suggest
40 gradual cooling during the initial stages of each interstadial phase, followed by abrupt
41 cooling back to stadial conditions.

42 A common explanation for these cycles involves changes in the Atlantic
43 meridional overturning circulation (AMOC), perhaps triggered by freshwater forcing
44 [*Clark et al.*, 2001; *Ganopolski and Rahmstorf*, 2001], but paleoceanographic evidence
45 for these changes remains elusive [*Elliot et al.*, 2002; *Piotrowski et al.*, 2008; *Pisias et*
46 *al.*, 2010]. Here we propose a mechanism to explain these millennial-scale climate cycles
47 involving abrupt changes in sea-ice cover, gradual regrowth of ice shelves, and warming
48 of intermediate-depth waters.

49

50 **2. Rapid Climate Change in Greenland Ice Cores**

51 $\delta^{18}\text{O}$ records from Greenland ice cores show that each DO cycle began with an
52 abrupt shift in $\delta^{18}\text{O}_{\text{ice}}$, occurring in as little as a few years [*Steffensen et al.*, 2008;
53 *Thomas et al.*, 2009], which was associated with a large warming, ranging from 8°C to

54 16°C [Severinghaus *et al.*, 1998; Huber *et al.*, 2006; Wolff *et al.*, 2010 and references
55 therein]. Other properties of the ice, including electrical conductivity [Taylor *et al.*,
56 1993a, 1993b], deuterium-excess [Dansgaard *et al.*, 1989, Steffensen *et al.*, 2008], dust
57 content [Fuhrer *et al.*, 1999], and methane concentrations [Brook *et al.*, 1996] changed in
58 less than a decade. At the same time, accumulation rates roughly doubled and
59 proportionally more precipitation fell in winter months [Alley *et al.*, 1993; Cuffey and
60 Clow, 1997].

61 Following the abrupt warming, the interstadial climate gradually cooled before
62 abruptly cooling back to stadial conditions. A stable stadial climate characterized by low
63 $\delta^{18}\text{O}_{\text{ice}}$ values was then maintained for the next hundreds to thousands of years until the
64 next abrupt warming, concluding the DO cycle. This characteristic trapezoid shape in
65 $\delta^{18}\text{O}_{\text{ice}}$ can be seen for all DO cycles, but their duration varies from ~1.1 to 8.6 kyr
66 (Figure 1A) [Andersen *et al.*, 2006]. Grootes and Stuiver [1997] found a strong peak at
67 1470 years in the power spectrum of DO cycles 1 through 13, but Schulz [2002] showed
68 that most of the power in the 1470-year band came from DO cycles 5-7 only. Due to
69 varying age models and statistical techniques, debate persists over whether a 1470-year
70 periodicity exists in the DO time series [Wunsch, 2000; Rahmstorf, 2003; Ditlevsen *et al.*,
71 2007]. Based on multiple proxy records with DO-like cycles, Piasias *et al.* [2010] found a
72 mode of variability with broad spectral power of ~1600 years rather than a sharp spectral
73 peak at 1470 years.

74 Many climate proxies around the globe show DO-like variability on similar time
75 scales. Proxies from the northern hemisphere show warmer (colder) and wetter (drier)
76 climates during DO interstadials (stadials) [Voelker *et al.*, 2002; Overpeck and Cole,

77 2006; *Pisias et al.*, 2010]. In the Antarctic EDML ice core, there is an inverse relation
78 between northern and southern hemisphere climate oscillations (bi-polar seesaw), with a
79 correlation between Greenland stadial duration and the amplitude of the Antarctic
80 temperature warming [*EPICA Members*, 2006].

81 Sediment cores from 40-50°N in the North Atlantic (the so-called ice rafted debris
82 (IRD) belt) show IRD from Icelandic and European sources associated with every DO
83 stadial [*Bond and Lotti*, 1995], but are dominated by larger IRD pulses from the
84 Laurentide ice sheet known as Heinrich events, associated with only every second to
85 fourth stadial (Figure 1A, 1C) [*Hemming*, 2004 and references therein]. In contrast, in the
86 Nordic Seas [*Voelker et al.*, 1998; *Dokken and Jansen*, 1999] and the Irminger Basin [*van*
87 *Kreveld et al.*, 2000; *Elliot et al.*, 2001], IRD pulses of roughly equal magnitude are
88 visible for every DO stadial, while characteristic Heinrich layers are absent (Figure 1B).
89 Planktonic $\delta^{18}\text{O}$ records show large negative excursions associated with Heinrich events
90 in both the Nordic Seas (Figure 1B) [*Voelker et al.*, 1998; *Rasmussen et al.*, 1996; *Elliot*
91 *et al.*, 1998; *van Kreveld et al.*, 2000] and the IRD belt (Figure 1C) [*Bond et al.*, 1992;
92 *Hillaire-Marcel and Bilodeau*, 2000; *Hemming*, 2004 and references therein], but in the
93 Nordic Seas, weaker negative spikes are also visible for the non-Heinrich stadials (Figure
94 1B).

95

96 **3. Previous Hypotheses for DO cycles**

97 The origin of DO cycles has commonly been explained by changes in the AMOC,
98 but a mechanism for forcing the AMOC at this timescale remains unknown and existing
99 proxy data do not show corresponding changes in the AMOC for every DO cycle. *Winton*

100 [1993] showed that rapid increases in the overturning rate (“flushing” events) could be
101 produced periodically in models by including a constant atmospheric transport of
102 freshwater from low to high latitudes. This mechanism operates on millennial time scales
103 without the need to dictate a periodicity. The magnitude of warming produced by
104 oscillations of the AMOC alone, however, was substantially less than the warming
105 reconstructed over Greenland during DO events [Huber *et al.*, 2006].

106 *Ganopolski and Rahmstorf* [2001] produced a time series of characteristically-
107 shaped DO cycles by forcing an intermediate complexity model with a sinusoidal
108 freshwater flux with a period of 1470 years, which caused large reductions and
109 subsequent resumptions in AMOC strength that resulted in temperature changes over
110 Greenland. We note, however, that there is no known physical mechanism to explain
111 such a sinusoidal fluctuation in the hydrological cycle. Moreover, despite what are likely
112 unrealistically high rates of overturning (~50 Sv) reached by this model, the simulated
113 warming was again considerably less than the reconstructed Greenland temperatures
114 [Huber *et al.*, 2006].

115 Although benthic $\delta^{13}\text{C}$ [Zahn *et al.*, 1997; Shackleton *et al.*, 2000; Elliot *et al.*,
116 2002] and neodymium [Piotrowski *et al.*, 2008; Gutjahr *et al.*, 2010] records from
117 intermediate and deep Atlantic sites indicate substantial changes in the AMOC during
118 DO stadials associated with Heinrich events, no significant changes are seen during non-
119 Heinrich stadials. This indicates that large changes in the AMOC could not have been the
120 primary mechanism behind all the DO cycles.

121 An alternative mechanism for causing abrupt DO warming involves changes in
122 sea-ice cover [Li *et al.*, 2005; Gildor and Tziperman, 2003]. By removing winter sea-ice

123 cover over a large part of the North Atlantic, *Li et al.* [2005] simulated an annual average
124 warming of up to 5-7°C over Greenland, consistent with the lower end of DO warming
125 reconstructed from $\delta^{15}\text{N}$ of gases trapped in the ice [*Huber et al.*, 2006]. In addition, the
126 simulation produced a doubling of accumulation rate and a shift to more wintertime
127 precipitation, also in agreement with observations from ice cores [*Alley et al.*, 1993;
128 *Cuffey and Clow*, 1997; *Svensson et al.*, 2008]. *Li et al.* [2010] also found that a reduction
129 in sea-ice cover in the Nordic Seas alone produced significantly more warming
130 (especially in winter) over Greenland's summit than removing sea-ice cover in the
131 western and central North Atlantic, suggesting that the Nordic Sea region may be critical
132 in terms of influencing the air temperature over Greenland.

133 *Li et al.* [2010] proposed that rapid sea-ice retreat from the Nordic Seas, possibly
134 in response to small changes in wind stress or heat transport, could explain the rapid
135 warming at the onset of a DO cycle. However, this same property of sea ice cannot
136 explain much of the remainder of the DO cycle, which includes the intervals of gradual
137 cooling during the interstadial phase and the sustained cold stadial climate, each of which
138 lasted hundreds of years. This suggests that some other mechanism is needed to set these
139 longer timescales in the DO cycle.

140

141 **4. A Hypothesis for DO Cycles**

142 We propose a conceptual model for DO cycles that explains their characteristic
143 temporal evolution and is supported by existing proxies of ice-sheet, climate and AMOC
144 variability. In particular, we adopt the sea-ice mechanism of *Li et al.* [2005; 2010] to
145 explain the fast-changing intervals of the DO cycles (Figure 2b, 2d). We then invoke an

146 ice shelf to explain the slower-changing phases of the DO cycles (Figure 2a, 2c). From
147 the perspective of the atmosphere, an ice shelf looks the same as sea ice in terms of its
148 albedo and its insulating effects, which reduce the release of heat from the ocean.
149 However, because ice shelves are much thicker than sea ice (100s of m vs. <10 m), they
150 are largely insensitive to small changes in heat transport or wind stress.

151 We first consider the influence of an ice shelf covering a large region of the ocean
152 east of Greenland in the Nordic Seas. Given the sensitivity analysis by *Li et al.* [2010]
153 and the number of proxies showing variability of the cryosphere on DO timescales in the
154 Nordic Seas (e.g. Figure 1B and others) [*Voelker et al.*, 1998; *Rasmussen et al.*, 1996;
155 *Elliot et al.*, 2002; *Dokken and Jansen*, 1999], we focus on an ice shelf along the eastern
156 Greenland margin that could influence sea-ice cover in this region. We propose that the
157 cooling effect of a large ice shelf combined with extensive sea-ice cover would result in
158 regionally cold surface temperatures due to the insulating properties of the ice shelf and
159 sea ice, as well as their effect on local albedo [*Li et al.*, 2005; 2010]. This stadial climate
160 would be maintained for as long as the ice shelf was present.

161 In the event of the ice shelf's collapse, potentially caused by warming of
162 subsurface waters (discussed below), the only remaining ice cover would be sea ice and
163 floating icebergs. A small change in wind stress or heat transport could quickly export or
164 melt this ice, resulting in a large increase in open-ocean area and a corresponding large
165 and abrupt warming over Greenland marking the start of a new DO cycle [*Li et al.*, 2005;
166 2010].

167 During the interstadial phase of a DO cycle, the near doubling of accumulation
168 over the Greenland Ice Sheet that accompanies the warmer climate [*Alley et al.*, 1993;

169 *Cuffey and Clow, 1997; Svensson et al., 2008*] would induce a more positive mass
170 balance, causing the ice shelf to begin reforming along the coast. Expansion of the ice
171 shelf to cover increasingly more ocean surface area would cause air temperatures to
172 gradually cool over Greenland. Once the shelf reached a critical size, it would cause sea
173 ice to rapidly expand through the sea-ice-albedo feedback [*Gildor and Tziperman, 2003*],
174 driving climate back to stadial conditions and completing the DO cycle. **The same cycle**
175 **could not be achieved with multi-year sea ice because its regrowth timescale is**
176 **inconsistent with the gradual decline of climate over the duration of the interstadial**
177 **phase.**

178 In summary, our hypothesis combines the ability of sea ice in the Nordic Seas to
179 explain the rapid transition into and out of the interstadial phase [*Li et al., 2010*] with a
180 gradually expanding ice shelf derived from eastern Greenland to (i) explain the
181 progressive cooling during the interstadial (Figure 2c), (ii) provide the mechanism to
182 trigger sea-ice growth to cause the rapid cooling (Figure 2d), and (iii) sustain the stadial
183 climate once the ice shelf reaches steady state (Figure 2a, 2e). The duration of the
184 interstadial phase is determined by the time required to regrow the ice shelf to a threshold
185 size, beyond which the local ice-albedo effect causes the rapid expansion of sea ice and
186 the corresponding switch to a stadial climate. After a time, ice-shelf collapse, potentially
187 due to subsurface warming, along with an associated rapid loss of sea ice causes the
188 abrupt warming that starts a new DO cycle.

189

190 **5. Discussion**

191 We summarize here proxy records, model results, and modern observations that
192 support key elements of our hypothesis for DO cycles. Multiple lines of evidence support
193 the presence of ice shelves in the northern high latitudes during the last glaciation.
194 Reconstructions of seawater salinity during the LGM show that the ocean was saltier than
195 expected from ice-sheet build-up alone [Adkins *et al.*, 2002]. Reconciling these
196 observations requires either a large change in the volume of groundwater or additional ice
197 shelves equivalent to seven times the volume of the modern Antarctic ice shelves [Adkins
198 *et al.*, 2002]. In addition, there is widespread evidence on the continental shelves
199 surrounding the Nordic Seas, including off eastern Greenland, of fast-flowing ice
200 extending to the shelf edge that may have fed ice shelves [Vorren *et al.*, 1998; Stokes and
201 Clark, 2001; Svendsen *et al.*, 2004; Evans *et al.*, 2009; Dowdeswell *et al.*, 2010].

202 Proxy records suggest substantial variability of the cryosphere in the Nordic Seas
203 on DO timescales. IRD records and planktonic $\delta^{18}\text{O}$ anomalies in the Nordic Seas
204 [Voelker *et al.*, 1998; Dokken and Jansen, 1999] and in the Irminger Basin [van Kreveld
205 *et al.*, 2000; Elliot *et al.*, 1998, 2001] suggest an increase in ice-rafting during each DO
206 stadial (Figure 1B). As discussed previously, these records showing similar-scale
207 variability for every DO stadial differ from those found further south in the IRD belt,
208 where the most prominent IRD and $\delta^{18}\text{O}$ signals are associated with Heinrich events
209 derived from the Laurentide Ice Sheet, and the signals during non-Heinrich DO stadials,
210 particularly in $\delta^{18}\text{O}$, are weak to absent (Figure 1C) [Bond *et al.*, 1992; Cortijo *et al.*,
211 1997; Labeyrie *et al.*, 1999; Hillaire-Marcel and Bilodeau, 2000].

212 An ice shelf constricting the Denmark Strait between Greenland and Iceland may
213 have played an important additional role in influencing sea-ice cover in the Nordic Seas.

214 Firstly, proxies of ice rafting in this area show a strong response on DO timescales
215 (Figure 1B) [Voelker *et al.*, 1998]. Additionally, during the glaciation, grounded ice
216 extended to the shelf break from both Greenland [Vorren *et al.*, 1998; Dowdeswell *et al.*,
217 2010] and Iceland [Hubbard *et al.*, 2006], narrowing the strait to a width of only ~150
218 km [Kosters *et al.*, 2004]. Today, the East Greenland Current passes south through the
219 Denmark Strait and exports substantial sea ice from the Arctic to the North Atlantic. If an
220 ice shelf restricted this outlet, **which is an ideal setting for growing an ice shelf due to**
221 **its shallow shelf bathymetry and proximity to two coastlines**, sea-ice export would
222 likely be impeded. A “log jam” of sea ice could build up north of the Denmark Strait,
223 contributing to further sea-ice expansion through the ice-albedo feedback. The removal of
224 the ice shelf would allow the East Greenland Current to resume, increasing sea-ice export
225 southward into the mid-North Atlantic. In this way, the ice shelf could indirectly
226 influence ice cover over a larger area of ocean.

227 Previously, Hulbe *et al.* [2004] proposed a similar mechanism involving the
228 destruction of an ice shelf in the Labrador Sea to explain Heinrich events, but this
229 hypothesis failed to explain why the ice shelf would collapse only during the cold stadial
230 phases [Alley *et al.*, 2005]. Shaffer *et al.* [2004] explained this relationship by suggesting
231 that warming of intermediate-depth waters associated with a large reduction in the
232 AMOC, such as that which occurred prior to Heinrich events [Zahn *et al.*, 1997; Clark *et*
233 *al.*, 2007; Piotrowski *et al.*, 2008; Pisias *et al.*, 2010; Gutjahr *et al.*, 2010], would cause
234 melting of the Hudson Strait ice shelf from below while surface temperatures remained
235 cold. Additional model results and proxy data provide support for this mechanism

236 [Rasmussen *et al.*, 2003; Clark *et al.*, 2007; Alvarez-Solas *et al.*, 2010, 2011; Marcott *et*
237 *al.*, 2011].

238 Similarly, we propose that subsurface warming caused the collapse of the
239 hypothesized ice shelf along the eastern Greenland margin. In the Nordic Seas,
240 Rasmussen and Thomsen [2004] found changes in benthic fauna that suggest intrusion of
241 warm intermediate waters during stadial phases of DO cycles [Rasmussen *et al.*, 1996;
242 Rasmussen and Thomsen, 2004]. Depleted benthic $\delta^{18}\text{O}$ signals during DO stadials in this
243 region are also consistent with warming of intermediate depth waters [Rasmussen *et al.*,
244 1996; Dokken and Jansen, 1999; Rasmussen and Thomsen, 2004], with a dominant
245 temperature control on these signals supported by Mg/Ca measurements [Jonkers *et al.*,
246 2010; Marcott *et al.*, 2011].

247 Several lines of evidence identify subsurface warming as an effective way to
248 destabilize an ice shelf from below. Modern observations show that warm waters at the
249 base of the ice tongue in front of Jakobshavn Isabrae in western Greenland [Holland *et*
250 *al.*, 2008] and an ice shelf in front of Pine Island glacier in Antarctica [Jenkins *et al.*,
251 2010] increased basal melting, causing thinning, retreat, and destabilization of those ice
252 shelves, leading to accelerated ice discharge. Ice shelf-ice stream models forced by
253 subsurface warming produce similar results [Walker *et al.*, 2009; Joughin *et al.*, 2010].

254 In climate model simulations, warming of intermediate waters in the North
255 Atlantic basin is a robust response to a large reduction in the AMOC [Knutti *et al.*, 2004;
256 Clark *et al.*, 2007; Mignot *et al.*, 2007; Liu *et al.*, 2009; Brady and Otto-Bliesner, 2011].
257 However, model runs show that subsurface warming can still develop with relatively
258 modest changes in the AMOC [Brady and Otto-Bliesner, 2011; Mahajan *et al.*, 2011] and

259 is accompanied by a southward shift in the site of convection [*Brady and Otto-Bliesner*,
260 2011]. In the context of our hypothesis, expansion of the ice shelf as well as increased
261 freshwater fluxes from iceberg calving and melting of sea ice transported southward may
262 have caused a slight reduction in the AMOC and a southward shift in convection, causing
263 subsurface warming to develop locally under the expanded ice shelf fringing Greenland
264 in the Nordic Seas. A decrease in flushing by the AMOC around the ice shelf may have
265 allowed the build-up of atmospherically-derived freshwater in the surface ocean that, in
266 addition to the melting of isotopically depleted icebergs calved off the ice shelf, could
267 have contributed to the light planktonic $\delta^{18}\text{O}$ observed in the region during stadials.

268 **During the LGM, the sea ice edge could have been too far south for the subsurface**
269 **warming to penetrate beneath the ice shelf, resulting in no DO events except**
270 **following Heinrich events when the amount and extent of subsurface warming was**
271 **greater.**

272 Although proxy evidence indicates that large reductions in AMOC strength only
273 occurred during Heinrich stadials [*Zahn et al.*, 1997; *Clark et al.*, 2007; *Piotrowski et al.*,
274 2008; *Pisias et al.*, 2010], existing ocean proxies may not be sensitive to the modest
275 AMOC reductions that models suggest can still induce subsurface warming. Antarctic ice
276 cores show warming events corresponding to the Heinrich stadials [*EPICA Members*,
277 2006], times when the AMOC was significantly reduced and interhemispheric heat
278 transport was weaker. Between these larger Antarctic warming events, smaller events
279 have been correlated with the non-Heinrich stadials [*Wolff et al.*, 2010], consistent with
280 minor changes in heat transport (and therefore AMOC strength) during these times.

281 Proxies outside of the Atlantic hint at global changes in intermediate depth
282 circulation occurring during DO stadials prior to the abrupt warming. High-resolution
283 sediment cores from the Santa Barbara basin show decreases in benthic $\delta^{18}\text{O}$ occurring
284 60-200 years prior to the abrupt decrease in planktonic $\delta^{18}\text{O}$ representing the surface
285 warming of the DO event [*Hendy and Kennett, 2003*]. This phasing was interpreted as a
286 shift in intermediate depth circulation bringing $\delta^{18}\text{O}$ -depleted water from the north
287 Pacific into the basin prior to the large-scale atmospheric reorganizing accompanying the
288 DO event warmed the surface waters [*Hendy and Kennett, 2003*]. In addition, high-
289 resolution ice core studies show that atmospheric N_2O began to rise prior to the rapid DO
290 warmings [*Flückiger et al., 2004*]. In models, global atmospheric N_2O production,
291 predominantly from the tropical Pacific, has been shown to vary as a result of changes in
292 the AMOC [*Schmittner and Galbraith, 2008*], suggesting the early rise in atmospheric
293 N_2O observed in ice cores could be an indicator of changes in Pacific and Atlantic ocean
294 circulations at intermediate depths prior to the main DO event.

295

296 **6. Conclusion**

297 We describe a new mechanism to explain DO cycles involving the formation and
298 collapse of an ice shelf fringing eastern Greenland, potentially extending across the
299 Denmark Strait. Our hypothesis explains the rapid transitions into and out of the
300 interstadial using the ability of sea ice to rapidly expand and contract, whereas the
301 slower-changing phases are explained by the presence or absence of an ice shelf. The
302 duration of the interstadial phase is set by the regrowth timescale of the ice shelf, and the
303 duration of the stadial phase is determined by the timing of ice-shelf removal, potentially

304 due to subsurface warming. Existing proxy evidence from the Nordic Seas supports the
305 idea of fluctuating ice volume in the region in time with DO cycles. Further proxy studies
306 could explore the IRD and meltwater fluxes resulting from such an ice-shelf break up.
307 Modeling work using an active sea-ice model could test the response of sea ice to the
308 presence or absence of an ice shelf fringing eastern Greenland. A combination of these
309 and other approaches can test the feasibility of this idea and illuminate the exact location
310 of the proposed ice shelf.

311

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318

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589

590 **Figure Captions**

591

592 **Figure 1.** Multiple proxies showing DO variability in the Nordic Seas (B) compared to
 593 Heinrich variability in the IRD belt (C). **A.** NGRIP $\delta^{18}\text{O}_{\text{ice}}$ vs. age model GICC05
 594 [Svensson *et al.*, 2008] **B.** Planktonic $\delta^{18}\text{O}$ (black line) and Lithic grain concentration
 595 (#/gram) (grey solid) vs. age model from core PS2644-5 [Voelker *et al.*, 1998] **C.**
 596 Planktonic $\delta^{18}\text{O}$ (black line), >125 μm size fraction (%) (dotted line), and Percent
 597 carbonate (%) (grey solid) vs. age from core MD95-2024 [Hillaire-Marcel and Bilodeau,

598 2000; *Weber et al.*, 2001] **D.** Map showing the location of the proxy records plotted in A-
599 C. Letters on the map correspond to subfigures.

600

601 **Figure 2.** Schematic of proposed DO oscillation mechanism. Phases of the DO cycle
602 labeled a-e with corresponding description of changes in cryosphere and Greenland
603 temperature occurring during each phase. 20-year resolution $\delta^{18}\text{O}_{\text{ice}}$ from NGRIP ice core
604 (grey line) [*Svensson et al.*, 2008] over the period 43-49 ka showing DO 12, with a 10-
605 point smoothing of the data (black line).