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A new mechanism for Dansgaard-Oeschger cycles

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Abstract

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

1. Introduction

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

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

2. Rapid Climate Change in Greenland Ice Cores

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

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

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

Many climate proxies around the globe show DO-like variability on similar time scales. Proxies from the northern hemisphere show warmer (colder) and wetter (drier) climates during DO interstadials (stadials) [*Voelker et al.*, 2002; *Overpeck and* Cole, 2006; *Pisias et al*., 2010]. In the Antarctic EDML ice core, there is an inverse relation between northern and southern hemisphere climate oscillations (bi-polar seesaw), with a correlation between Greenland stadial duration and the amplitude of the Antarctic temperature warming [*EPICA Members*, 2006].

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

3. Previous Hypotheses for DO cycles

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

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

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

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

An alternative mechanism for causing abrupt DO warming involves changes in sea-ice cover [*Li et al.*, 2005; *Gildor and Tziperman*, 2003]. By removing winter sea-ice cover over a large part of the North Atlantic, *Li et al*. [2005] simulated an annual average warming of up to 5-7°C over Greenland, consistent with the lower end of DO warming reconstructed from δ^{15}N of gases trapped in the ice [*Huber et al*., 2006]. In addition, the simulation produced a doubling of accumulation rate and a shift to more wintertime precipitation, also in agreement with observations from ice cores [*Alley et al*., 1993; *Cuffey and Clow*, 1997; *Svensson et al*., 2008]. *Li et al*. [2010] also found that a reduction in sea-ice cover in the Nordic Seas alone produced significantly more warming (especially in winter) over Greenland's summit than removing sea-ice cover in the western and central North Atlantic, suggesting that the Nordic Sea region may be critical in terms of influencing the air temperature over Greenland.

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

4. A Hypothesis for DO Cycles

We propose a conceptual model for DO cycles that explains their characteristic temporal evolution and is supported by existing proxies of ice-sheet, climate and AMOC variability. In particular, we adopt the sea-ice mechanism of *Li et al*. [2005; 2010] to explain the fast-changing intervals of the DO cycles (Figure 2b, 2d). We then invoke an ice shelf to explain the slower-changing phases of the DO cycles (Figure 2a, 2c). From the perspective of the atmosphere, an ice shelf looks the same as sea ice in terms of its albedo and its insulating effects, which reduce the release of heat from the ocean. However, because ice shelves are much thicker than sea ice (100s of m vs. <10 m), they are largely insensitive to small changes in heat transport or wind stress.

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

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

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

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

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

5. Discussion

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

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

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

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

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

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

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

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

In climate model simulations, warming of intermediate waters in the North Atlantic basin is a robust response to a large reduction in the AMOC [*Knutti et al.*, 2004; *Clark et al.*, 2007; *Mignot et al.*, 2007; *Liu et al.*, 2009; *Brady and Otto-Bliesner*, 2011]. However, model runs show that subsurface warming can still develop with relatively modest changes in the AMOC [*Brady and Otto-Bliesner*, 2011; *Mahajan et al.*, 2011] and is accompanied by a southward shift in the site of convection [*Brady and Otto-Bliesner*, 2011]. In the context of our hypothesis, expansion of the ice shelf as well as increased freshwater fluxes from iceberg calving and melting of sea ice transported southward may have caused a slight reduction in the AMOC and a southward shift in convection, causing subsurface warming to develop locally under the expanded ice shelf fringing Greenland in the Nordic Seas. A decrease in flushing by the AMOC around the ice shelf may have allowed the build-up of atmospherically-derived freshwater in the surface ocean that, in addition to the melting of isotopically depleted icebergs calved off the ice shelf, could have contributed to the light planktonic δ^{18}O observed in the region during stadials. **During the LGM, the sea ice edge could have been too far south for the subsurface warming to penetrate beneath the ice shelf, resulting in no DO events except following Heinrich events when the amount and extent of subsurface warming was greater.**

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

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

6. Conclusion

We describe a new mechanism to explain DO cycles involving the formation and collapse of an ice shelf fringing eastern Greenland, potentially extending across the Denmark Strait. Our hypothesis explains the rapid transitions into and out of the interstadial using the ability of sea ice to rapidly expand and contract, whereas the slower-changing phases are explained by the presence or absence of an ice shelf. The duration of the interstadial phase is set by the regrowth timescale of the ice shelf, and the duration of the stadial phase is determined by the timing of ice-shelf removal, potentially due to subsurface warming. Existing proxy evidence from the Nordic Seas supports the idea of fluctuating ice volume in the region in time with DO cycles. Further proxy studies could explore the IRD and meltwater fluxes resulting from such an ice-shelf break up. Modeling work using an active sea-ice model could test the response of sea ice to the presence or absence of an ice shelf fringing eastern Greenland. A combination of these and other approaches can test the feasibility of this idea and illuminate the exact location of the proposed ice shelf.

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Figure Captions

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

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

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