Abrupt climate shifts in Greenland due to displacements of the sea ice edge

Camille Li[1*], David S. Battisti[1], Daniel P. Schrag[2], Eli Tziperman[2,3]

[1]Department of Atmospheric Sciences, University of Washington
408 ATG, Box 351640, Seattle WA 98195 USA
[2]Department of Earth and Planetary Sciences, Harvard University
20 Oxford Street, Cambridge MA 02138 USA
[3] Division of Engineering and Applied Sciences, Harvard University
20 Oxford Street, Cambridge MA 02138 USA

[*]To whom correspondence should be addressed; E-mail: camille@atmos.washington.edu.

An atmospheric circulation model is used to show that small reductions in sea ice extent in the North Atlantic are capable of explaining the abrupt changes in temperature, snow accumulation and oxygen isotopes recorded in Greenland during the Dansgaard-Oeschger (D-O) events of the last glacial period. Model simulations indicate that reduced sea ice cover in this region produces warming that is especially pronounced in winter and an accumulation increase that occurs primarily in summer. Mechanisms for driving such displacements of sea ice could be small changes in ocean thermohaline circulation (OTC) or rearrangements of the tropical atmosphere-ocean system.

The Dansgaard-Oeschger (D-O) events that punctuated the last glacial period (50–10 kyr BP) are abrupt warming episodes recorded in Greenland ice cores (*1*). Each event is characterized by a large temperature rise (7–10°C) over several decades, with the warm conditions lasting for 200–600 years before a more gradual return to the glacial state (*2*). Recent studies have found that the warming in Greenland is coincident with abrupt changes in other parts of the Northern Hemisphere as well as the tropics (*3–7*). A leading hypothesis attributes D-O events to an internal oscillation of the ocean thermohaline circulation (OTC) (*8*). By switching the OTC from its sluggish glacial mode to one which features an increase in ocean heat transport (OHT) into the North Atlantic, this oscillation could cause an abrupt warming in Greenland (*2, 9*). However, state-of-the-art climate models and theory suggest that even large changes in mid- and high latitude sea surface temperature (SST) produce weak temperature responses in Greenland, whether they occur during modern or glacial climates (*10–12*). If OTC changes are indeed responsible for D-O events, a critical question is how they can produce the large, abrupt changes observed in the Greenland ice cores and elsewhere.

Sea ice is a key component of the climate system and can affect it through a variety of feedbacks. The presence of sea ice increases the planetary albedo, causing more sunlight to be reflected from the Earth's surface and lowering global mean temperature. Sea ice can also influence regional air temperature in winter by insulating the atmosphere from the substantial heat capacity of the ocean. When sea ice is absent, the ocean can absorb heat in the summer and release it back to the atmosphere in the winter, thereby moderating the extreme cold of the polar night in high latitudes. The existence of such feedbacks points to the possibility of rapid changes in sea ice cover resulting from relatively weak forcing (*13*). Several studies have identified displacements of the sea ice edge as an alternative mechanism for D-O events (*14–16*) with the possibility of rapid, switch-like changes in sea ice cover caused by small OTC variability (*17, 18*).

We demonstrate through atmospheric general circulation model (GCM) experiments that removal of a relatively small amount of sea ice in the North Atlantic can explain the D-O warming signal observed in Greenland. Furthermore, these sea ice changes are consistent with the snow accumulation and oxygen isotope records from the Greenland ice cores. The associated change in meridional OHT implied by the model is small.

We used the NCAR Community Climate Model version 3 (CCM3) atmospheric GCM to run simulations of a scenario with reduced North Atlantic sea ice cover and warmer North Atlantic sea surface temperature relative to a control Last Glacial Maximum (LGM) climate (see Fig. 1 for prescribed boundary conditions) configured with Peltier ice sheets (*19*), revised CLIMAP SST (*20*) and sea ice (*21*), orbital parameters for 21kyr BP, 200 ppm CO_2 and 350 ppb CH_4. This control LGM simulation generates reasonable values of annual mean temperature, annual accumulation and maximum/minimum temperatures at the Greenland Summit (*22*).

The scenario with reduced sea ice around Greenland (scenario I, see Table 1 for details) shows warming in the North Atlantic region with magnitudes reaching 7°C (*23*) at the Greenland Summit (panel (b) of Fig. 1). This Greenland warming is comparable to the 7–10°C temperature rise during D-O events determined through measurement of gas fractionation in air bubbles trapped within Greenland ice (*24*). The air temperature response in scenario I is localized, but the prescribed SST model limits teleconnections, and in particular, lacks potential tropical linkages that have been proposed to exist in glacial climates (*25*). These linkages, in conjunction with potentially enhanced glacial teleconnections (*26*), could lead to feedbacks that would both reinforce the warm temperatures in Greenland and produce far-field responses throughout the Northern Hemisphere. The climate footprint from a simulation incorporating these feedbacks could be compared with records such as those from the Cariaco Basin (*5*), Hulu Cave (*6*), Bermuda Rise (*3*) and Santa Barbara Basin (*4*).

Looking beyond the annual mean picture, results from scenario I and an additional simulation with extreme sea ice retreat to 80N around Greenland in summer alone (scenario II) point to winter as important for generating Greenland warmth (see Table 1). Less sea ice in winter reduces the areal extent of insulation over the relatively warm ocean, which is a significant source of heat to the atmosphere during the polar night. In scenario I, winter accounts for 2°C of warming in the 7°C annual mean signal. Summer changes may also play a role as less sea ice provides a darker surface for absorbing incoming solar radiation. However, the summer changes in scenario I contribute only 1°C to the annual mean signal, and even a large reduction in sea ice extent during summer months such as in scenario II produces only a 2°C warming in the annual mean.

We can delve further into the issue of seasonality by considering constraints provided by the snow accumulation and oxygen isotope records in ice cores. These records indicate that warming during D-O events was accompanied by a 50–100% increase in accumulation (*27,28*) and a 3–4 per mil increase in δ^{18}O (*29,30*). Table 1 and Figure 2 show temperature, accumulation and accumulation-weighted temperature from our model simulations broken down as monthly or seasonal contributions to the annual signal. Scenario I shows a doubling of total annual accumulation, which is within the limits of the observational range. Although the distribution of accumulation shifts slightly towards winter, most of the actual increase occurs in summer. The sea ice changes in scenario II, which occur only in summer, also produce a doubling of accumulation, but with little effect on Greenland temperature. Additional experiments (*31*) indicate that further increases in summer SST produce snowfall amounts that are much larger than those observed. Thus, if sea ice is involved, D-O events must include neither extreme sea ice retreat nor excessive ($>$2.5°C) North Atlantic warming in summer. These results support a recent suggestion that D-O events are primarily a wintertime phenomenon (*32*).

The δ^{18}O record has traditionally been used as a measure of air temperature (*33*), but more recent work has shown that other factors influence the isotope signal in Greenland (*24,34*). The complications are related in large part to the fact that the isotope signal is recorded by, and hence dependent on, accumulation. Models generally agree that during glacial winters, extensive sea ice cover inhibits evaporation over the North Atlantic and very zonal atmospheric circulation prevents cyclones from advecting moisture over Greenland (*10, 35, 36*). As a consequence, Greenland sees very little snowfall in winter compared to summer, a result that is supported by evidence from close-off porosity measurements (*37*). This pronounced seasonality in accumulation rates during glacial times compared to today introduces a warm (summer) bias to the ice core record (*10, 35, 36, 38*). Indeed, Werner *et al.* (*36*) showed that snow seasonality is the dominant control for the glacial/interglacial transition (*39*), although others have argued for the importance of other factors such as moisture origin (*40*) and tropical cooling (*41*) in determining the isotopic composition of Greenland snowfall.

If this is also true for D-O events, then we can calculate an accumulation-weighted temperature change associated with D-O warming that should correspond directly to the muted (5–7°C) isotopic paleothermometer estimate rather than the more pronounced ($\sim$10°C) estimate derived from N fractionation (*24*). From Table 1, we see that scenario I does indeed produce a damped "isotope" signal (ΔwT = 2.5°C) compared to the actual annual mean temperature signal from the model (ΔT = 7°C), but that it is in fact too *cold* or *depleted*. An examination of the DJF columns in Table 1 reveals that even a small increase in wintertime snowfall shifts the seasonality of accumulation sufficiently that the cold Greenland winter contributes more to the accumulation-weighted temperature. Thus, despite its tendency to weaken the D-O isotope signal, a change in the seasonality of precipitation alone may not be able to explain the δ^{18}O record entirely. Ongoing work aims to ascertain if other effects such as changes in the source regions or transport (i.e., the amount of processing between origin and Summit) of Greenland-bound water vapor, both of which conceivably enrich the isotope signal during times of reduced sea ice (*36, 40*), are involved.

This study has shown that relatively small changes in winter sea ice extent in the North Atlantic are capable of producing the observed warming in Greenland during D-O events. Although rapid displacements of the sea ice edge are responsible for the abrupt warming signal, they must be driven by other parts of the climate system. The driver could be local (e.g. OTC changes in the North Atlantic (*2,18,42*)) or reside in more distal regions such as the tropics (*26*). If the ocean is the primary driver, a successful hypothesis must explain a strengthening of the OTC in the North Atlantic that initiates abruptly and persists for several hundred years before gradually returning to a weaker (glacial) mode. Modelling studies have identified such oscillations in the OTC (*9*) supported by the climatological mean state of the glacial ocean (*2*). In addition to the complications for the OTC hypothesis discussed earlier is the apparent phase locking of D-O events on a 1470 year cycle for over 50 kyr (*43, 44*). Furthermore, model experiments indicate that changes in OTC lead to short-lived (on the order of decades) changes in sea ice extent (*12*) which induce a strong, restoring atmospheric response in the modern climate (*45*).

Although our model does not include an interactive ocean, we can get a sense of the OTC changes consistent with the reduced sea ice scenario through the surface heat budget. By making the assumption that any imbalance in the budget must be accounted for by ocean circulation, and then integrating from the North Pole to a given latitude, we estimate the amount of heat transported by the ocean to sustain the specified sea ice and SST distributions (Fig. 3). This calculation shows a very small change in implied OHT between the LGM and reduced sea ice (I) scenarios, with a difference equivalent to less than several Sverdrups of overturning using modern day scaling (*46*). This result appears in contrast to those from studies in which warming over the North Atlantic region is attributed to very large changes in ocean circulation (*2*) rather than changes in sea ice (*18*). The apparent inconsistency is not surprising, however, as changes in sea ice extent have proven more effective than changes in midlatitude SST for generating large amplitude, far-field climate responses (*11, 47*).

Because sea ice is sensitive to forcing from the atmosphere as well as the ocean (*48*), an alternative driver for sea ice retreat is a change in the surface wind stress in the North Atlantic (*49, 50*). These surface wind anomalies themselves could have non-local origins such as the interaction of atmospheric circulation with the land-based ice sheets or changes in the tropical atmosphere-ocean system (*26*). Further efforts to understand the factors and feedbacks that control the advance and retreat of the sea ice edge are required in order to resolve this issue. Nonetheless, the results presented here support abrupt displacements of the sea ice edge as a leading candidate for explaining the D-O signal recorded in Greenland.

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23. The temperature and accumulation changes quoted for the reduced sea ice scenario corresponds to at least several standard deviations of its internal variability. This estimate provides a conservative measure of the statistical significance of the results since the internal variability of this scenario is larger than that of either the modern or glacial control climates of the model.

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Table 1: Temperature at 2 m reference height, accumulation and accumulation-weighted temperature (weighted T) for the four model scenarios: modern, LGM, I (reduced sea ice) and II (reduced sea ice in summer only, with the ice line retreating to 80N around Greenland). For temperature and accumulation, winter (DJF) and summer (JJA) breakdowns are included in addition to the annual averages. The accumulation values in parentheses are the fraction of the total annual accumulation contributed by the given season. The columns marked ΔT, Δacc and ΔwT show the annual mean difference relative to the LGM simulation. These results are an average over the four grid points near the Greenland Summit (*23*).

exp	**2m temperature (C)**				**accumulation (cm/y)**				**weighted T (C)**	
	DJF	JJA	ANN	ΔT	DJF (%)	JJA (%)	ANN	Δacc (%)	ANN	ΔwT
modern	-33	-8	-22		20.5 (22)	34.2 (36)	23.6		-18.6	
LGM	-63	-24	-45		0.6 (4)	11.0 (70)	4.0		-28.3	
I	-55	-20	-38	7	1.2 (4)	20.1 (62)	8.0	+100	-25.5	2.5
II	-63	-19	-43	2	0.6 (2)	25.5 (79)	8.1	+100	-22.4	5.8

Figure 1: Comparison of LGM control experiment and the reduced sea ice scenario (I). **(A)** Sea surface temperature boundary conditions (degrees Celsius) for the LGM scenario (left) and reduced sea ice scenario (right). Maximum (February) and minimum (August) sea ice extents are indicated with the solid and dotted red lines, respectively. Scenario I has a maximum sea ice extent equivalent to the LGM perennial ice cover, and a minimum sea ice extent equivalent to the modern day minimum ice cover. The ice thickness is 2 metres, which is a typical value for the Arctic today. **(B)** The difference in surface air temperature between the two experiments (degrees Celsius).

Figure 2: Seasonal cycle of temperature (top), monthly accumulation (middle) and monthly accumulation as a percentage of total annual accumulation (bottom) at the Greenland Summit.

Figure 3: Implied ocean heat transport (OHT) in the Atlantic Ocean for the modern, LGM control and reduced sea ice (scenario I) simulations calculated by integrating the surface heat imbalance from the North Pole southwards.