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

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[1] An atmospheric circulation model is used to show that a reduction in sea ice extent in the North Atlantic produces a climatic response consistent with abrupt changes in temperature and snow accumulation recorded in Greenland during the Dansgaard-Oeschger (D-O) events of the last glacial period. The model simulations exhibit warming that is especially pronounced in winter and an accumulation increase that occurs primarily in summer, though the fractional accumulation increase is greater in winter. The seasonality of the combined temperature and accumulation changes is also consistent with the oxygen isotope signal, explaining why estimates of the magnitude of D-O warming from oxygen isotopes are damped relative to direct estimates. Mechanisms for driving such displacements of sea ice could be changes in ocean thermohaline circulation (OTC) or rearrangements of the tropical atmosphere-ocean system. **Citation:** Li, C., D. S. Battisti, D. P. Schrag, and E. Tziperman (2005), Abrupt climate shifts in Greenland due to displacements of the sea ice edge, *Geophys. Res. Lett.*, *32*, L19702, doi:10.1029/2005GL023492.

1. Introduction

[2] 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 [*Dansgaard et al.*, 1993]. A typical event is characterized by a large annual mean temperature rise (typically 5–10°C [*Ganopolski and Rahmstorf*, 2001; *Severinghaus et al.*, 2003]) over several decades, with the warm conditions lasting for 200–600 years before a more gradual cooling, sometimes followed by an abrupt return to the glacial state. Recent studies have found that the warming in Greenland is coincident with abrupt changes in other parts of the Northern Hemisphere and global tropics [*Rahmstorf*, 2002, and references therein]. A leading hypothesis attributes D-O events to an internal oscillation of the ocean thermohaline circulation (OTC) [*Broecker et al.*, 1990]. 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 [*Winton*, 1993; *Ganopolski and Rahmstorf*, 2001]. The climate models used to illustrate this theory all contain simplified atmospheric dynamics that do not simulate temperature, precipitation and wind anomalies accurately away from regions where the SST and sea ice changes occur. Theory and more sophisticated climate models indicate that large changes in mid- and high latitude sea surface temperature (SST) alone resulting from shifts in the OTC produce a temperature response near Greenland that is weak compared to the D-O jumps, regardless of whether they occur under modern or glacial boundary conditions [*Fawcett et al.*, 1997; *Seager et al.*, 2002; *Vellinga et al.*, 2002]. If OTC changes are indeed responsible for D-O events, a critical question is how such a modest climate forcing can be reconciled with the large signals recorded in Greenland ice cores.

[3] Sea ice is a key component of the climate system and can affect it through a variety of feedbacks. In addition to the well-known ice-albedo feedback, sea ice has an important influence on regional air temperature in winter by insulating the atmosphere from the substantial heat capacity of the ocean. When sea ice is absent, the ocean absorbs heat in summer and releases it back to the atmosphere in 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 [*Maykut and Untersteiner*, 1971]. Several studies have proposed displacements of the sea ice edge as a mechanism for D-O events [*Dansgaard et al.*, 1989; *Alley et al.*, 1993; *Broecker*, 2000; *Denton et al.*, 2005] with the suggestion of rapid, switch-like changes in sea ice cover caused by either subtle shifts in wind stress near the ice edge, or by small OTC variability [*Gildor and Tziperman*, 2003; *Kaspi et al.*, 2004].

[4] We demonstrate through atmospheric general circulation model (AGCM) experiments that removal of sea ice in the North Atlantic can explain the magnitude of 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. While the sea ice studies described above use dynamical oceans to capture the timescale and timing pertinent to the OTC mechanism for abrupt climate events, the approach in our sensitivity study is to impose conditions at the sea surface and ask how large an effect they will have in Greenland. Although our model lacks ocean dynamics and atmosphere-ocean interactions, its better representation of atmospheric dynamics allows for good estimates of what Greenland will experience given these displacements in North Atlantic sea ice.

2. Experimental Design

[5] We used the NCAR Community Climate Model (CCM3) AGCM to simulate scenarios with reduced North Atlantic sea ice cover and warmer North Atlantic SST relative

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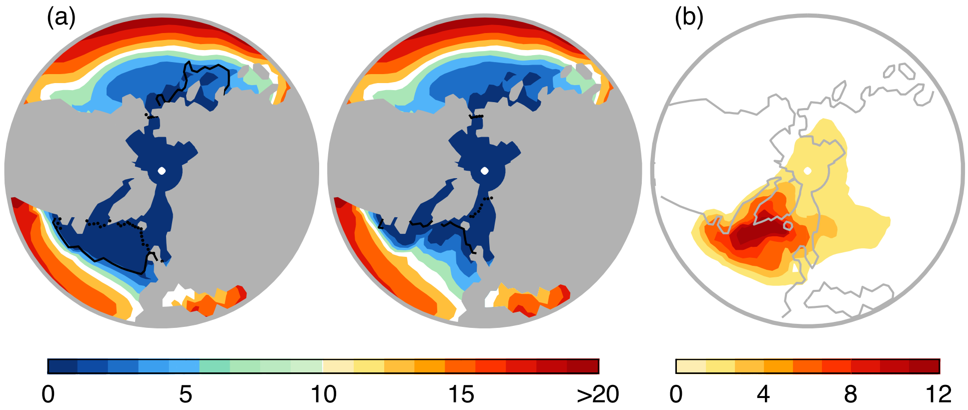

Figure 1. Comparison of LGM and reduced sea ice scenario I. (A) Annual mean sea surface temperature boundary conditions (degrees Celsius) for LGM (left) and reduced sea ice scenario (right). Maximum (February) and minimum (August) sea ice extents are indicated with the solid and dotted lines, respectively. 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 perennial 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 simulations (degrees Celsius).

to a Last Glacial Maximum (LGM) climate configured with Peltier ice sheets [*Peltier*, 1994], revised CLIMAP SST [*Crowley*, 2000] and sea ice [*Sarnthein et al.*, 2003], 21kyr BP insolation, 200 ppm CO_2 and 350 ppb CH_4 (Figure 1) Spectral smoothing lowers the LGM Greenland Summit by 900 geopotential meters. Extrapolating to the true summit elevation using the relatively linear "land surface lapse rate" of Greenland for each month of the year yields a more concordant (colder) LGM Summit temperature of $\sim$−50°C. Assuming that snowfall at the Summit scales with saturation vapor pressure, the associated error in accumulation is less than 10%.

[6] In a suite of reduced sea ice experiments, the atmospheric response in the vicinity of Greenland was relatively insensitive to the details of the prescribed sea ice changes. We will mainly discuss results from a simulation with maximum extent corresponding to the LGM perennial ice line, minimum extent corresponding to the modern perennial ice line, and ice distributions interpolated between these extremes for the intervening months (scenario I). This reduction, while somewhat arbitrary, is comparable to the sea ice variability associated with small (2 Sv) changes in Atlantic overturning in an LGM climate simulated by NCAR's fully coupled Community Climate System Model CCSM3 (B. Otto-Bliesner et al., unpublished data, 2005).

3. Results

[7] Reduced sea ice scenario I (see Table 1) shows warming in the North Atlantic region with an annual temperature change of 7°C around the Greenland Summit (Figure 1b). This warming is comparable to the 5–10°C temperature rise during D-O events determined from gas fractionation in air bubbles trapped within Greenland ice [*Severinghaus and Brook*, 1999; *Severinghaus et al.*, 2003]. The temperature response in scenario I is localized as the prescribed SST model limits teleconnections and lacks tropical atmosphere-ocean feedbacks demonstrated to be important in glacial climates [*Chiang et al.*, 2003]. These linkages, in conjunction with potentially enhanced glacial teleconnections [*Yin and Battisti*, 2001], could both reinforce the warm temperatures in Greenland and produce far-field responses throughout the Northern Hemisphere.

[8] Looking beyond the annual mean picture, results from scenario I and an additional simulation with sea ice retreat to 80N around Greenland in summer alone (scenario II) point to winter as important for generating Greenland warmth (see

Table 1. Temperature at 2 m Reference Height, Accumulation and Accumulation-Weighted Temperature (Weighted T) for the Four Simulations: Modern, LGM, I (Reduced Sea Ice) and II (Reduced Sea Ice in Summer Only, With the Ice Line Retreating to 80N Around Greenland)[a]

	2 m Temperature, C				Accumulation, cm/y				Weighted T, C	
Experiment	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

[a]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. These results are an average over 70N–75N and 34W–48W near the Greenland Summit. Temperature and accumulation changes quoted for the reduced sea ice scenario correspond to at least several standard deviations of its internal variability and are significant at the 95% confidence level.

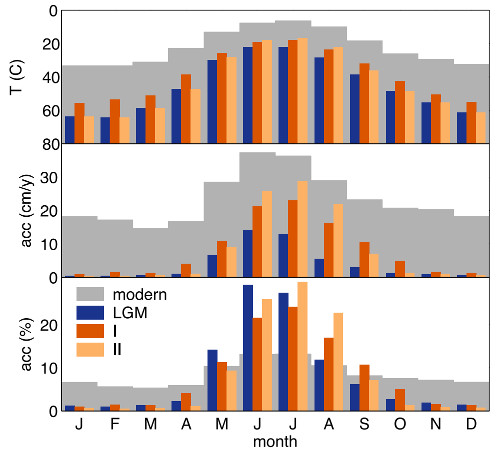

Figure 2. Seasonal cycle of temperature (top), monthly accumulation (middle) and monthly accumulation as a percentage of total annual accumulation (bottom) at the Greenland Summit. The Greenland Summit is taken to be 70N–75N and 34W–48W; all quantities shown are an area-weighted average over this box.

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 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 signal; even a large reduction in sea ice during summer months such as in scenario II produces only a 2°C warming in the annual mean.

[9] We can delve further into the issue of seasonality by considering constraints provided by snow accumulation and $\delta^{18}0$ in ice cores. These records indicate that warming during D-O events was accompanied by a 50–100% increase in accumulation [*Dahl-Jensen et al.*, 1993; *Cuffey and Clow*, 1997] and a 3–4 per mil increase in $\delta^{18}0$ [*Grootes and Stuiver*, 1997]. 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 (C. Li et al., unpublished data) indicate that further increases in summer SST produce much more snowfall than 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 observation-based evidence that D-O events are primarily a winter phenomenon [*Denton et al.*, 2005, and references therein].

[10] The $\delta^{18}0$ record has traditionally been used as a measure of air temperature [*Dansgaard*, 1964], but other factors also influence the isotope signal in Greenland [*Cuffey et al.*, 1994; *Severinghaus and Brook*, 1999]. The complications are related in large part to the fact that the isotope signal is recorded by, and hence dependent on, accumulation. Since Greenland sees very little snowfall in winter compared to summer [*Fawcett et al.*, 1997; *Krinner and Genthon*, 1997; *Raynaud et al.*, 1997; *Werner et al.*, 2000], there is a pronounced seasonality in accumulation rates during glacial times that introduces a warm (summer) bias to the ice core record [*Steig et al.*, 1994]. Although some have argued for the importance of other factors in determining the isotopic composition of Greenland snowfall [*Charles et al.*, 1994; *Boyle*, 1997], *Werner et al.* [2000] showed that snow seasonality is the dominant control for the glacial/interglacial transition. Indeed, the differences in mean temperature (ΔT = 22°C) and accumulation-weighted temperature (ΔT = 9°C) between our LGM and control experiments reproduce the disparity in the glacial/interglacial temperature transition from borehole and $\delta^{18}O$ measurements.

[11] 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 to the muted (5–7°C) isotopic paleothermometer estimate rather than the more pronounced gas fractionation (5–10°C) estimates [*Severinghaus et al.*, 2003]. 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 (ΔT = 7°C) from the model, but that it is in fact too depleted (cold). Even a small increase in winter snowfall shifts the seasonality of accumulation sufficiently that the cold Greenland winter contributes more to the accumulation-weighted temperature (Table 1). Thus, despite its tendency to weaken the D-O isotope signal, a change in the seasonality of precipitation alone may not explain entirely the $\delta^{18}0$ record. Ongoing work aims to ascertain if effects such as changes in source regions or transport of Greenland-bound water vapor, both of which conceivably enrich the isotope signal when there is less sea ice [*Charles et al.*, 1994; *Werner et al.*, 2000], are involved.

4. Discussion

[12] This study has shown that changes in winter sea ice extent in the North Atlantic are consistent with observed signals in temperature, accumulation and $\delta^{18}O$ in Greenland during D-O events. Although rapid displacements of the sea ice edge may be responsible for the abrupt warming signal, they must be driven by other parts of the climate system. The driver could be local (OTC changes in the North Atlantic [*Ganopolski and Rahmstorf*, 2001; *Kaspi et al.*, 2004; *Knutti et al.*, 2004]) or reside in more distal regions such as the tropics [*Yin and Battisti*, 2001].

[13] If ocean-driven, a successful hypothesis must explain a strengthening of North Atlantic OTC that initiates abruptly and persists for several centuries before returning to a weaker mode. Modelling studies have identified such OTC oscillations [*Winton*, 1993; *Ganopolski and Rahmstorf*, 2001], but the idealized boundary conditions (bathymetry, basin geometry) and coarse grids used are known to be problematic for accurate simulation of ocean circulation. Furthermore, model experiments indicate that in the modern climate, changes in OTC lead to short-lived (on the order of decades) changes in sea ice extent [*Vellinga et*

al., 2002] which induce a restoring atmospheric response [*Deser et al.*, 2004]. This ocean-driven hypothesis instead requires positive feedbacks to sea ice changes from the overlying atmosphere in the LGM to allow for the observed D-O warming timescale. Although our model does not include an interactive ocean, we can obtain indirect evidence of ocean changes by assuming that any imbalance in the surface heat budget is accounted for by ocean circulation. Integrating this imbalance from the North Pole to 40N, we estimate a 15% decrease in ocean heat transport (OHT) needed at this latitude to sustain the reduced sea ice distribution.

[14] Because sea ice is sensitive to forcing from the atmosphere as well as the ocean [*Bitz et al.*, 2005], an alternative driver for sea ice retreat is a change in surface wind stress in the North Atlantic [*Fang and Wallace*, 1994]. The surface wind anomalies could themselves have non-local origins such as the interaction of atmospheric circulation with land-based ice sheets or changes in the tropical atmosphere-ocean system [*Yin and Battisti*, 2001]. Further efforts to understand the factors and feedbacks that control the advance and retreat of sea ice are required in order to resolve this issue. Nonetheless, the results presented here support the idea that displacements of the sea ice edge during D-O events are recorded in Greenland ice cores, and are key to interpreting records of abrupt climate events in the North Atlantic.

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