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The Younger Dryas termination and North Atlantic Deep Water formation: Insights from climate model simulations and Greenland ice cores

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Abstract. Results from the GISP2 and GRIP ice cores show that the termination of the Younger Dryas (YD) climate event in Greenland was a large and extremely fast climate change. A reinitiation of North Atlantic Deep Water formation following a shutdown, and its associated winter release of heat to the atmosphere, has been suggested as the most likely cause of this climate transition. To test this idea, two general circulation model experiments using GENESIS have been completed for YD time (12,000 calendar years ago): one with low heat flux in the Nordic Seas (10 W/m^2, deep water shutdown) and one with high Nordic Sea heat flux (300 W/m^2, active deep water formation). Comparison of Greenland climate differences between these experiments with the ice core records shows that when deep water is turned on, much of the YD termination warming is achieved. The increase in precipitation is underestimated because of a model tendency to overestimate summertime precipitation, which obscures the dominantly wintertime response to the specified forcing. The winter storm track shift toward Greenland contributes much of the climate change at the YD termination.

Introduction

The Younger Dryas (YD) climate event is the best-known in a series of abrupt returns to almost full glacial conditions during the last deglaciation. New Greenland ice core data show that the termination of this event was a large and extremely fast climate change, occurring in decades or less, and possibly in as little as 1 to 3 years [*Dansgaard et al.*, 1989; *Alley et al.*, 1993; *Mayewski et al.*, 1993; *Taylor et al.*, 1993]. Both the magnitude and rapidity of the Younger Dryas - Preboreal (PB) climate transition have broad implications for our understanding of variability in the climate system, especially in view of current concern over anthropogenic climate change [e.g., *Broecker*, 1995]. Much attention has therefore been focussed on identifying possible climate forcing mechanisms that operate on timescales much shorter than orbital variations.

Rapid climate oscillations such as the Younger Dryas have been attributed to sudden changes in the North Atlantic thermohaline circulation and associated poleward advection of warm surface waters [e.g., *Broecker and Denton*, 1989; *Broecker et al.*, 1985, 1990]. Previous climate model studies [e.g., *Bryan*, 1986; *Manabe and Stouffer*, 1988; *Birchfield and Broecker*, 1990; *Wright and Stocker*, 1993] have shown that the thermohaline overturn in the North Atlantic can oscillate between two modes, "on" and "off," as a function of surface salinities. *Broecker and Denton* [1989] attributed the onset of Younger Dryas cooling to a shutdown of North Atlantic Deep Water (NADW) formation when Laurentide ice sheet meltwater was diverted from the Mississippi to the St. Lawrence drainage and caused

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a rapid freshening of North Atlantic surface waters. In this model, the Younger Dryas ended when the Atlantic salt conveyor reestablished itself and warm surface waters were again advected to high latitudes in the North Atlantic. In addition to the drainage diversion there is substantial evidence for a large iceberg discharge from the Hudson Strait into the North Atlantic at this time (Heinrich event, H-0) and these in combination could have generated the low salinity cap necessary to shut down deep convection [*Miller and Kaufman*, 1990; *Bond et al.*, 1993; *Andrews et al.*, 1994].

In this study we use an atmospheric general circulation model (AGCM) which includes an explicit parameterization for ocean heat transport to test the hypothesis that the Younger Dryas to Preboreal (PB) climate transition resulted from a restart of NADW formation. Two experiments for Younger Dryas-age boundary conditions were carried out: one with low ocean heat convergence in the Nordic Seas (Norwegian and Greenland Seas) representing the NADW shutdown phase, and one with high ocean heat convergence in the Nordic Seas representing NADW formation and advection of warm surface waters into this region. The difference in climate results between the two experiments is compared with the high-resolution climate record from the GISP2 ice core.

The Younger Dryas Termination in Greenland

The Younger Dryas to Preboreal climate change at Summit, Greenland, is well documented in the various proxy climate records obtained from the new GRIP and GISP2 ice cores. The upper two thirds of the GISP2 core was dated by counting annual layers [*Alley et al.*, 1993; *Meese et al.*, 1994] which allows an assignment of calendar-year ages to the climate records, and the high resolution of the ice core allows for detailed inspection of the climate transitions. The Younger Dryas began ~12,700 ±200 (calendar) years ago and ended ~11,500 ±200 years ago [*Alley et al.*, 1993].

Electrical conductivity measurements of the ice core show an abrupt decrease (10 years or less) in alkaline dust loading at the Younger Dryas termination [*Taylor et al.*, 1993], and chemical analyses of the ice also show a sharp decrease in atmospheric loading of crustal material and sea salt, as well as large fluctuations in ammonium [*Mayewski et al.*, 1993; *Alley et al.*, 1995]. The size and rapidity of these changes probably indicate a shift in atmospheric circulation patterns rather than changes in the dust source areas.

The $\delta^{18}O$ (per mil relative to SMOW) of ice shows a 3.5 to 4 per mil increase over 50 years at the Younger Dryas termination [*Dansgaard et al.*, 1989; *Johnsen et al.*, 1992; *Grootes et al.*, 1993]. The isotopic thermometer has been calibrated against borehole temperatures for the warmings from the Little Ice Age (0.55 per mil per °C [*Cuffey et al.*, 1994) and from the glacial maximum (0.33 per mil per °C [*Cuffey et al.*, 1995]). No direcct calibration was obtained for the Younger Dryas because its borehole-temperature signal has largely diffused away, although the isotopic data do contain temperature information [*Cuffey et al.*, 1994, 1995]. These calibrations yield warmings of 7°-8°C to 10°-12°C at the Younger Dryas termination. Changes in moisture-source temperature [*Dansgaard et al.*, 1989] and perhaps location [*Charles et al.*, 1994] may have enhanced the isotopic change, causing an overestimate of the temperature change. Analysis of the isotopic composition of trapped gases apparently has detected the thermal diffusion signal of a large, abrupt warming at the end of the Younger Dryas [*Severinghaus et al.*, 1996]. Preliminary calculations based on the inferred gas age-ice age difference at the Younger Dryas termination, the reconstructed accumulation rates, and an empirical firn densification model that yields the gas age-ice age difference as a function of temperature and accumulation are consistent with a moderate to large warming, perhaps 7° to 8°C [*Severinghaus et al.*, 1996].

At the Younger Dryas termination, ice accumulation rates doubled in a matter of decades, and much of this change occurred in as little as 1 to 3 years [*Alley et al.*, 1993]. The change in accumulation rates at this transition (10%/°C for a 7°C warming, and no less than 6%/°C assuming an upper limit warming of 12°C) is much higher than would be expected from a purely thermodynamic effect (the ability of warmer air to carry more moisture, 4%/°C [*Johnsen et al.*, 1989]). This requires dynamic changes affecting accumulation with the warming and most likely a storm track shift toward Greenland [*Kapsner et al.*, 1995]. Such shifts in storm tracks are also argued for by *Mayewski et al.*, [1993] based on rapid changes in the loading of dust and other particulates in the ice.

Climatic oscillations recorded in North Atlantic marine sediment cores support the hypothesis of a link between the ocean's thermohaline circulation and the abrupt climate changes of the last deglaciation. Fine-scale color changes related to abundances of *Neogloboquadrina pachyderma* (*s*) in Deep Sea Drilling Project core 609 in the central Atlantic resemble the $\delta^{18}O$ record in the Camp Century core (south Greenland), which suggests a common climatic origin [*Broecker et al.*, 1990]. The high-resolution Troll 3.1 core from the Norwegian trench records sea surface warming of 5°C or more in fewer than 40 years at the end of the Younger Dryas due to increased inflow of warm Atlantic surface water to the Norwegian Sea [*Lehman and Keigwin*, 1992].

Model and Boundary Conditions

The GENESIS (version 1.02) general circulation model [*Thompson and Pollard*, 1995] used in this study is an extensively modified version of the National Center for Atmospheric Research (NCAR) Community Cli-

mate Model CCM1 [*Williamson et al.*, 1987]. The atmospheric model is coupled to a 50-m slab mixed layer ocean in which poleward ocean heat transport is included as a zonally symmetric function of latitude based on present-day observations using 0.3 times the "0.5 X OCNFLX" case of *Covey and Thompson* [1989]. This reduction gives a best fit to present zonal mean sea surface temperatures [*Thompson and Pollard*, 1995]. A region of enhanced ocean heat flux during winter in the Nordic Seas was added to prevent unrealistic buildup of sea ice. This flux warms the mixed layer whenever surface water temperature falls below 1.04°C in a rectangular region between 66° and 78°N and -10° and 56°E, and increases linearly to 500 W/m^2 if the ocean cools to its freezing point (-1.96°C) [*Thompson and Pollard*, 1995]. This simulates the buffering effect of the deepening mixed layer in winter, and advection of heat by warm ocean currents [e.g., *Hibler and Bryan*, 1987]. At each time step after the Nordic Sea adjustment is made, an additive global adjustment is made between 55°N and 55°S to keep the global integral of ocean heat convergence at zero.

The present-day performance of the model is comparable to that of previous coarse-grid models with predicted sea surface temperatures [*Thompson and Pollard*, 1995]. The primary errors are too warm high-latitude land areas in summer associated with too few clouds, and too large a global precipitation value due to too active penetrative convection in the lower troposphere and an overly large value of the roughness length over open ocean (which affects evaporation rates). The model does predict reasonable values for surface temperature, diurnal ranges of temperature, atmospheric energy fluxes, strength of jet stream maxima, and locations of precipitation maxima.

The Younger Dryas experiments were designed for 12,000 calendar years ago, the approximate midpoint of the cool climate episode. Paleogeography (land-sea distribution, continental ice sheet distribution, and paleotopography) for 12,000 years ago was taken from *Peltier* [1994] and is shown in Figure 1. The present-day solar constant of 1370 W/m^2 and the Earth's orbital configuration for 12,000 years ago determined the latitudinal and seasonal distribution of insolation following *Berger* [1978]. Atmospheric CO_2 is set at 250 ppm from the Vostok ice core record [*Barnola et al.*, 1987] (present control value is 340 ppm). A global intermediate vegetation type (savannah) was selected along with intermediate values for soil texture and color.

The basis for the ocean heat transport sensitivity test is the enhanced Nordic Sea heat convergence in winter. In the first experiment the additional Nordic Sea heat flux is turned off (actually reduced to 10 W/m^2) in the model to simulate the NADW shutdown and the cold phase of the Younger Dryas. This experiment is termed "Nordic Sea Heat Flux Off" (Heat Off). The second experiment has the Nordic Sea heat flux turned on to simulate the surface effects of active NADW formation, and the warmer climate following the Younger Dryas (Preboreal). This experiment is termed "Nordic Sea Heat Flux On" (Heat On).

Model Results

For both experiments, the model was executed with an AGCM spectral resolution of R15 (4.5° latitude by 7.5° longitude) and a 2° by 2° surface model resolution. Unless otherwise noted, all fields are means averaged over 10 consecutive years of an equilibrated run.

Surface Temperature

The specification of Younger Dryas boundary conditions produces substantial change in global climate relative to the present day. Figure 2 shows zonally aver-

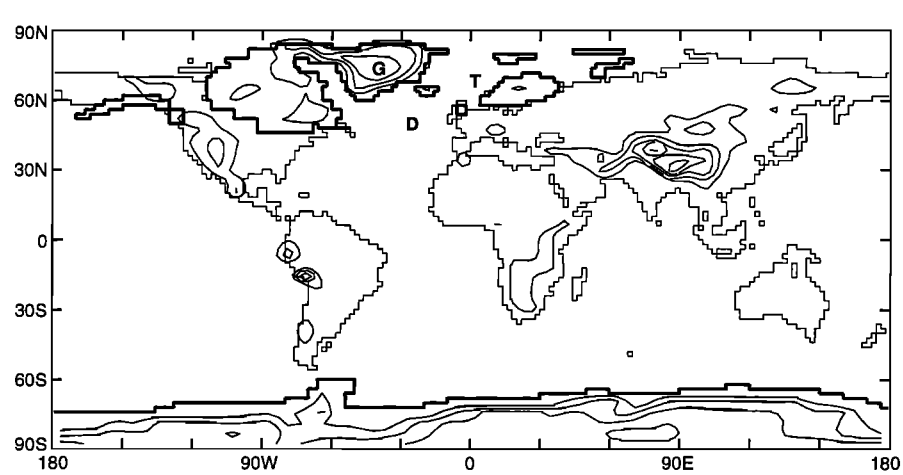

Figure 1. Paleogeographic map for 12,000 calendar years ago (adapted from *Peltier* [1994]) at the surface model resolution of 2° by 2°. Heavy lines show the extent of ice sheets; thin lines show topography with a 1000-m contour interval. G marks the GISP2 site, central Greenland, T is Troll 3.1 core, Norwegian Sea, and D is DSDP Site 609, North Atlantic.

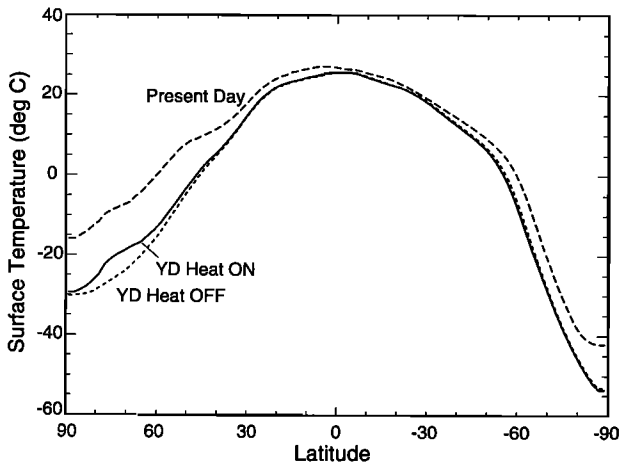

Figure 2. Zonally averaged surface temperature for three GENESIS experiments: the long-dashed line is the present-day control, the solid line is the YD Nordic Sea Heat On experiment, and the short-dashed line is the YD Nordic Sea Heat Off experiment.

aged annual temperature profiles for three simulations, the two Younger Dryas experiments and a present-day control experiment. The YD experiments are about 2°-3°C cooler in the tropics than the present day and show temperatures increasingly cooler than present ones toward the poles. The northern hemisphere cooled more than the southern hemisphere relative to the present day in both YD experiments due to the specification of the Laurentide and Fennoscandian ice sheets.

Temperature differences between the Younger Dryas Heat Off and Heat On experiments are largest between 60° and 80°N. In the southern hemisphere, temperature differences are approximately a degree or less. The Heat On experiment is warmer than the Heat Off experiment in the northern hemisphere and cooler in the southern hemisphere.

Figures 3 and 4 show the predicted December-January-February (DJF) and June-July-August (JJA) surface temperatures for the YD Heat Off and Heat On experiments, respectively. The seasonal temperature structure is very similar in both experiments for much

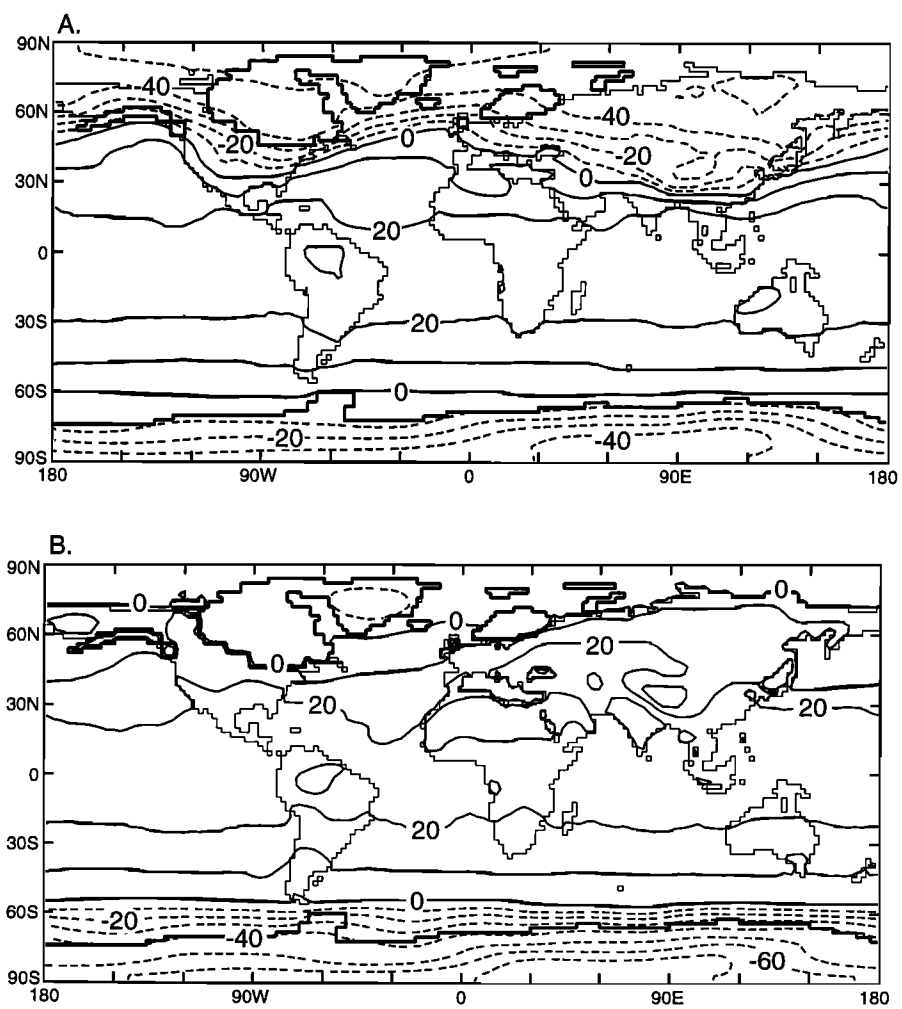

Figure 3. Model-predicted surface temperature (10°C contour interval) for the YD Heat Off experiment: (a) December-January-February (DJF) and (b) June-July-August (JJA).

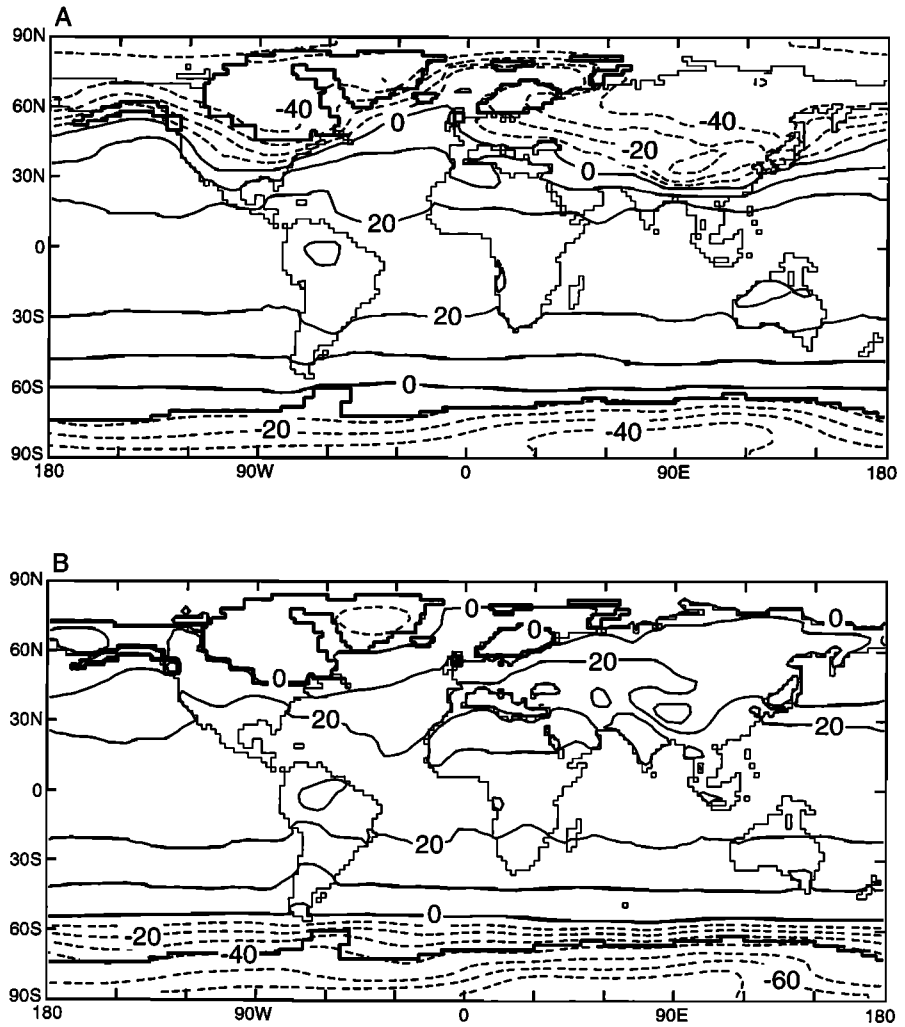

Figure 4. Model predicted surface temperature (10°C contour interval) for the YD Heat On experiment: (a) DJF and (b) JJA.

of the globe with the exception of the North Atlantic region. In the Heat Off case, the DJF 0°C isotherm runs almost east-west, from the U.S. mid-Atlantic coast to the southern British Isles. A sharp temperature gradient exists over the North Atlantic and temperatures of -40° to -60°C are found in northern North America and Greenland. The summer temperature pattern over North America is complex, with subfreezing temperatures over the specified ice sheets and very sharp gradients along the ice margins. Central Greenland temperatures are the coldest in the northern hemisphere. In the Heat On experiment, the DJF 0°C isotherm runs northeast from the mid-Atlantic coast to north of the British Isles. Large portions of the Norwegian and Barents Seas are at, or just slightly below, 0°C. The sharpest temperature gradients occur in eastern North America along the SE margin of the Laurentide ice sheet, and along the SE Greenland coast. North American and Greenland interior temperatures range from -40° to -50°C. In North America a very sharp temperature contrast runs along the Laurentide ice sheet margin with subfreezing temperatures over the ice itself and temperatures of 10°C and higher south of the ice margin. Central Greenland again has the coldest northern hemisphere summer temperatures.

The difference in annual average surface temperature between the two YD experiments is shown in Figure 5; values from the Heat Off run are subtracted from the Heat On run values. There is a large positive difference in temperature in the circum-Atlantic region which reaches a maximum of 25°C in the Barents Sea region. Much smaller but hemisphere-wide differences are also apparent, with the northern hemisphere slightly warmer and the southern hemisphere slightly cooler than in the Heat On experiment (see also Figure 2). Only a small portion of western North America is warmer in the Heat Off simulation. The largest temperature differences are centered in the region of the ocean heat flux forcing

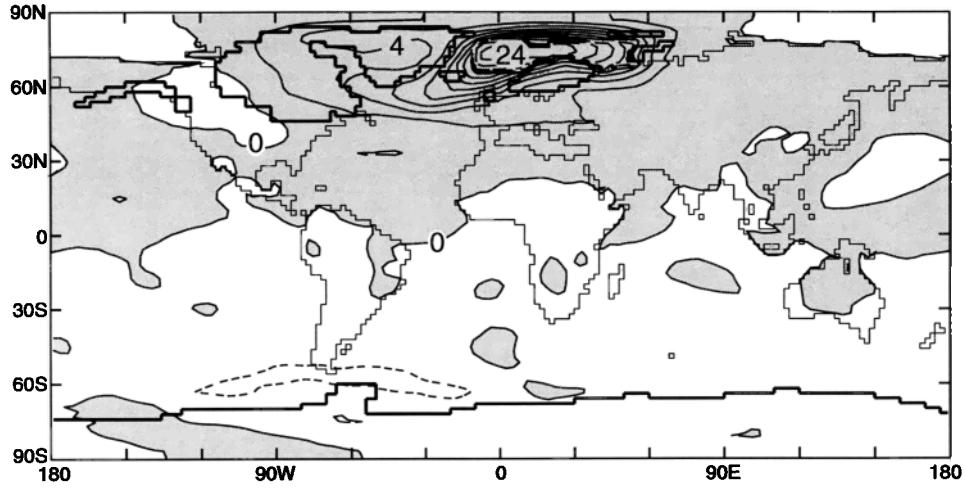

Figure 5. Annual average surface temperature differences (Heat On minus Heat Off). Positive differences are shaded, contour interval is 2°. Negative difference contours are dashed.

and are easily understood in this context; the smaller but more widespread differences are more complex and will be discussed below.

Temperatures at the GISP2 Site

The predicted surface temperatures for the GISP2 site in central Greenland (76.2°N, 38.5°W) for both YD experiments are averages of the values from the four surface grid cells surrounding the site. Table 1 shows the actual model grid point temperatures and temperatures that are corrected for a lowered Greenland summit. GCMs that use a spectral method of solution must first smooth the surface topography which usually results in a small adjustment to the surface elevation. The surface of central Greenland, however, is lowered a full kilometer (from 3 km to 2 km) because of the steep slope of the ice sheet margins. We apply a correction using the dry adiabatic lapse rate of 10°C/km to the predicted surface temperature values, and reduce predicted precipitation values by 4%/°C (see Table 2). With a 1-km elevation difference, surface temperatures are reduced by 10°C and precipitation is reduced to 66.5% of the original rate.

The annual average surface temperature difference (Heat On versus Heat Off experiments) for the GISP2 site is 2.8°C. This difference is strongly seasonal, with much of it occurring in winter and virtually identical summer (JJA) temperatures in both experiments (Figure 6).

Sea Ice

In the GENESIS climate model, sea ice forms whenever the ocean surface temperature drops below -1.96°C. The large differences in North Atlantic ocean surface temperatures between the two YD experiments do generate significant differences in the sea ice. In the Heat Off case, the DJF ice margin runs from northern Newfoundland to the northern tip of the British Isles (Figure 7) and the JJA margin lies slightly north of its winter position. The Heat On experiment has a DJF sea ice margin that runs from Newfoundland to Iceland and then north across the northern part of the Barents Sea

Table 1. GENESIS-predicted surface temperatures for the Greenland summit

	Model Predicted Surface Temperatures, °C		
Younger Dryas Experiment	DJF	JJA	Annual
Heat Flux Off	-55.53 (-65.53)	-14.93 (-24.93)	-38.94 (-48.94)
Heat Flux On	-49.88 (-59.88)	-14.27 (-24.27)	-36.17 (-46.17)
Temperature difference	5.65	0.66	2.77

Surface temperature values corrected for elevation are given in parentheses.
DJF, December-January-February; JJA, June-July-August.

Table 2. GENESIS-predicted precipitation values for the Greenland summit

Younger Dryas Experiment	Model Predicted Precipitation, cm/month		
	DJF	JJA	Annual
Heat Flux Off	0.62 (0.41)	3.34 (2.22)	18.48 (12.29)
Heat Flux On	1.07 (0.71)	3.52 (2.34)	21.08 (14.02)
Precipitation difference	0.45 (0.30)	0.18 (0.12)	2.70 (1.79)
Increase	(+73%)	(+5%)	(+15%)

Precipitation values corrected for elevation are given in parentheses.
DJF, December-January-February; JJA, June-July-August.

(Figure 7). The JJA margin also lies just northward of its winter position. The major difference between the two simulations is in the much wider extent of sea ice in the Norwegian and Barents Seas in the colder experiment (Heat Off) and is a direct result of the specified ocean heat flux forcing. The distribution of sea ice in these two experiments strongly affects the nature of the atmospheric circulation in the North Atlantic.

Winter Storm Tracks

Synoptic-scale cyclones are the dominant contributors of precipitation to Greenland, accounting for 90% of the total moisture convergence [*Robasky and Bromwich*, 1994]. Winter storms are closely tied to the jet stream, which occurs where the average tropospheric meridional temperature gradient is greatest. The strongest temperature contrasts and highest wind speeds in winter occur at the eastern edges of continents, and as we have noted in the case of the Younger Dryas time period, this contrast is enhanced by the presence of the Laurentide ice sheet.

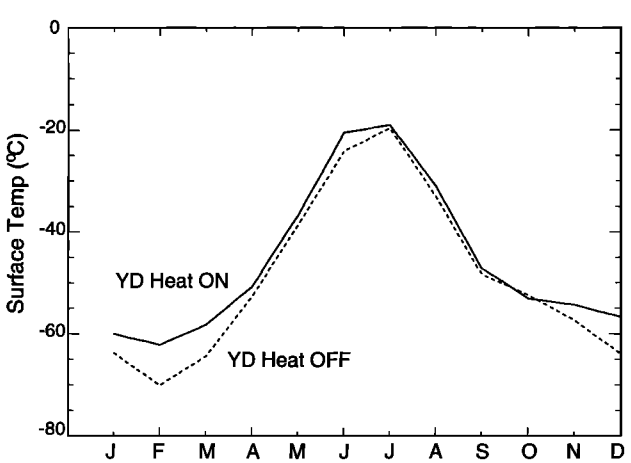

Figure 6. Model predicted annual surface temperature cycle at the GISP2 site, Greenland, for the YD Heat On (solid line) and the YD Heat Off (dashed line) experiments.

We determine the position of the winter storm tracks in the Younger Dryas simulations by taking the standard deviation of the 500-mbar geopotential height field, following *Blackmon et al.* [1976]. For mid latitude transient events, a time filter of 2.5 to 6 days is normally applied to give a good measure of regions where strong high and low pressure systems pass frequently. At higher latitude regions such as Greenland, however, cyclonic disturbances can persist for periods longer than 6 days, and so we apply a lower-frequency filter of 2 to 20 days for better resolution of the storm tracks. Variances in the geopotential height of 80 m and greater are used to define the position of the storm track.

The northern-hemisphere winter storm track for the YD Heat Off experiment is a prominent feature which extends from eastern Siberia across much of northern North America and across the North Atlantic into northern Europe (Figure 8). The region of highest baroclinic activity is anchored on the southeastern margin of the Laurentide ice sheet, and this extends across the Atlantic towards Iceland. Central Greenland has significant storm activity, but the major axis of the Atlantic storm track lies to the southeast, just poleward of the DJF sea ice margin.

The winter storm track for the Heat On experiment (Figure 8) is reduced both in extent and intensity compared with that of the Heat Off experiment, yet it remains a very prominent feature. It also begins in eastern Siberia and crosses into northern North America but ends in the North Atlantic. Again, the most intense portion of the storm track is anchored on the southeast margin of the Laurentide ice sheet, although it does not show quite as much variability in the height field as in the colder simulation. The axis of the North Atlantic storm track extends from this position to the northeast toward Greenland, roughly following the DJF sea ice margin. Central Greenland shows more variance in geopotential height in the Heat On simulation than in the Heat Off run, consistent with a higher frequency of

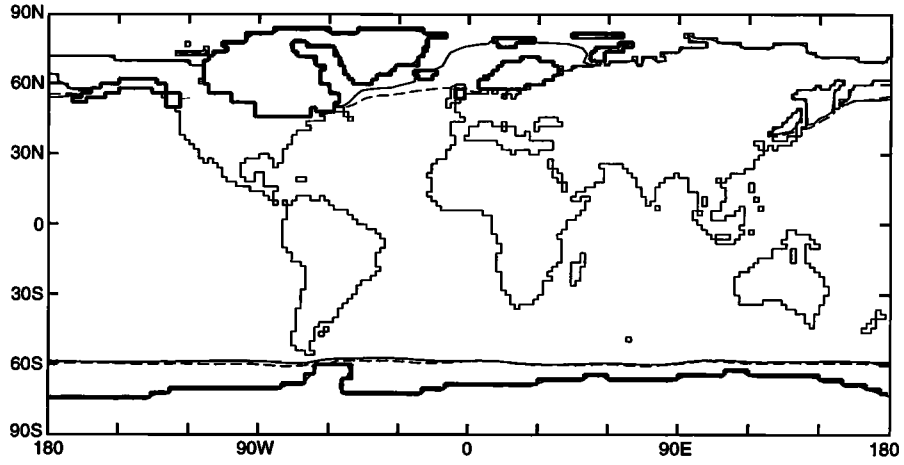

Figure 7. Model predicted DJF sea ice distribution (50-cm thickness) for the YD Heat Off (dashed line) and the YD Heat On (solid line) experiments.

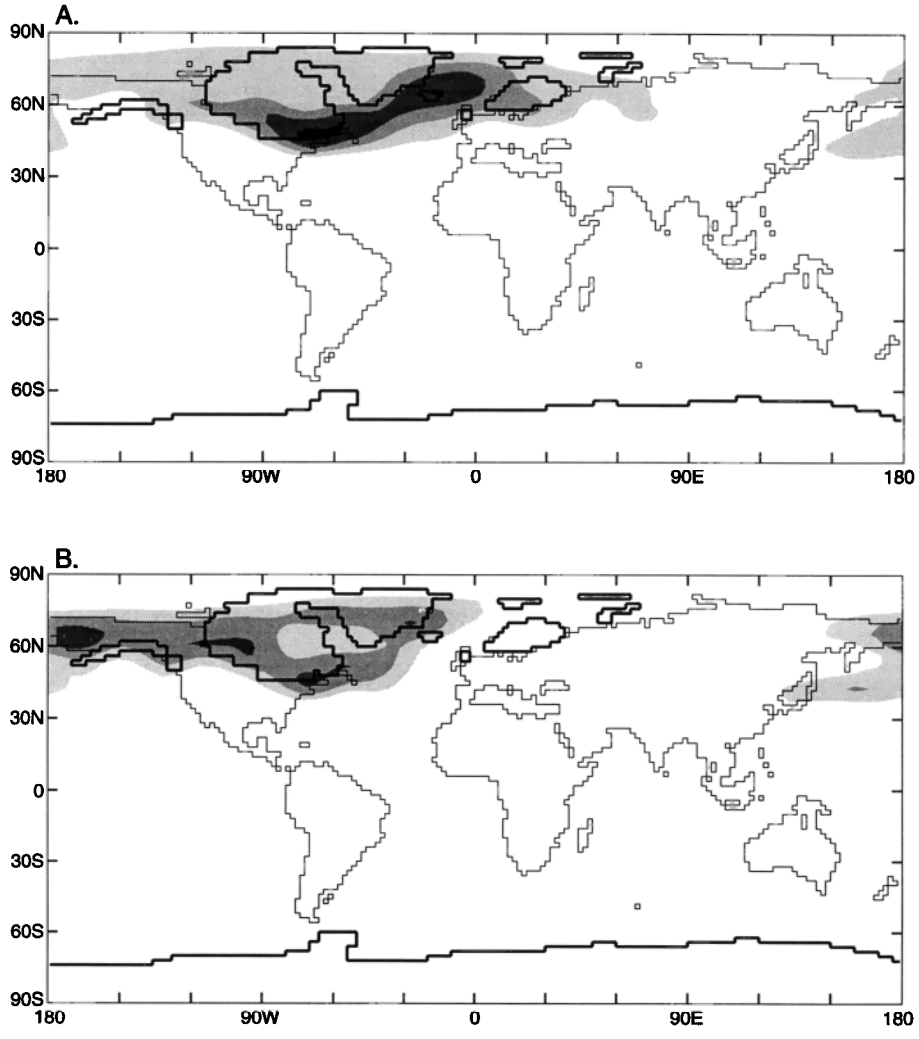

Figure 8. Winter (DJF) storm tracks (standard deviation of 2 to 20 day time-filtered geopotential height field) for 5 years of the (a) YD Heat Off and (b) YD Heat On experiments. Contours start at 80 m with a 10-m interval.

cyclones reaching the interior. A second storm track affecting Greenland lies farther to the north, running from the north central part of the Laurentide ice sheet across central Greenland. A relative minimum in height field variance lies between these two axes, over the Labrador Sea. The explanation for this may be a northward propagation of cyclones through the Nordic Seas to the higher latitudes and then a westward drift across northern and central Greenland. In the Heat Off case the more zonal temperature pattern apparently blocks the northward propagation of cyclones through the ice-covered Nordic Seas. There are significant differences in the path of winter storms between the two YD simulations, with more storm activity over central Greenland in the Heat On case. The maximum intensity of northern hemisphere winter storm activity is greater in the Heat Off simulation.

Precipitation

The predicted DJF precipitation fields for both experiments are shown in Figure 9. In both cases, precipitation is low over the northern hemisphere continents with the exception of northwestern North America. The precipitation maxima over the North Atlantic follow their respective winter storm tracks (south of the main axis of the tracks) with Greenland receiving more precipitation in the YD Heat On experiment. In both experiments, a JJA North American precipitation belt follows the southern margin of the Laurentide ice sheet, and the Middle East receives summer monsoonal precipitation (a result similar to that of *COHMAP Members* [1988]).

Annual precipitation differences between the two YD experiments are shown in Figure 10; values from the Heat Off run are subtracted from the Heat On run. The two regions of greatest difference in precipitation are over the Norwegian/Barents Seas and the tropics. The large changes in the tropics are consistent with a northward shift of the intertropical convergence zone (ITCZ) in the Heat On experiment. Northwestern North America receives more precipitation in the Heat Off case than in the Heat On; this is one of the few locations in the

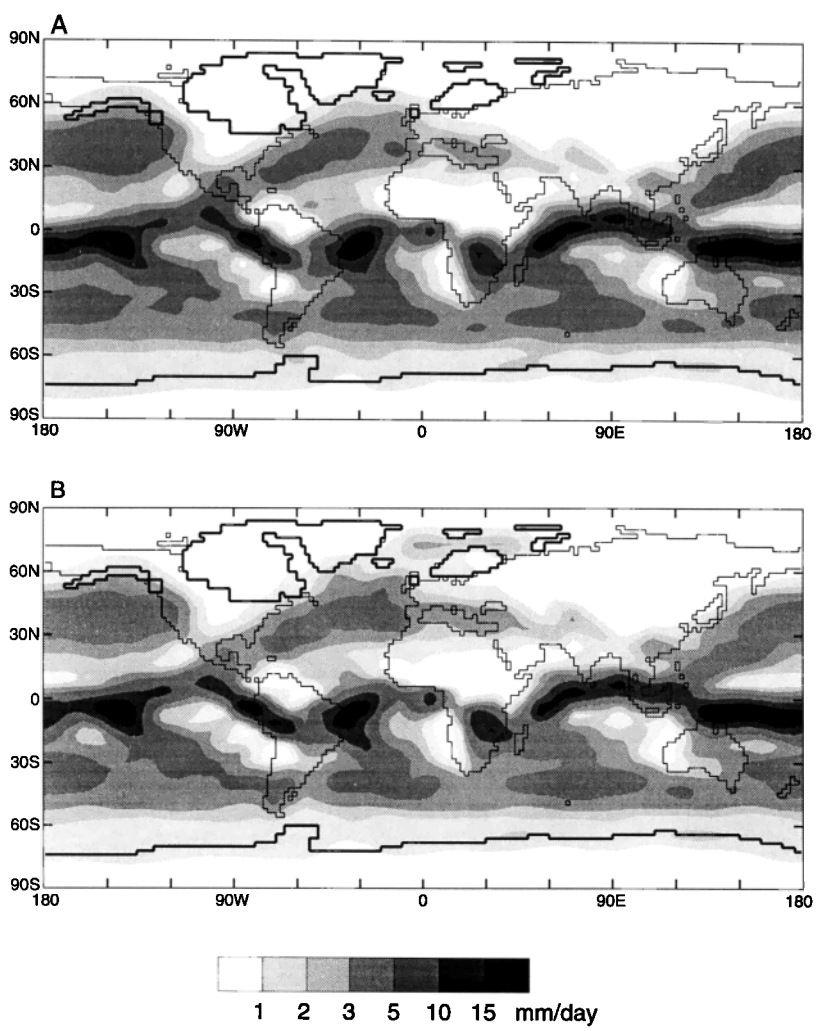

Figure 9. Model predicted DJF precipitation rates (millimeters per day, variable contour interval) for the (a) YD Heat Off and (b) YD Heat On experiments.

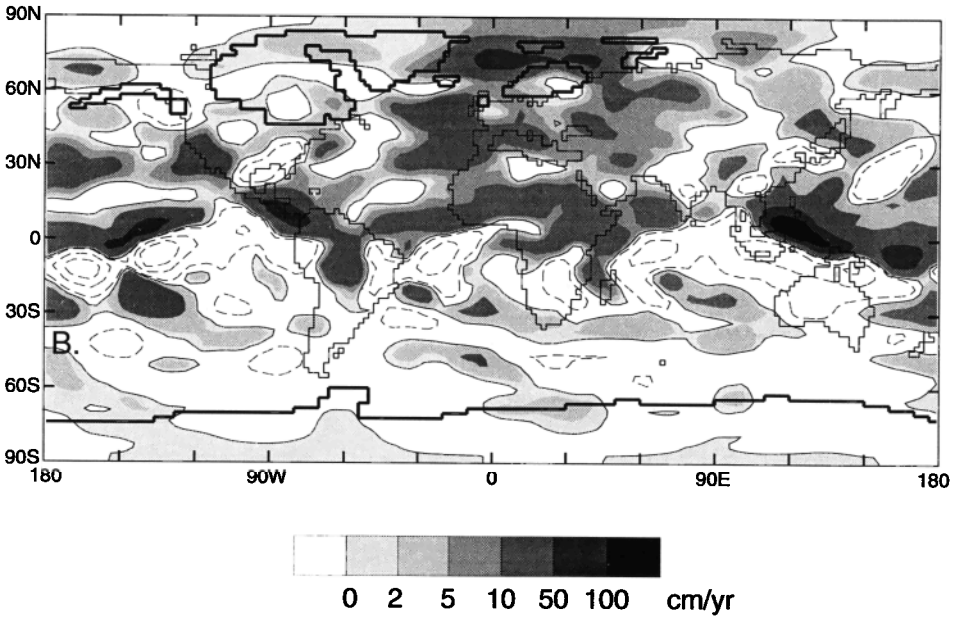

Figure 10. Annual average precipitation rate differences (centimeters per year, variable contour interval), YD Heat On minus YD Heat Off experiments. Positive differences are shaded. Negative contours are -5, -50 and -100 cm/yr.

northern hemisphere where this occurs. Precipitation rates are higher over central and northern Greenland and over the Arctic Islands in the Heat On experiment and are about the same in both experiments in southern Greenland.

Precipitation at the GISP2 Site

Predicted values for precipitation at the GISP2 site with the elevation correction are given in Table 2. On an annual average basis, the change in precipitation from the Heat Off to Heat On experiment is small with an increase of 1.5 cm/yr, or 15%. There is a very strong seasonal component to this annual average, with most of the increase occurring in winter (Figure 11).

Discussion

We have conducted two experiments representing the Younger Dryas to Preboreal climate transition, in which the amount of ocean heat convergence specified in the Nordic Sea is varied. This isolates the predicted response of Greenland's climate to a sudden reinitiation of North Atlantic Deep Water formation and associated changes in ocean surface heat flux. Comparison of the predicted climate change with the actual climate change reconstructed from the GISP2 ice core should allow an assessment of how much of the change can be attributed to the turn-on of NADW formation. However, before we can place confidence in these results, we need to assess how well the GENESIS model performs for the present day at the summit of Greenland.

Performance of GENESIS for the Modern GISP2 Site

The surface temperature record of the GISP2 site has been obtained from long-term satellite passive microwave brightness temperature trends and short-term automatic weather station data [*Shuman et al.*, 1995].

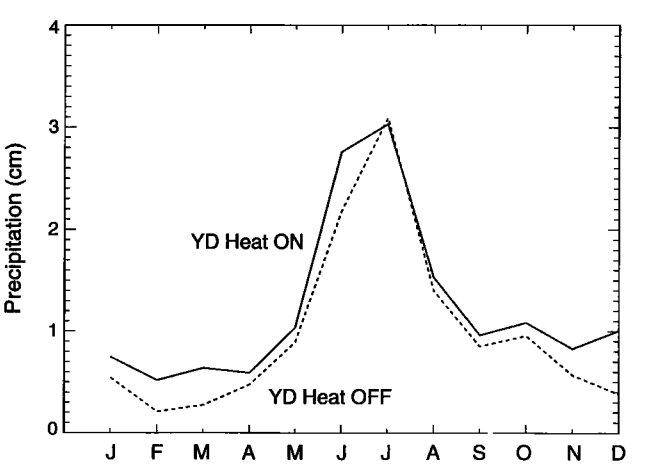

Figure 11. Model-predicted annual precipitation cycle at the GISP2 site, Greenland for the YD Heat On (solid line) and YD Heat Off (dashed line) experiments.

An interesting feature shown by these records is that the summit area experiences prominent secondary warm periods in late fall/early winter in addition to the primary summer warm period. The annual average temperature at GISP2 is -31°C and the annual temperature cycle ranges from -46°C in winter to about -8°C in summer - a range of 38°C (Figure 12a; data from *Shuman et al.* [1996]).

A present-day GENESIS control experiment with the Nordic Sea Heat Flux On, a 2° by 2° surface resolution, and a 4.5° by 7.5° atmospheric resolution has a predicted annual average surface temperature for the summit of Greenland of -16.0°C. The topography of the Greenland ice sheet is also truncated by a full kilometer in this experiment, so with the elevation correction the mean annual temperature is -26.0°C. This is 5°C warmer than is observed, which is consistent with the *Thompson and Pollard* [1995] observation that high-latitude areas are in general predicted to be too warm in GENESIS. The annual range in temperature is quite well simulated, however, ranging from -46°C in winter to -11°C in summer (Figure 12b), a 35°C annual range. The late fall/early winter secondary warm intervals found in the Greenland temperature records are well predicted by the model (Figure 12) with similar timing and magnitudes. Thus we conclude that GENESIS is able to simulate the temperature structure for central Greenland including both the annual temperature range and the secondary warm interval, to within a few degrees of error.

The modern annual accumulation rate at GISP2 is 24.7 cm/yr [*Alley et al.*, 1993]. Accumulation rates have a low seasonality with winter averages of 1.5 to 2 cm/month and summer averages of approximately 2.5 cm/month [*Bromwich et al.*, 1993; *Shuman et al.*, 1995]. The GENESIS control experiment predicts annual precipitation rates of 35.1 cm/yr for the Greenland summit, which is about 40% higher than observed. There is a strong annual cycle in predicted precipitation with winter values of 2 cm/month (close to observed) and summer values of 5 to 7 cm/month (much higher than observed). The overprediction in summer precipitation is a result of the model overestimation of evaporation rates over the open ocean (the large roughness length discussed above). The error in evaporation rates is more pronounced in summer, when surface temperatures are high over the moisture source regions for Greenland. In winter, when the source temperatures are cooler, less evaporation occurs and the error is diminished. Thus we conclude that while the geographic patterns of precipitation are largely well simulated [*Thompson and Pollard*, 1995], the seasonal amplitude of precipitation predicted for the summit of Greenland is too large, with winter values close to, and summer values higher than observed. We expect that a similar summer overprediction will occur for other experiments using this model version.

Younger Dryas - Preboreal Climate Differences

The Nordic Sea heat flux experiments predict a substantial climate change in Greenland as a consequence of switching on North Atlantic Deep Water formation. The annual average temperature increase from the Heat Off to Heat On experiments at the GISP2 site is 2.8°C (Table 1), which is approximately half of the temperature increase indicated by the ice core $\delta^{18}O$ profile. The actual increase in accumulation rate is 100%, much larger than the few tens of percent predicted.

The predicted temperature and precipitation differences are very seasonal in character (Tables 1 and 2; Figures 6 and 11). There is a large temperature increase in DJF and a small increase in JJA. Seasonal precipitation differences are even more pronounced, with the largest increase in DJF. The annual surface temperature ranges are larger for the Younger Dryas than for the present day because of the increased high-latitude summer insolation 12,000 years ago [*Berger*, 1978] so we expect that there will be enhanced precipitation seasonality for the YD also. In fact, there is a very pronounced seasonal cycle in precipitation for both YD experiments (Figure 11). However, the overprediction of summer precipitation in the present-day control experiment suggests that the predicted YD summer precipitation values are probably also too high and that

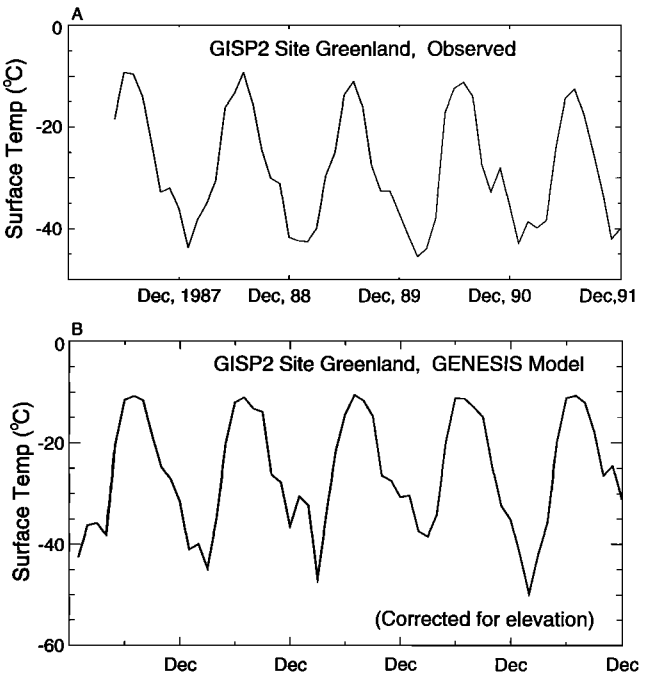

Figure 12. (a) Recent near-surface air temperature for the GISP2 site, Greenland (data from *Shuman et al.* [1996]) and (b) 5 years of model-predicted surface temperatures for the GISP2 site, Greenland, for the present-day control experiment. (Both are monthly averages).

the seasonal range in precipitation should be lower for both experiments. The relative contributions from the winter seasons to the annual precipitation totals should then be higher. As most of the precipitation increase from the Heat Off to the Heat On experiment occurs in winter, we should also expect the annual average precipitation difference to be larger and therefore closer to the observed value.

The predicted change in precipitation seasonality at the Younger Dryas to Preboreal climate transition has important implications for the isotopic record of temperature change. In the cold YD, low winter precipitation amounts bias the annual average $\delta^{18}O$ value towards summer when precipitation is higher (because it is the snowfall itself that contains the isotopic thermometer). In the warmer PB, winter precipitation increases markedly, and the thermometer samples winter temperatures more frequently. The change in $\delta^{18}O$ in the ice core will then underestimate the actual temperature increase, all else being equal [*Steig et al.*, 1994]. The significant differences between the modern spatial $\delta^{18}O$ - T calibration of 0.67 per mil per °C [*Johnsen et al.*, 1989] and the temporal $\delta^{18}O$ - borehole T calibrations of 0.51 and 0.33 per mil per °C [*Cuffey et al.*, 1994, 1995] may well result from this type of change in seasonality of precipitation.

Predictions of large winter changes in temperature and precipitation are not surprising given that the forcing factor, North Atlantic Deep Water formation, is a seasonal (winter) phenomenon. Present-day Europe is considerably warmer in winter lying downwind of this heat source than it would be otherwise. Greenland, however, lies upwind of this heat source. When NADW is "switched on," the increases in temperature and precipitation at the GISP2 site must be due to a more complex process than simple downwind (westerly) advection of heat and moisture.

The shifts in atmospheric circulation patterns, especially the winter storm tracks, from the Younger Dryas to Preboreal appear to be the explanation. In the Heat Off experiment, the North Atlantic storm track is very intense and extends east from the southern edge of the Laurentide ice sheet out into the Atlantic (Figure 7). Its counterpart in the Heat On experiment is not as intense, does not extend as far into the Atlantic, and curls up onto central and northern Greenland from the Norwegian Sea (Figure 7). The resulting change for central Greenland is increased regional winter storminess in concert with hemispherically reduced storminess. Both aspects of this prediction are well supported by analysis of the ice core. *Kapsner et al.* [1995] demonstrated that a storm track shift onto Greenland was required to explain the extra increase in accumulation relative to temperature across the YD-PB transition, and the abrupt decrease in dust and particulate loading in the ice across this transition [*Mayewski et al.*, 1993; *Taylor et al.*, 1993; *Alley et al.*, 1995] is consistent with reduced hemispheric storminess (and reduced peak wind speeds).

The spatial patterns of increased surface temperature and precipitation (annual difference plots in Figures 5 and 10) are consistent with the shift in winter storm tracks. For both climatic variables, the largest changes occur in northern Greenland and the magnitude of change decreases from east to west. This GISP2 site actually lies in a temperature difference minimum (Figure 5). In the Heat On experiment, winter storms drift westward across central to northern Greenland from the open Norwegian Sea, delivering more precipitation and warming these regions by advection of warmer air into the interior of the ice sheet and by increased cloud cover which traps surface- emitted longwave radiation. The trapped IR is a significant part of the radiation budget at these high latitudes because incoming winter insolation is essentially zero. There may also be some component of increased latent heating of the atmosphere associated with the storms, but this is probably subordinate to the other mechanisms.

The shift in storm tracks over Greenland explains why the winter temperature (and precipitation) differences are so large. The Heat Off experiment, in which winter storms track away from Greenland, has very cold winters and a reduced to nonexistent secondary warm peak (Figure 13). Winters in the Heat On experiment, where winter storms track close to Greenland, are considerably warmer and have the secondary warm peak noted in the present-day control experiment and in modern climate data. This difference in winter temperature structure shows that the shift in the North Atlantic storm track drives regional climate change in Greenland at the YD termination.

The high correspondence of the climate model results to the ice core data suggests that a sudden reinitiation

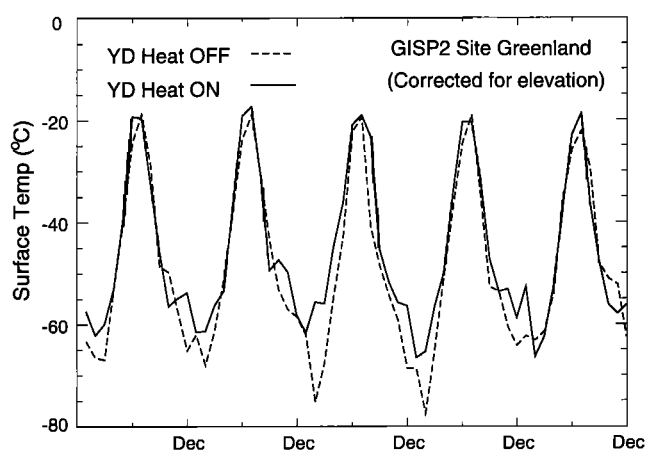

Figure 13. Five years of model predicted annual average surface temperature for the GISP2 site, Greenland, for the YD Heat Off (dashed line) and YD Heat On experiments.

of NADW formation after an extended period of shutdown is a very plausible explanation for the large and rapid climate change in Greenland at the end of the Younger Dryas. However, the results do not explain all of the observed climate change, and this raises three possibilities. (1) The hypothesis is incorrect, and the apparent success of the results is merely a chance occurrence. (2) The hypothesis is correct, but the model is not fully sensitive to the climate forcing. (3) The hypothesis is partially correct, but some other climate forcing factor(s) have also contributed to the climate change.

We reject the first possibility on the basis of widespread paleoceanographic evidence for NADW formation shutdown during the Younger Dryas, the bistable nature of deep ocean convection in the North Atlantic in a number of ocean circulation models, and the degree to which our atmospheric modeling results match patterns of change in the ice core record. This leads us to conclude that the model is not fully sensitive to this climate forcing and/or that other factors have acted in combination with the Atlantic thermohaline circulation.

The coarse atmospheric resolution of the GENESIS experiments may have contributed to an underestimation of the actual climate change. Close examination of Figures 5 and 10 shows that the GISP2 site actually lies in relative minima for both temperature and precipitation differences. As we have determined that these differences are largely due to storm track migration, a higher- resolution model experiment with a more realistic topography could change slightly the position of the Greenland winter storm tracks and increase the differences between the YD experiments.

For other climate forcing factors to be considered for the YD termination, two important criteria must be met. They must operate on a very rapid timescale and have a global effect. The Younger Dryas has long been recognized in terrestrial climate records of northwestern Europe, and more recently in Greenland ice cores and deep-sea sediment cores from the North Atlantic basin. With improved dating techniques, cooling events contemporaneous with the North Atlantic Younger Dryas have been recognized in a number of widespread localities (see *Peteet* [1995] for review): eastern and western North America, Alaska, the South American Andes, New Zealand, and the western Pacific Ocean. In East Africa a YD-termination-contemporaneous change from a more arid to a more humid climate is shown by dramatic changes in lake levels and salinities [*Street-Perrott and Perrott*, 1990; *Roberts et al.*, 1993]. If all of these events are causally related, then the forcing factor(s) affecting the North Atlantic climate must have had a direct global impact, or at least an indirect impact via climate feedbacks or teleconnections.

At the time of the YD termination, rapidly changing factors that might have had a global impact include changes in atmospheric greenhouse gases (CH_4, CO_2) and changes in atmospheric dust content (which can affect radiation balance through albedo effects and IR absorption). Ice core records from Greenland and Antarctica show increases from the YD to PB in both CH_4 and CO_2 [*Sowers and Bender*, 1995], although they are relatively small compared to longer-term variations (CO_2 increases by 10 ppmv and CH_4 increases by 220 ppbv). The increase in CH_4 is relatively larger than that for CO_2 and could have contributed a small amount to the YD end warming. The Greenland ice core record also shows a sharp decrease in atmospheric dust loading from YD to PB [*Taylor et al.*, 1993; *Mayewski et al.*, 1993; *Alley et al.*, 1995], but the actual climatic response to this change is not well understood. The Antarctic ice core records do not show an increase in atmospheric dust content during the YD [*Jouzel et al.*, 1995], so this climatic forcing appears to be restricted to the northern hemisphere.

The ocean circulation is another possible candidate for transmitting a climate signal around the globe. Our model results suggest that the rapid climate changes in tropical and subtropical Africa could have resulted from the small changes in tropical sea surface temperatures by a mechanism similar to that proposed by *Street-Perrott and Perrott* [1990]. The turnoff of NADW formation in the North Atlantic produces a cooling which extends well into the northern subtropics (Figure 5). This is coupled to reduced precipitation (and precipitation-evaporation) in large parts of Africa and central America (Figure 10) and is consistent with the lower lake levels in East Africa during the YD. When North Atlantic Deep Water formation resumes, the opposite pattern occurs with higher precipitation in East Africa. Though of marginal statistical difference, this result is strengthened by its strong agreement with the paleoclimatic data. The atmospheric methane variations recorded in the ice cores could have resulted in part from such hydrologic changes affecting tropical and subtropical vegetation.

The model results for southern hemisphere localities (Antarctica, New Zealand) are not consistent with the regional paleoclimatic records in that no pronounced cooling during the YD is predicted. This may be because we do not include the effects of CO_2 and CH_4 in the experiments, or because the actual climate signal involves ocean circulation and cannot be simulated without a true ocean general circulation model. If the interaction between NADW (slightly warmer and saltier) and southern ocean water masses changes significantly when deep water formation rates change in the North Atlantic, then it is possible that the southern hemisphere climate system could be affected. This idea cannot be tested without a more detailed model of ocean circulation.

Conclusions

A variety of paleoclimatic indicators in the GISP2 ice core show that the YD climate event ended abruptly in

the North Atlantic region, possibly in as little as 1 to 3 years. This termination is characterized by a 3.5 to 4.0 per mil increase in $\delta^{18}O$ of ice (which corresponds approximately to a 7°C temperature increase), a doubling of the summit accumulation rate, and an increase in regional storm activity but a decrease in storm activity in broader areas. A reorganization of the North Atlantic thermohaline circulation has been suggested to explain both the rapidity and the size of the YD termination.

The GENESIS climate model contains a provision for ocean heat transport which allows us to test the effects of NADW-formation-related experiments for YD conditions, one with low and one with high ocean heat flux in the Nordic Seas (representing deep water shutdown and reinitiation), were compared to the GISP2 ice core record of climate change. These results show the following:

1. Annual average surface temperature increases by 2.8°C at the Greenland summit when high ocean heat flux is applied in the Nordic Seas. This is approximately half of the temperature increase recorded in the ice core.

2. Annual average precipitation rate increases by a few tens of percent, as compared to the 100% increase in accumulation rate in the ice core.

3. Both of these increases have a pronounced seasonality, with most of the change occurring in winter; this result has implications for interpretation of ice core $\delta^{18}O$ paleothermometry.

4. The large increase in winter precipitation is "swamped" by an overprediction of summer precipitation. This explains, in part, the model underestimation of the annual accumulation rate increase.

5. North Atlantic sea ice extent is greatly reduced in the high Nordic Sea ocean heat flux experiment. The predicted changes in sea ice distribution roughly correspond to polar front migrations recorded in deep-sea sediment cores.

6. North Atlantic winter storm tracks shift toward Greenland in the warmer climate experiment (Heat On), away from the more zonal orientation in the cold climate experiment (Heat Off). They also weaken in intensity over North America and the western Atlantic, which is consistent with the reduced continental dust and sea salt loading in the ice cores.

7. Heightened storminess over Greenland in the warm climate experiment contributes to the large increase in predicted winter precipitation rates.

8. A present-day control experiment shows that the secondary late fall/winter temperature maximum in Greenland summit surface temperature data is caused by the passage of winter storms. The warmer temperatures result from heat advection into central Greenland and from increased cloud cover which traps surface-emitted IR.

9. This winter-storm-related secondary temperature maximum occurs at the GISP2 site in the (YD) warmer climate experiment, and does not in the cold climate experiment.

10. The GISP2 site climate changes (in temperature and precipitation) are not fully explained by the model experiments, indicating that the model is not fully sensitive to the ocean heat flux forcing and/or that some other climate forcing factor(s), have acted in concert with the switches in thermohaline circulation.

11. Abrupt changes in subtropical hydrologic balances could have been caused by reorganizations in the ocean circulation system which act to modify tropical sea surface temperatures. A model-predicted increase in East African precipitation when high Nordic Seas heat flux is applied, and cross-equatorial ocean heat transport increases to compensate, is consistent with higher lake levels in that region at the end of the YD.

12. The predicted changes in subtropical hydrologic balance should have affected tropical and subtropical vegetation and could have contributed to the higher atmospheric CH_4 levels at the close of the YD.

Together, these results strongly support the hypothesis that a change in oceanic heat flux in the Nordic Seas is linked to the Younger Dryas but suggest that processes or feedbacks not modeled here were also important.

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