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CLIMATIC EFFECTS OF REDUCED ARCTIC SEA ICE LIMITS IN THE GISS II GENERAL CIRCULATION MODEL

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Abstract. In this paper we present results of an atmospheric general circulation model (GCM) experiment in which Arctic sea ice limits were substantially reduced in all months. March sea ice limits were set equivalent to modern September limits, and all sea ice was removed in September. Sea ice coverage for other months varied between these two extremes. This climate sensitivity experiment makes predictions about mean northern hemisphere atmospheric conditions (including temperature, pressure, wind patterns, and precipitation) consistent with these boundary constraints. The major effects of reduced sea ice limits are observed in winter. They include large regional warming of the circum-Arctic region, northward migration of the Icelandic low pressure system, and strengthening of the Azores high. Changes in net heating over the North Atlantic Ocean suggest that increases in sea surface temperatures and salinities in this region would also accompany reductions in Arctic sea ice limits. In the wind field, a weakening of the polar easterlies and an intensification of cyclonic circulation over the Norwegian-Greenland Sea suggest that surface water exchange between the Atlantic and Arctic would increase when sea ice limits are reduced. However, zonally averaged changes in strength of the westerlies or upper level jet stream are minimal. The late Pliocene cooling of the North Atlantic Ocean and North American Arctic margin may have been linked in part to the development of perennial sea ice cover.

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INTRODUCTION

For over three decades, the history of Arctic Ocean sea ice cover and its role in controlling global climatic variations has been a topic of active research. Ewing and Donn [1956] and Donn and Ewing [1966] proposed that the Arctic played a critical role in driving the waxing and waning of the late Pleistocene ice sheets. Their hypothesis that the Arctic Ocean is ice-free early in glacial periods and acts as a moisture source for ice sheet growth was eventually abandoned, in large part because no evidence for an ice-free Arctic could be found for the last few glacial cycles.

In the following decades, much of the work in this field has been carried out by two researchers in particular, D. Clark and Y. Herman, who came to almost opposite conclusions as to the history of Arctic ice cover. Clark et al. [1980] and Clark [1982a, b] proposed that the Arctic Ocean has been ice-covered for the last 5 m.y., with sea ice extent and thickness never less than is observed today. This conclusion was later modified by Gilbert and Clark [1982/1983], who proposed that the Arctic was characterized by very thin or even periodically absent sea ice between 1.8 and 0.7 Ma but with perennial ice cover prior to 1.8 Ma and after 0.7 Ma. These suggestions contrast with those of Herman and Hopkins [1980], who proposed that perennial Arctic ice cover first formed about 0.7 Ma.

The views of Herman and Hopkins seem to be more in agreement with geologic evidence on land. Repenning [1983], Funder et al. [1985], Carter et al. [1986], and Repenning et al. [1987] all present data suggesting that the margins of the Arctic were seasonally ice-free in the late Pliocene, possibly until 1.6 Ma. Scattered floral and faunal data indicate that climate conditions were often warmer than at present. Whether these warm periods alternated with intervals of colder climate distinguished by perennial sea ice is uncertain. Records from deep-sea cores indicate that from 2.8 to 1.6 Ma continental ice sheets and deep ocean temperature varied with a predominately 41,000-year rhythm characteristic of changes in the Earth's obliquity. Whether Arctic sea ice limits also responded to variations in high-latitude radiation has been difficult to establish due to the paucity of well-dated continuous records from the Arctic Ocean. In addition, the pre-Pleistocene geologic record from land, which gives information regarding the history of the ocean margins, is discontinuous and could easily be biased towards preservation during warm intervals.

It is probable that two important steps occurred in the long-term evolution of perennial Arctic sea ice cover: the development of year-round sea ice during colder (glacial) periods of insolation/obliquity minima, and the further establishment of permanent year-round sea ice during warm (interglacial) periods of insolation/obliquity maxima. By the latest Pleistocene, perennial sea ice appears to have become firmly established within the Arctic, characterizing both interglacial and glacial periods [Scott et al., 1989]. However, when this final step in the evolution of modern Arctic sea ice cover occurred is uncertain. According to a summary of Arctic geologic data by Repenning et al. [1987], modern Arctic Ocean sea ice limits could have developed either at the beginning of the Eubronian cold period in Europe (around the Olduvai subchron, ~1.7 Ma) or possibly as late as the Brunhes magnetic chron (~0.7 Ma) when Herman and Hopkins [1980] report a substantial increase in Arctic Ocean ice rafting.

Regardless of the disagreement and uncertainty over when present-day perennial sea ice cover formed, few dispute that the Arctic ice sheet must play an important role in controlling the thermal regime of the Arctic and closely adjoining polar regions. In this paper, we explore the role of variable Arctic sea ice extent as a climate forcing factor in locations more remote from the Arctic. Paleoceanographic studies [Raymo et al., 1986; Loubere and Moss, 1986] show that prior to the 2.65-2.40 m.y. interval when large ice sheets first accumulated in the northern hemisphere [Shackleton et al., 1984; Raymo et al., 1989], significant variations in North Atlantic foraminiferal fauna were occurring at Milankovitch-scale frequencies, suggesting that sea surface temperature (SST) variations were also occurring. The primary purpose of this GCM experiment is not to explain the factors that produced the sea ice changes, but to determine whether large-scale variations in the extent of Arctic ice cover, perhaps driven by orbital insolation variations, can climatically impact areas remote from this region, in particular the North Atlantic Ocean. Our approach of isolating the effects of a single boundary condition change (sea ice cover) provides a deeper understanding of the unique impact of one component of the climate system as well as the feedbacks that exist within the overall climate system.

MODEL

The global atmospheric circulation model used in this experiment is described in detail by Hansen et al. [1983]. This model (the GISS II GCM) solves simultaneous conservation equations for mass, energy, momentum, and moisture. The model consists of 8° x 10° (latitude by longitude) horizontal grid boxes with nine atmospheric layers. Topographic heights are specified as the area-weighted mean of each grid box which is fractionally divided into ocean, land (based on the 1° resolution Scripps topography [Gates and Nelson, 1975]), ocean ice, and land ice.

The radiation calculation incorporates the effects of all significant atmospheric gases and aerosols, and includes the effects of calculated cloud coverage and height. Diurnal and seasonal cycles are modeled. Ocean albedo varies with surface wind speed and solar zenith angle. Land albedo (as well as ground hydrology) is a function of vegetation type and snow cover, both of which can vary seasonally. The effects on surface albedo of snow age, depth, and masking by vegetation are also specifically modeled. Snowfall occurs in the model when the lowest level air temperatures are <0°C and snow accumulates if the ground temperature is ≤0°C. In the control run, ocean temperatures are prescribed based on monthy mean values [Robinson and Bauer, 1981] linearly interpolated from the middle of each month; sea ice coverage is based on mean ice conditions for the years 1953-1977 [Walsh and Johnson, 1979].

Modeled cloud cover in the Arctic was compared to observations by Barry et al. [1986] who concluded that it was realistic. In addition, the model uses a smoothing technique at high latitudes which removes very small scale noise but results in too much transient eddy energy right near the poles. This eddy energy does not appear to be associated with transports so its influence in this experiment is probably negligible. A complete evaluation of the control run used for this experiment, including the effects of the spatial differencing scheme used at the poles [Arakawa, 1972], is given by Hansen et al. [1983]. In the following sections, we will focus primarily on the significant differences between the GISS II GCM control run and the experiment in which we imposed major reductions in Arctic sea ice extents.

EXPERIMENT

In the experiment described here, significant reductions in Arctic Ocean sea ice were prescribed for all months. In addition, many of the bays and waterways of the Canadian Archepelago are also generally ice-free in the model, unlike today when these areas are almost entirely locked in year-round ice. Presently, mean annual minimum and maximum sea ice extents are in September and March, respectively. In the experiment, these months are also considered to be the seasonal extremes: all northern hemisphere sea ice is removed in September; in March, sea ice limits are set equivalent to modern September limits (Figure 1). This choice is consistent with Pliocene sea otter remains recovered from northern Alaska [Carter et al., 1986] which suggest that winter ice limits during at least some intervals in the late Pliocene may have been very close to modern summer limits.

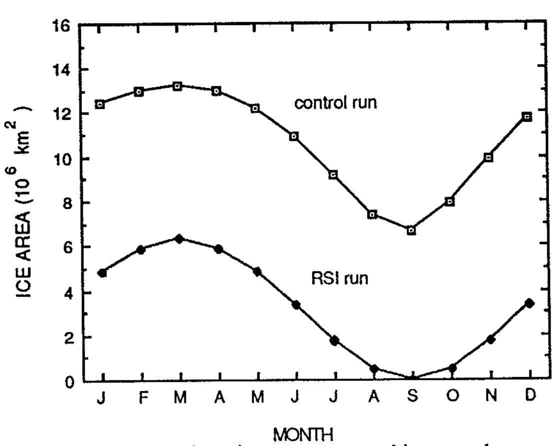

Fig. 1. Monthly areal sea ice coverage used in control run and experiment with reduced sea ice (RSI) limits.

Because there is no multiyear ice in the model (i.e., all ice melts in September), ice thickness (z_{ice}) is always 1 m, consistent with observed thicknesses of modern first-year ice. Fractional open water, or percent leads, is defined by $f_o = 0.1/z_{ice}$ and thus is equal to 10% in the experiment. The run covers the full seasonal cycle, with imposed sea ice limits varying sinusoidally between seasonal extremes in March and September (Figure 1).

In the control run (present conditions), the SST of open water, including leads, is typically -2°C in areas of extensive sea ice cover. In the experiment, we imposed very cold SSTs (-1°C) in ocean areas from which the sea ice was removed. Elsewhere SSTs remained at present-day prescribed values. We did this to focus the sensitivity test on only the climatic impact of removing sea ice, rather than on the degree of warming of the ice-free waters. All other initial conditions in this experiment, including vegetation, were prescribed at modern values.

The model was initialized from a five year control run and run for 2 years; the length of the control and experimental runs are sufficient given that SSTs are specified and not allowed to change. In addition, we discuss only those model results which are several times the level of normal fluctuation in the 5-year control run (>2 s.d.) and are similar in both years. Because sea ice and SSTs are prescribed and cannot come into equilibrium with the atmosphere, the interpretation of the results is limited. By not allowing SSTs to increase in the Arctic, we may be minimizing the impact of the imposed sea ice changes. However, while the absolute magnitude of predicted changes may be uncertain, the relative magnitude and direction of the climate changes away from the Arctic can be estimated from the resulting energy imbalances, as discussed below. In addition, the lack of dynamics implies that the model cannot distinguish, for instance, whether large negative heat fluxes in an ice-free Arctic would be sustained by ocean heat transports from lower latitudes, or whether cooling of the water column would eventually result in reformation of ice. Here, we report the first-order atmospheric effects of major reductions in Arctic sea ice, recognizing that the coupling of this model to an interactive ocean-ice model will improve the predictive qualities of this experiment.

RESULTS OF THE EXPERIMENT

In this section, we outline the major changes in atmospheric circulation and climate that result from the imposed reductions in sea ice. We refer to this experiment as the RSI run, for "reduced sea ice." Results for winter (December, January, and February) are discussed first, followed by results for summer (June, July, and August).

Winter

Figures 2a and 2b show average winter surface air temperatures for the control and RSI runs; Figure 2c shows the difference in winter surface air temperatures between the two runs. The major effects of reduced sea ice limits are restricted to the high northern latitudes. Not surprisingly, surface temperature increases greater than 16°C occur over the Arctic Ocean and Baffin Island. The northern Norwegian-Greenland seas warms by 2°C in the south to over 20°C in the north and the Labrador Sea area warms by up to 12°C in the RSI run. Little change is observed over the northern Atlantic or Pacific oceans when sea ice limits are reduced.

Over land, a warming of 4°-18°C occurs over northern Canada and Alaska, as well as over the northern coast of Asia, with progressively smaller temperature effects towards the continental interiors. Except for a 2°-4°C warming over northern Scandanavia when sea ice is reduced, no significant surface air temperature changes are observed over Europe. Greenland temperatures increase by up to 14°C, with the least change in the southeast section of the island. Other than a warming of the polar air masses that penetrate southeastward over east central Canada, no significant changes in surface air temperature are observed at mid-latitudes. The rapid drop off with distance of the temperature anomaly (Figure 3) is consistent with previous modeling results [North, 1984; Manabe and Broccoli, 1985] and the suggestion of Hyde et al. [1989], that the e-folding distance of a temperature perturbation is approximately 2000 km in the absence of other radiative coupling (e.g., a CO_2 change).

The polar warming associated with reduced sea ice limits results in a major decrease in sea level pressure over the Arctic (Figures 4a-4c) in the RSI experiment. Sea level pressures are 4-10 mbar lower over the Arctic Ocean when sea ice limits are reduced, causing a significant weakening of the Arctic polar high. Another significant change is a weakening and northward movement of the Icelandic low. (Note the relative size and location of the 10 mbar (1010 mbar) contour in Figures 4a and 4b.) The northward movement results in a localized strengthening of the low over the Norwegian-Greenland Sea (Figure 4c).

The migration and overall weakening of the Icelandic low is associated with the warming of the Arctic region. Today, thermal contrast between cold polar air masses (originating over North America, Greenland, and the Arctic) and the relatively warm North Atlantic Ocean results in extemely high ocean-atmosphere heat fluxes in the Norwegian-Greenland Sea and other regions of the North Atlantic, especially along the eastern North American seaboard (Figure 5a). In the RSI run (Figure 5b), the decreased surface air temperature gradient between North America and the North Atlantic (Figure 2c) leads to a significant decrease in these heat fluxes implying that SSTs would increase. In addition, a consideration of the subpolar/temperate North Atlantic region, defined as the area between 31° and 55°N and 15° and 55°W, shows that the average excess of evaporation over precipitation in winter increases from 0.23 to 0.67 mm/d in the RSI case. Such a change suggests that North Atlantic salinities would also increase (by as much as 1‰) if Arctic sea ice were reduced.

Another surface pressure change associated with the reduction in sea ice limits is an increase in the strength of the Azores high (Figure 4c). The general strengthening of this high-pressure system over the eastern North Atlantic, western Europe, and Scandanavia is linked to the northward retreat of the Icelandic low. A stronger winter high-pressure system over eastern North America also characterizes both years of the model run; however, the observed changes in surface pressure are not significantly greater than the level of variation in the control run and thus are not considered statistically significant. Similarly, the stronger low over southwest Asia and the variations in highs and lows over central Asia are not large relative to observed variations in the control run.

In the Arctic region, the decrease in ice cover results in increased ocean-atmosphere heat fluxes, especially along the Arctic margin where the largest winter sea ice changes were imposed (e.g., along the northern Russian coastline or over Baffin Bay; Figures 5a and 5b). Increases of up to 100 W/m^2 are observed in winter. The atmospheric warming over the circum-Arctic region, which results in the reduction of Arctic sea level pressure, is driven by these high heat fluxes from the

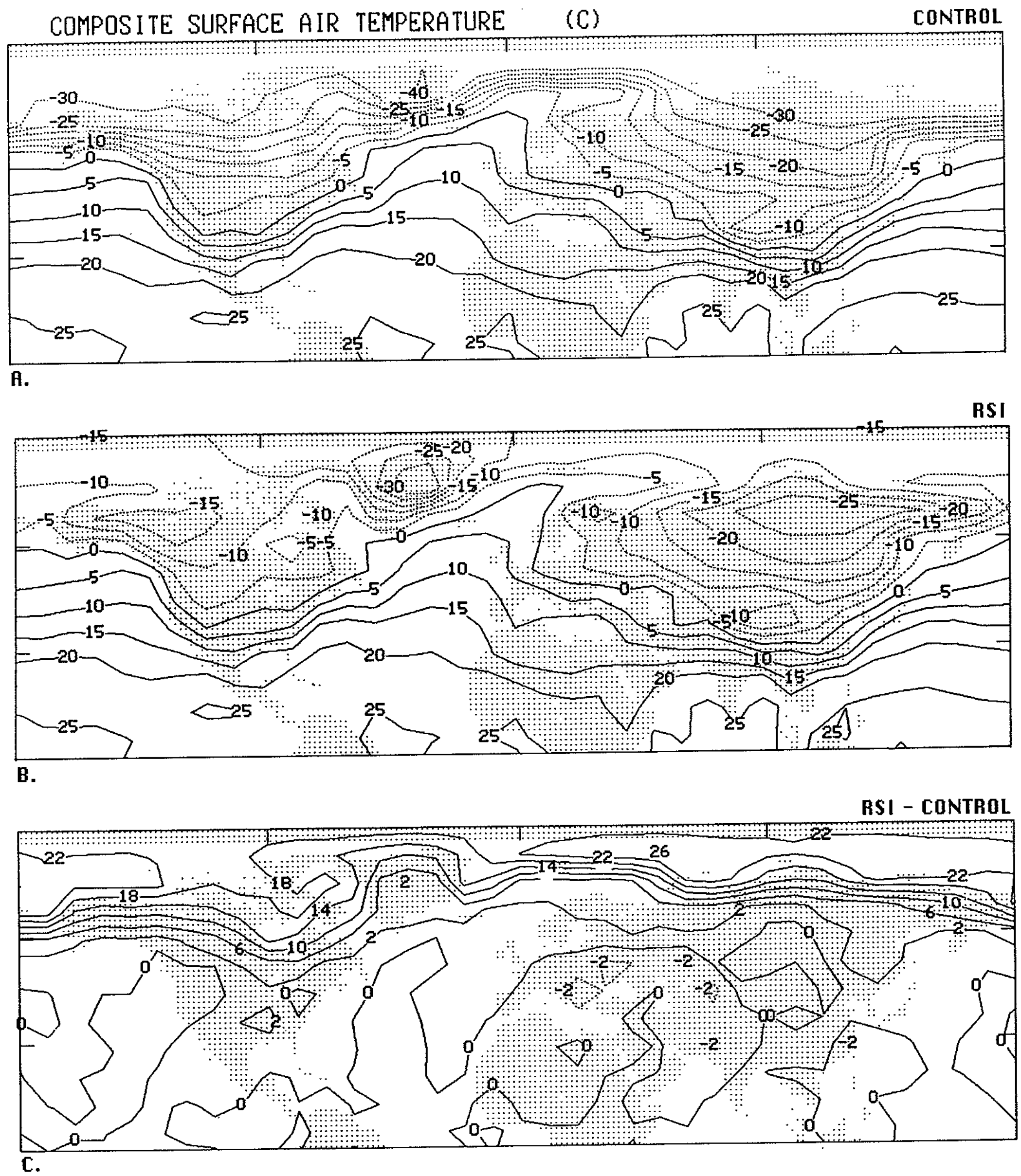

Fig. 2. Winter (DJF) surface air temperature (°C) for (a) control run and (b) RSI run. (c) Difference plot for RSI-control. Negative contours are dashed.

open ocean. On average, an additional flux of 40-80 W/m^2 is observed over the Arctic in areas where sea ice is reduced (Figure 5c).

In this sensitivity test, average planetary albedo changes very little in winter (Table 1) suggesting that oceanic heat losses, from ice-free areas of the Arctic, would not be replenished by increased absorption of solar radiation (e.g., if SSTs were allowed to vary). Ice-albedo feedbacks are relatively unimportant for two reasons: first, the high angle of incidence (and hence high reflectivity) of incoming insolation in polar regions diminishes the influence of surface albedo changes caused by sea ice reductions. This is especially true in winter when incoming solar radiation is at a minimum; second, although surface albedo decreases significantly when sea ice limits are reduced (Table 1), atmospheric albedo increases in the Arctic region due to increases in evaporation and hence cloud cover (which has a high albedo). A 167% increase in evaporation is observed at high northern latitudes (>67°N) in winter of the RSI run (from 0.3 to 0.80 mm/d; Table 1). Keeping in mind the uncertainties associated with clouds in GCMs, this change results in a small increase in the already high amount of cloud cover (from 61 to 65%) which offsets the lower surface albedo of open water in the RSI run. In addition, the increase in evaporation over precipitation would presumably result in higher Arctic Ocean salinities, enhancing thermohaline overturn in this region.

The surface temperature and pressure changes discussed above result in a number of alterations to winter surface wind

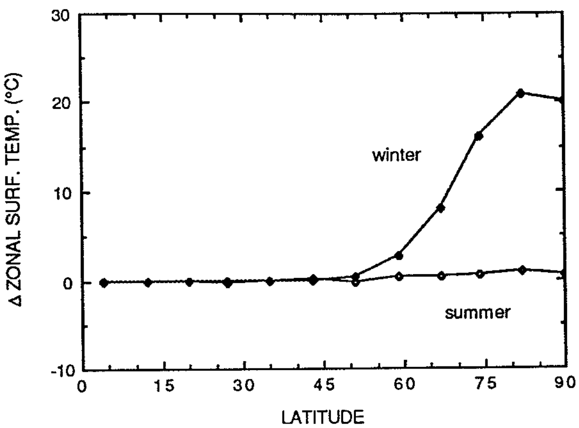

Fig. 3. Difference in zonally averaged surface temperatures for winter (DJF) and summer (JJA) between RSI run and control run.

patterns (Figures 6a-6c). Significant differences between the control and RSI runs include decreased strength of polar easterlies along the circum-Arctic margins, especially off the northern Scandanavian coast, due to a weakening of the polar high pressure system; enhanced cyclonic circulation over the Norwegian-Greenland Sea around a more localized Icelandic low; and stronger anticyclonic circulation centered over Portugal, around the area of the strengthened Azores high. Stronger anticyclonic circulation over the Gulf Stream region (associated with the slightly higher pressures in this region) is also observed when sea ice limits are reduced; however, this change is not significant based on standard deviations for this region. No significant changes are observed in the tropics.

The weakening of the polar easterlies, which today tend to oppose ocean transport from the subpolar North Atlantic to the Arctic, would lead to enhanced exchange between the water masses of these two basins when sea ice limits were reduced. In addition, the strengthened cyclonic circulation over the Norwegian-Greenland Sea would also tend to enhance exchange between the Arctic and subpolar Atlantic. Such an increase in ocean (and heat) transport from the south would help to maintain seasonally ice-free conditions in the Arctic.

Although surface wind speeds increase in the areas of the Icelandic low and Azores high, zonal average wind speed at midlatitudes (e.g., mean strength of the westerlies) is not significantly lower in the RSI run (Table 2) even though the surface air temperature gradient (from 43° to 67°N) decreased from 22°C to 16°C (s.d. 0.5°C) in winter. This can be ascribed to the high stability of the atmosphere in winter which restricts the effects of polar warming to the lowest layers of the atmosphere where its influence on zonal wind speed is damped. For this same reason, both the 200-mbar thermal gradient and the strength of the 200-mbar level (jet stream) winds remain essentially unchanged in winter (Table 2). However, while zonally averaged changes in winds are not large, Figure 6c shows decreased westerlies over northern Canada and Eurasia where the changes in temperature gradient and sea level pressure are largest.

Summer

In June, July, and August, changes in surface air temperature with reduced summer sea ice limits are much less extreme than are observed in winter (Figure 3). A 3°C warming observed over northern Greenland in the RSI run is only just significant (Figures 7a-7c). In the sea level pressure field, no changes are significant given the natural variation of the control run. Thus, relative changes in surface temperature and pressure between the control run and the RSI run are much greater in winter, even though Arctic sea ice cover was significantly reduced in the summer season.

This result can be attributed to the decreased stability of the atmosphere in summer versus winter. Reduced solar heating of polar regions in winter leads to the development of a stably stratified troposphere at polar latitudes with oceanic heat trapped near the surface. In summer, significantly increased radiative heating of the Arctic Ocean results in this region acting as a net heat sink (Figures 8a and 8b), replacing heat lost in the previous winter. When sea ice limits are reduced (and SSTs are held at -1°C), summer heating of the Arctic Ocean increases by 20-40 W/m^2 (s.d. < 10 W/m^2). This additional heating can be ascribed to large reductions in planetary albedo, due primarily to decreased snow and ice cover in the RSI run (Table 3).

Because of the smaller ocean-air temperature contrasts in the North Atlantic in summer, heat fluxes in this region (Figures 8a-8c) are smaller than is observed in winter. The reduction of sea ice limits apparently has little effect on ocean-atmosphere heat exchange in the North Atlantic, with changes never more than 30 W/m^2 (s.d. = 15-20 W/m^2) between the two runs. As a result, evaporation/precipitation ratios and cloudiness, in the temperate/subpolar North Atlantic and at higher latitudes (Table 3), show no significant changes in the RSI run in summer. In addition, no significant changes are observed in surface wind patterns, the hemispheric temperature gradient, or in zonal wind speeds in the summer season (Table 2).

DISCUSSION

Seasonally, the largest changes in surface air temperature, sea level pressure, and wind speed and direction occur in winter. The reduction of sea ice cover in winter results in a significant surface warming of the Arctic region and a reduction in surface pressure. The Icelandic low shifts northward and weakens overall while the Azores high becomes stronger. Major changes in surface wind patterns between the two runs reflect alterations in sea level pressure patterns described above, as well as changes in the winter temperature gradient over land masses (e.g., decreased speed of the polar easterlies over northern Europe and Asia and locally intensified circulation around the Icelandic low and Azores high). However, no large zonal decreases in the strength of the surface westerlies or the 200-mbar jet stream are observed when sea ice is reduced.

The circum-Arctic warming is driven by the large increases in ocean-to-atmosphere heat fluxes which result when sea ice limits are reduced in winter. This warming is particularly pronounced over adjacent areas of North America, where large reductions in ice cover were imposed over Hudson and Baffin bays. The considerable increases observed in winter surface air temperature attest to the sensitivity of the high latitude climate, particularly over northern Canada, to changes in sea ice boundary conditions.

In contrast, no significant warming is observed in summer when sea ice is reduced (Table 3). Today, the Arctic is a net sink of heat in summer; reductions in sea ice enhance this flux slightly with little overall atmospheric effect and no obvious far-field effects. Therefore, because we would not expect summer surface temperatures to decrease over North America, a late Pliocene increase in Arctic sea ice limits

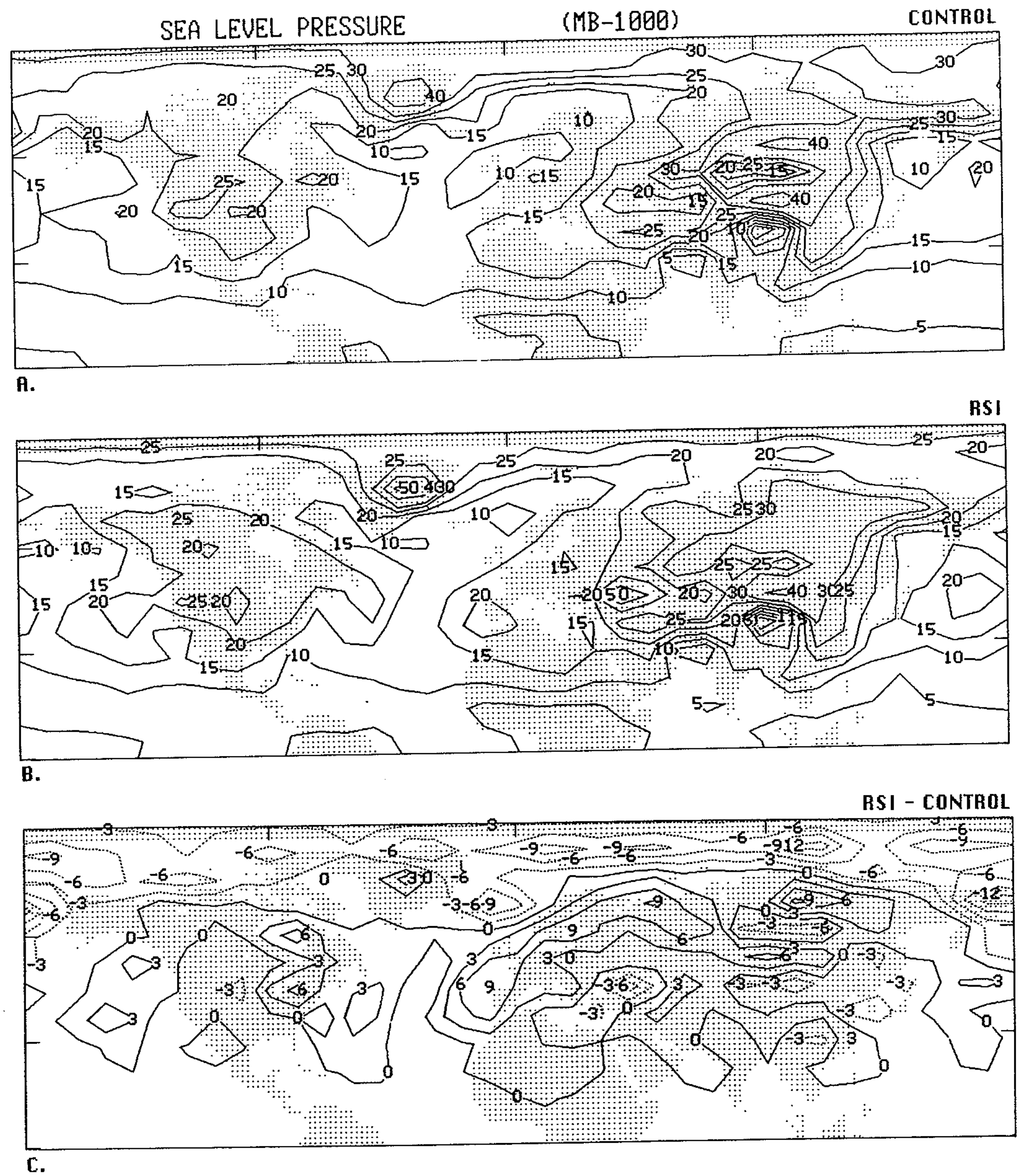

Fig. 4. Winter (DJF) sea level pressure for (a) control run and (b) RSI run. (c) Difference plot for RSI-control. Negative contours are dashed.

would not necessarily lead to enhanced ice growth on land. This argument assumes that summer temperature is the critical factor controlling the mass balance of a continental ice sheet [e.g., Weertman, 1976; Imbrie and Imbrie, 1980].

A map of the model's modern average annual net heating at the surface (Figure 9a) shows that large amounts of heat are today transferred from the ocean to the atmosphere in the Gulf Stream-North Atlantic Current region and in the Norwegian-Greenland seas. (Other places where the ocean supplies heat to the atmosphere, although not to such a great extent as in the North Atlantic, are in the Kuroshio Current, the Brazil Current, and in the Agulhas Current. The ocean acts primarily as a sink for atmospheric heat in the important upwelling cells off western Africa, North America, and South America.) By comparison, a significant decrease in the ocean heat flux is observed in the Gulf Stream and Norwegian-Greenland Sea regions when perennial sea ice cover is removed (Figure 9b). This effect is linked to the strong warming over North America and Greenland and the subsequent decrease in the influence of polar air masses from these regions. Our results show that this warming occurs primarily in winter and suggest that winter sea surface temperatures would increase significantly in the North Atlantic Ocean were sea ice reduced. Over the northwestern Pacific, changes in surface net heating are not as significant as in the Atlantic (Figure 9c), reflecting the more restricted warming over eastern Asia when sea ice is reduced (Figures 2c and 7c).

To maintain a seasonally ice-free Arctic some mechanism

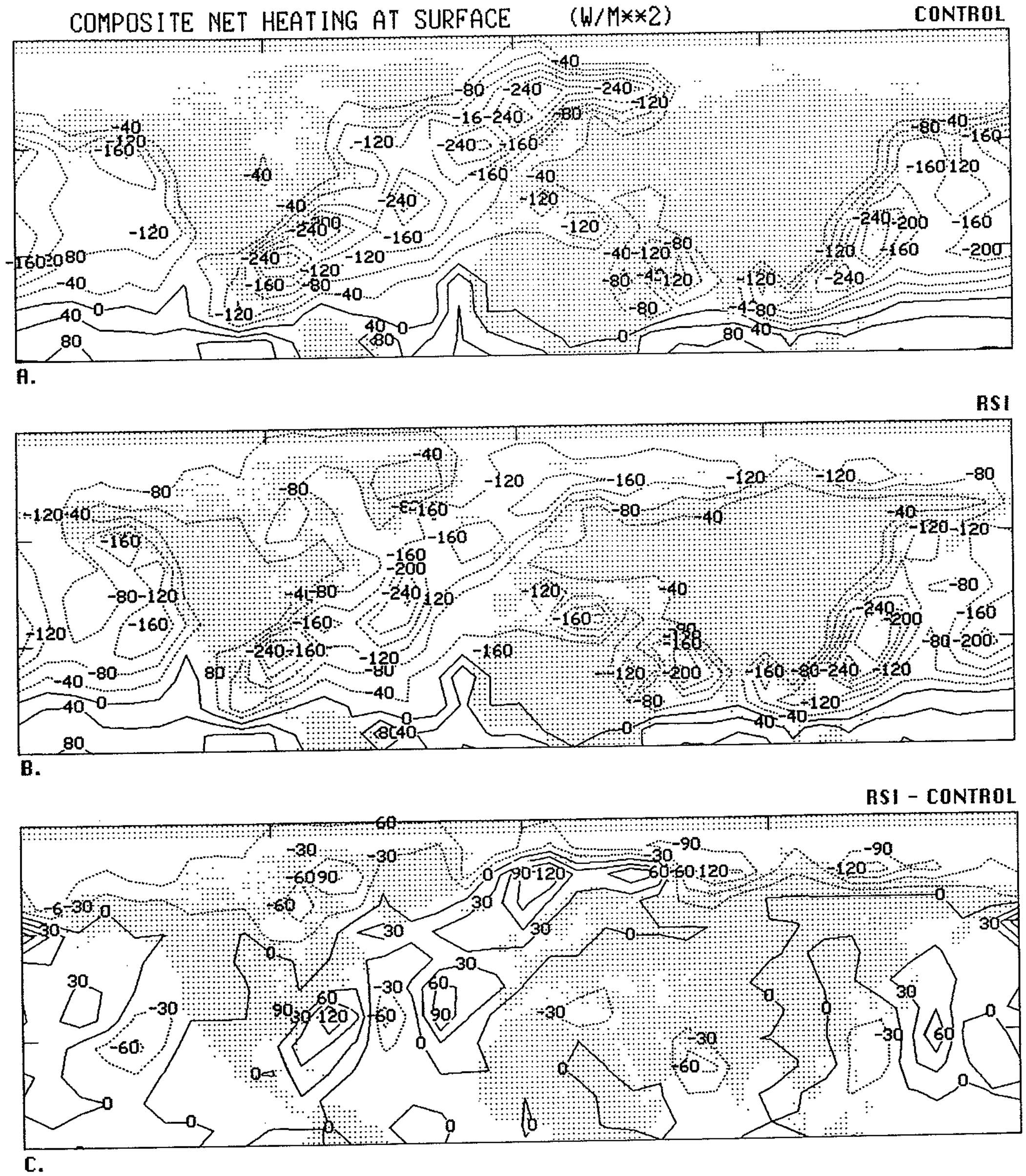

Fig. 5. Winter (DJF) net heating at surface (in W/m²) for (a) control run and (b) RSI run. (c) Difference plot for RSI-control. Negative contours are dashed.

other than direct solar heating in summer is needed. Annual surface heat fluxes (Figures 9b and 9c) show a net deficit of heat in the Arctic when sea ice limits are reduced. In the absence of an independent mechanism to warm the polar regions (e.g., higher atmospheric ρCO_2), the higher ocean-atmosphere heat fluxes observed in the Arctic would have to be balanced by increased oceanic heat transport from mid latitudes. If no source of oceanic heat were available, the Arctic water column would eventually be cooled and sea ice would reform. However, two factors suggested by this experiment would act to enhance ice-free conditions: the lower oceanic heat losses observed in the North Atlantic and Norwegian-Greenland Sea when sea ice is reduced suggest that water entering the Arctic through the Fram Straight would be warmer in the RSI case. In addition, a weaker Arctic high-pressure cell would be more favorable to ocean transport through the Fram Straights via alterations in the surface wind fields. Both these effects would act as positive feedbacks to maintain a seasonally ice-free Arctic.

It will be important to address whether equilibrium ice-free conditions can be maintained in the Arctic using coupled ocean-atmosphere circulation models with interactive SSTs and realistic ocean heat transports. In this way we can get further insight into the changes in climate boundary conditions needed to maintain an ice-free Arctic Ocean. For instance, Semtner's [1987] study using a coupled ocean-ice model suggests that a 2°-4°C global warming (in this case associated with a doubling of atmospheric CO_2) would maintain ice-free

TABLE 1. Average Values for Northern Hemisphere in Winter (December, January, and February)

	Control	RSI†	Δ (s.d.)
Ice cover, %	4.9	1.8	
Surface temperature, °C	8.41	9.87	+1.46 (0.12) *
Snow cover, %	22.6	19.3	-3.3 (0.38) *
Planetary albedo, %			
0-90°N	30.7	30.0	-0.7 (0.28) *
67-90°N	86.7	86.5	-0.2 (0.17)
Surface albedo, %			
0-90°N	11.4	10.7	-0.7 (0.10) *
67-90°N	86.6	83.4	-3.2 (0.39) *
Cloud cover, %			
0-90°N	53.7	53.4	-0.3 (0.70)
67-90°N	61.4	64.7	+3.3 (2.3)
Precipitation, mm/d			
0-90°N	3.23	3.20	-0.03 (0.10)
67-90°N	0.87	1.25	+0.38 (0.08) *
Evaporation, mm/d			
0-90°N	3.27	3.26	-0.01 (0.06)
67-90°N	0.30	0.80	+0.50 (0.05) *

*Significant differences between two runs (>2 s.d.).
†Average of years 1 and 2.

conditions in summer but have little effect on sea ice extent in winter. Ultimately, the presence of relatively warmer flora and fauna around the Arctic margins in the Pliocene and earlier in the Cenozoic (discussed below) suggests that a mechanism must exist to maintain significantly reduced sea ice limits in the Arctic Ocean.

Comparisons to Other Models

With a few exceptions, previous GCM modeling experiments have focused on sea ice as a predicted variable, testing its response to imposed changes such as increased CO_2 [e.g., Manabe and Stouffer, 1980; Manabe and Wetherald, 1980] or increased or decreased solar insolation [Kutzbach and Gallimore, 1988; Hansen et al., 1984]. In these studies, it is difficult to determine which climate responses are due to changes in sea ice and which are due to the initial forcing. The more direct approach, imposing reductions in Arctic sea ice and observing the impact on northern hemisphere climate, has been used by Newson [1973] and Herman and Johnson [1978]. Newson [1973] examined the effects of the complete removal of sea ice in a very brief (one-page) note. Herman and Johnson [1978] examined the predicted climate effects of the modest changes in sea ice extent observed over the last century. By contrast, we examine an intermediate case, that of seasonally present ice cover. Such an experiment is suggested by the results of an earlier climate study [Parkinson and Kellogg, 1979], in which Arctic sea ice reformed in winter even when large temperature increases (up to 9°C) were imposed over the Arctic.

Some comparisons can be made with the results of Newson [1973], who examined the removal of Arctic sea ice in winter using the GCM of Corby et al. [1972]. His experiment, which was run for 80 winter days, showed that high-latitude surface temperatures increased dramatically while mid-latitude continental temperatures decreased slightly. The mid-latitude cooling was ascribed to a southward displacement and weakening of the midlatitude westerlies, due to a decrease in the equator to pole thermal gradient.

By contrast, the RSI experiment, which did not remove sea ice entirely, showed much smaller winter temperature increases over the Arctic: <25°C maximum change reported here, versus a >40°C change reported by Newson [1973]. In addition, because the polar warming is restricted near the surface by the high atmospheric stability, no change in the strength of the upper level westerlies is observed when sea ice limits are reduced (Table 2). Warming, rather than cooling, is observed at mid-latitudes in the RSI experiment, although only over central Canada can these changes be considered statistically significant.

Our results can also be compared to those of Simmonds [1981] and Simmonds and Budd [1990], who examined the effects of reduced sea ice limits around Antarctica. In these experiments winter sea ice cover around Antarctica was reduced considerably (to summer limits and by 50%, respectively). In both cases, significant atmospheric heating of about 6°C was observed above the sea ice anomaly, in qualitative agreement with our results. However, a pronounced weakening of the zonally-averaged upper westerlies was also observed in the southern hemisphere, contrary to the response of the northern hemisphere to reductions in Arctic sea ice cover. The fact that Antarctic sea ice extends to lower latitudes than Arctic sea ice may in part account for the differences in the observed response.

Lastly, the RSI experiment can be compared to that of Hansen et al. [1984] who found, using the same GCM with interactive SSTs, that a 4 W/m^2 increase in global net heating was observed immediately after CO_2 levels in the atmosphere were doubled. This radiative imbalance was ultimately converted to a 4.2°C global temperature increase when the model was run to equilibrium (35 years). That study attributed some of this large climatic effect (~1°C) to the role of ice-albedo feedbacks, both in the Arctic and Antarctic. However, they pointed out that the ice-albedo feedback was non-linear and would contribute only a few tenths of a degree change in the absence of the accompanying moisture and cloud feedbacks. In agreement with this suggestion, our model, which shows no significant changes in hemispheric cloud cover (Tables 1 and 3) or relative humidity, has a 0.1 W/m^2 increase in the annual average heat flux when Arctic sea ice limits are reduced, implying little ultimate change in global temperatures.

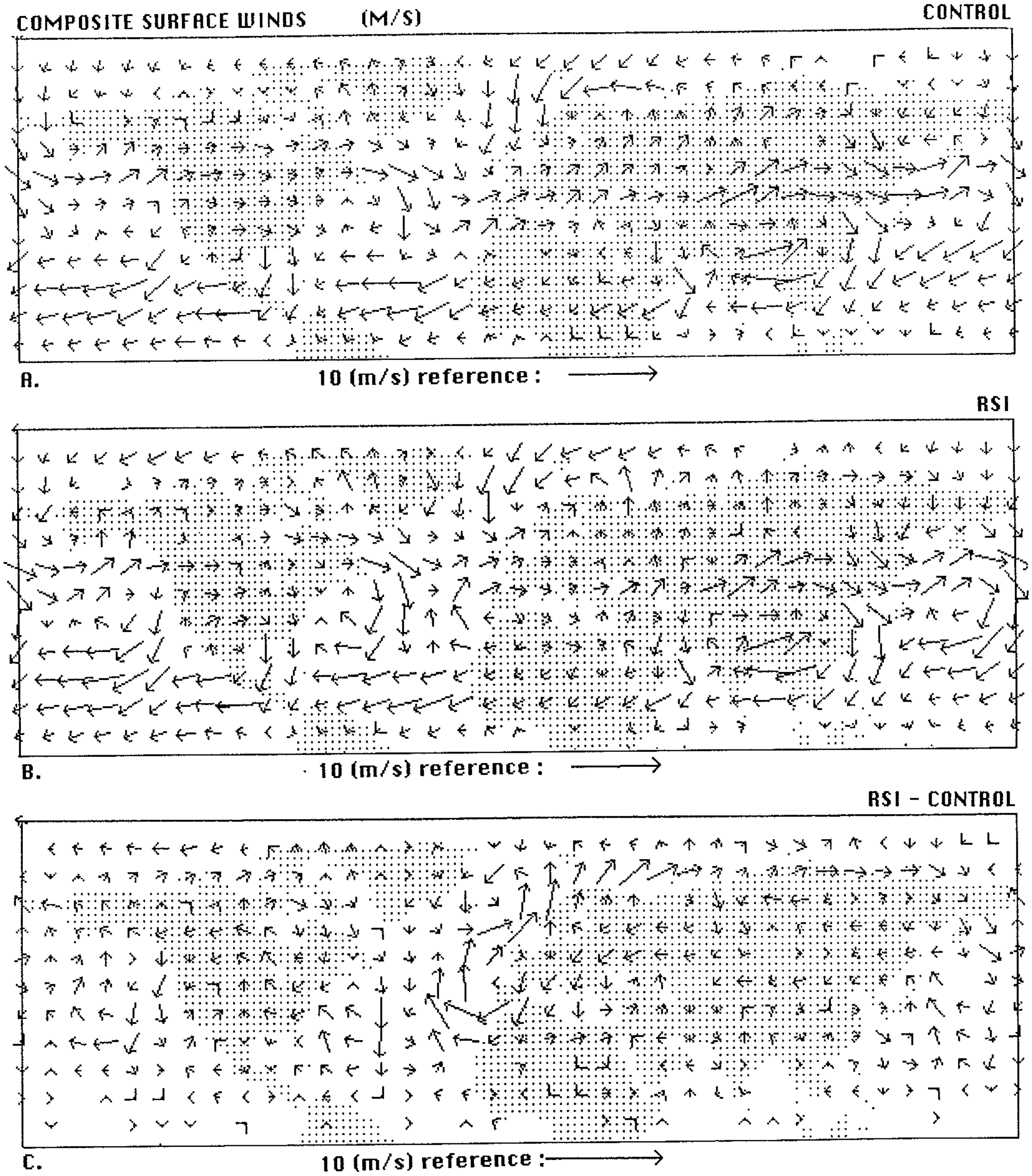

Fig. 6. Winter (DJF) surface wind fields (in m/sec) for (a) control run and (b) RSI run. (c) Difference plot for RSI-control.

TABLE 2. Changes in Zonal Wind Speed Between Control and RSI Run for Latitude Band 47°-63°N

	Control	RSI†	Δ (s.d.)
	Winter		
U-comp. surface winds	1.73	1.34	-0.39 (1.4)
V-comp. surface winds	0.25	0.23	-0.02 (0.20)
U-comp. 200-mbar winds	17.1	16.5	-0.6 (2.3)
V-comp. 200-mbar winds	-0.11	-0.08	0.03 (0.06)
	Summer		
U-comp. surface winds	-0.9	-1.04	-0.14 (0.94)
V-comp. surface winds	-0.12	-0.02	0.10 (0.08)
U-comp. 200-mbar winds	10.4	10.4	0.0 (1.4)
V-comp. 200-mbar winds	-0.01	-0.02	-0.01 (0.04)

Speeds in meters per second.
†Average of years 1 and 2.

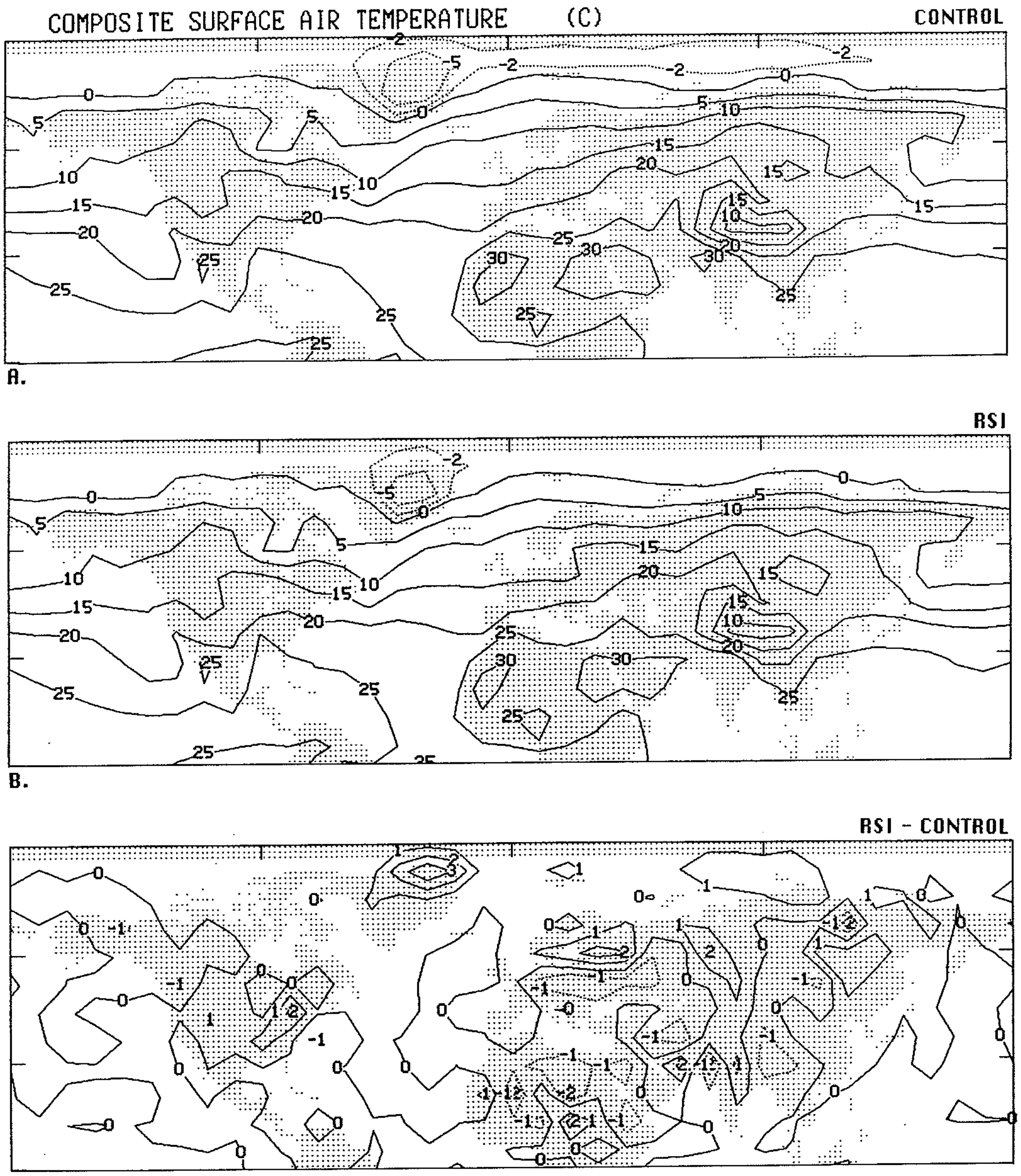

Fig. 7. Summer (JJA) surface air temperature (°C) for (a) control run and (b) RSI run. (c) Difference plot for RSI-control. Negative contours are dashed.

Comparison to Paleoclimatic Data

Our experiment suggests what climatic changes can be expected from a very substantial reduction in Arctic sea ice limits. By comparing to geologic proxies of sea ice, temperature, ocean circulation, and other key elements of the climate system, we may be able to better constrain the history of Arctic sea ice cover in the late Cenozoic.

Data from a number of geologic sections indicate that the circum-Arctic was significantly warmer in the late Pliocene. A pollen-based climate reconstruction from the Kap Kobenhavn formation of Northern Greenland [Funder et al., 1985] suggests summer temperatures 6°C higher than at present, and sedimentologic structures suggest an ice-free coastline in this region. Although the dating of this section is uncertain and could range from 3.1 to 1.6 m.y., Funder et al. [1985] proposed an age of about 2.0 Ma for this site. The results of the RSI experiment do not predict such large increases in summer temperature when sea ice is reduced (i.e., only 1°C); however, by holding sea-surface temperature fixed at -1°C, we have minimized the climatic effects of removing sea ice, particularly along the southernmost margins of the ice cover where summer insolation is highest.

At Ocean Point in northern Alaska, pollen reconstructions from the Gubik Formation also suggest a warmer climate, represented by a mixed conifer forest, in the late Pliocene

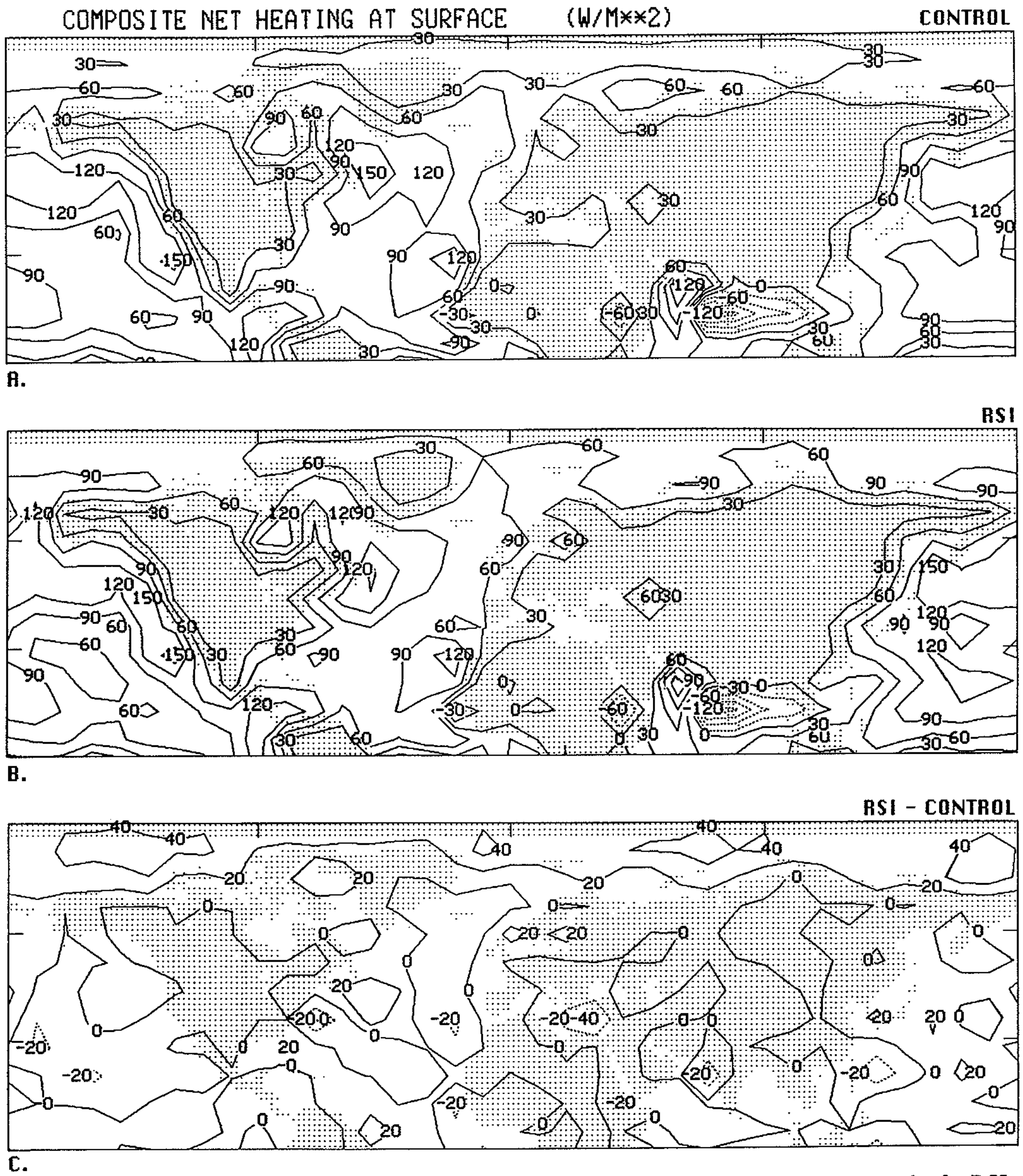

Fig. 8. Summer (JJA) net heating at surface (in W/m^2) for (a) control run and (b) RSI run. (c) Difference plot for RSI-control. Negative contours are dashed.

[Nelson and Carter, 1985]. The absolute age of this section, however, is again not well-constrained and is placed sometime between 3.0 Ma and 2.5 Ma by C.A. Repenning and E.M. Brouwers (manuscript in preparation, 1990) based on faunal remains and correlation to the nearby Fish Creek formation. By analogy with modern pollen assemblages near Anchorage, Nelson and Carter [1985] proposed that mean annual air temperature decreases of approximately 13°C have occurred since the deposition of these beds, with much smaller decreases in mean summer temperature. These model results are consistent with the observation of larger temperature reductions in winter (16°-20°C) versus summer (~1°C), although we show only an 8°C reduction in mean annual temperature at this location using specified SSTs.

In the Fish Creek section, also in northern Alaska, invertebrate (mollusc) and vertebrate (sea otter) evidence, shown to be younger than 2.5 Ma by paleomagnetostratigraphy, indicates that the Arctic margin could not have been frozen for more than one month a year at this location. Similarly, geologic evidence from a number of other sites, summarized by Repenning et al. [1987] and C.A. Repenning and E.M. Brouwers (manuscript in preparation, 1990), reflects a generally warmer circum-Arctic climate in the late Pliocene, with some indications by pollen of alternating warmer and colder intervals. Unfortunately, the ages of most of these site are poorly constrained but are believed to have been deposited prior to the Pliocene-Pleistocene boundary (1.6 Ma).

The large increases in air temperature predicted by the land

TABLE 3. Average Values for Northern Hemisphere in Summer (June, July, and August)

	Control	RSI†	Δ (s.d.)
Ice cover, %	3.6	0.7	
Surface temperature, °C	19.8	19.8	0.0 (0.09)
Snow cover, %	2.2	0.8	-1.4 (0.21) *
Planetary albedo, %			
0-90°N	29.5	28.9	-0.6 (0.27) *
67-90°N	53.6	48.4	-5.2 (1.07) *
Surface albedo, %			
0-90°N	13.4	13.0	-0.4 (0.20)
67-90°N	40.4	20.0	-20.4 (2.02) *
Cloud cover, %			
0-90°N	47.8	48.3	+0.5 (0.4)
67-90°N	63.3	61.7	-1.6 (3.2)
Precipitation, mm/d			
0-90°N	3.33	3.42	+0.09 (0.04) *
67-90°N	1.46	1.51	+0.05 (0.10)
Evaporation, mm/d			
0-90°N	2.97	3.03	+0.06 (0.03)
67-90°N	0.71	0.76	+0.05 (0.06)

*Significant differences between two runs (>2 s.d.).
†Average of years 1 and 2.

data suggest that seasonally ice-free conditions must have existed at least intermittently around the margins of the Arctic between 3.5 and about 2.5 Ma. However, because no land sections have been firmly dated in the critical late Pliocene/early Pleistocene interval after 2.5 Ma, it remains uncertain when the present-day perennial sea ice cover developed in the Arctic. The deep sea record from this ocean is also ambiguous. Herman and Hopkins [1980], using sedimentologic data, and Gilbert and Clark [1982/1983], using dynocyst data, both proposed that modern Arctic sea ice cover formed near the beginning of the Brunhes chron (0.73 Ma). However, the available geologic indices do not give unambiguous proof of the presence or absence of perennial sea ice either before or after this time [Clark and Hanson, 1982; Gilbert and Clark, 1982/1983].

Another possibility is that permanent perennial sea ice developed near the Pliocene/Pleistocene boundary [Scott et al., 1989], perhaps at the beginning of the Eubronian cold period in Europe which falls within the Olduvai subchron (1.66-1.88 Ma) [Repenning et al., 1987]. Two changes in the North Atlantic region, consistent with the alterations in atmospheric circulation suggested by RSI experiment, occur at about this time. The first is the abrupt increase in abundance of the foraminifera *N. pachyderma* (sinistral) in the North Atlantic at 1.66 Ma [Raymo et al., 1986]. An increase in this species, which today is most abundant in polar environments [Kipp, 1976], could be indicative of a significant drop in winter SSTs associated with the development of perennial Arctic sea ice cover. Similarly, the influx of boreal species to the Mediterranean Sea just after the Olduvai [Colalongo and Pasini, 1980; Tauxe et al., 1983] may be related a cooling of the open North Atlantic Ocean. In both cases, however, we cannot rule out evolutionary adaptation to a different environment as an explanation for these changes.

A second paleoceanographic change that occurred about this time was the beginning of pronounced North Atlantic Deep Water (NADW) suppression during glaciations [Raymo et al., 1990]. The decrease in NADW production observed for the last glacial maximum has been ascribed to the dual effects of lowered North Atlantic sea surface temperatures [Boyle and Keigwin, 1987] and increases in sea ice cover [Kellogg, 1980]. GCM modeling results [Manabe and Broccoli, 1985] have shown that continental ice sheets play an important role in cooling the adjacent North Atlantic Ocean during glaciations and the results reported here indicate that variations in Arctic sea ice extents can also significantly impact North Atlantic SSTs and salinities.

Specifically, three changes predicted by the RSI model would promote thermohaline overturn in the the Norwegian-Greenland and Labrador Seas and thus formation of North Atlantic Deep Water: an enhancement of surface water salinities in the North Atlantic region resulting from an increase in evaporation relative to precipitation; a localized strengthening of the Icelandic low over the Norwegian-Greenland Sea; and an increase in the salinity of water leaving the Arctic, driven by increased evaporative fluxes in this region when sea ice limits are reduced. (Recent studies by Aagaard [1988] and Aagaard and Carmack [1989] suggest that the Arctic Ocean may play a vital role in controlling thermohaline overturn in the North Atlantic by influencing the supply of fresh water delivered via the East Greenland Current.) By enhancing glacial cooling of the North Atlantic, decreasing North Atlantic salinities, and increasing delivery of low-salinity Arctic water via the East Greenland Current, development of perennial Arctic sea ice cover in the early Pleistocene may have resulted in the enhanced glacial suppression of NADW observed after 1.5 Ma.

While the above linkages may be speculative, neither the apparent cooling of the North Atlantic and Mediterranean at 1.6 Ma nor the relatively stronger suppression of North Atlantic Deep Water after this time can be ascribed to increases in global ice volume. Benthic oxygen isotope records [Shackleton and Hall, 1989; Raymo et al., 1990] show no significant differences in $\delta^{18}O$ signal amplitude for 0.5 m.y. before or after the Plio-Pleistocene boundary. The development of modern perennial Arctic sea ice cover, which would not affect oceanic $\delta^{18}O$ values nor influence the size of continental ice sheets (for reasons discussed earlier), may have caused, or at least reinforced, these changes.

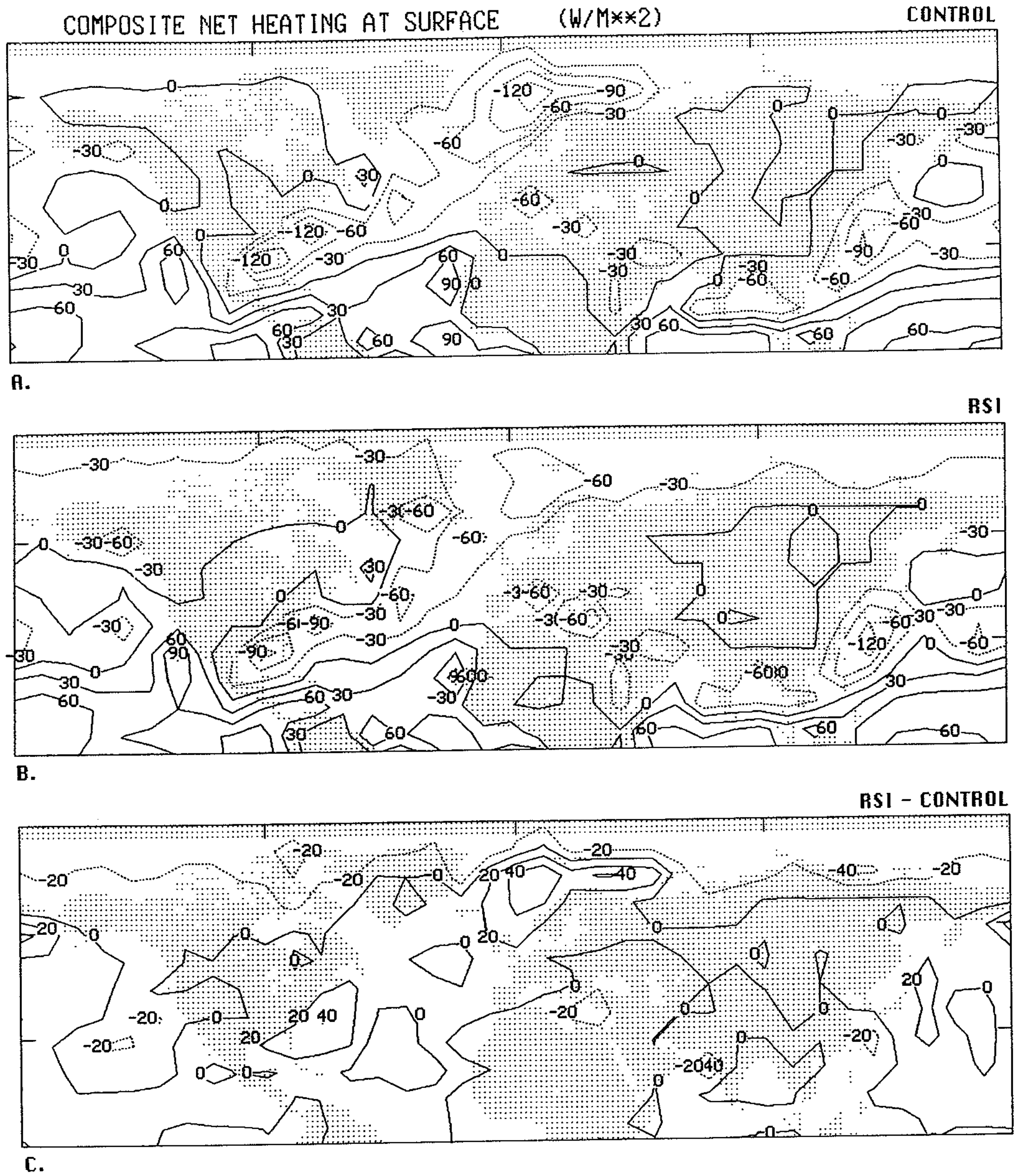

Fig. 9. Annual average net heating at surface (in W/m^2) for (a) control run and (b) RSI run. (c) Difference plot for RSI-control. Negative contours are dashed.

SUMMARY

Several studies [e.g., Clark, 1982a] have taken the view that gradual late Cenozoic cooling of the northern hemisphere eventually reached thresholds where first seasonal, and then perennial sea ice formed in the Arctic Ocean. In this view, a relatively moderate climatic cooling (5°-10°C) over the span of several million (or tens of millions) of years would have led to a relatively sudden cooling of the Arctic (15°-20°C or more) when sea ice formed, which in turn would have had large impacts on northern hemisphere climate [Clark, 1982b].

The results of our experiment are not entirely consistent with this hypothesis. Removing sea ice results in minor temperature changes in summer, with larger changes of 15°-20°C observed in winter. However, these changes are for the most part confined close to regions from which sea ice has been removed, primarily the Eurasian margins of the Arctic Ocean and the Canadian archipelago. This suggests that the late Cenozoic appearance (or increase) of Arctic sea ice may have had a relatively modest impact on northern hemisphere climate, except in regions proximal to the Arctic and in the North Atlantic Ocean. This conclusion is only valid within the boundary constraints of the cold SST values (-1°C) imposed in ice-free regions.

The large winter temperature anomaly observed over northern Canada, and the additional thermal effects advected farther to the southeast, result in large part from the removal of sea ice from the continental shelf of the Canadian archipelago. This archipelago may be a relatively young feature geologically. Hopkins [1967] suggested that flooding of the Canadian archipelago first occurred during the Pliocene, perhaps during the middle Pliocene spread of Pacific mollusc species into the Atlantic between 3.5 and 3.1 Ma. It is not known whether or not there was a time when the archipelago was submerged but ice-free in winter. If so, icing over of the archipelago would have produced a major cooling over North America in winter. Thus, tectonic submergence of this region could have been an important event in the evolution of climate in the interior of North America, leading to an amelioration of regional climates if it was initially ice-free and then a cooling as modern ice cover developed.

Many uncertainties concerning Arctic sea ice remain. Geologic estimates of sea ice limits and Arctic temperatures (summarized by Repenning et al. [1987]) casts doubt on the possibility that perennial sea ice, as extensive as today, could have existed prior to 2.5 Ma. However, when modern sea ice cover formed in the Arctic remains unresolved; did it form gradually or develop in steps, and how did ice cover respond to glacial-interglacial variations in climate? The lack of circum-Arctic geologic data from the early Pleistocene and the added influence of large continental ice sheets after 2.4 Ma hinder our ability to determine exactly when perennial sea ice cover formed in the Arctic Ocean.

In addition, we have ignored the ultimate question of why a major increase in Arctic sea ice cover would occur. The strong salinity stratification of the present Arctic Ocean suppresses deep convection and turbulent diffusion with the underlying waters, thus reducing the upward transfer of heat which could melt sea ice [Aagaard and Coachman, 1975; Aagaard et al., 1981]. The strong pycnocline is maintained primarily by influxes of low salinity water from the discharge of major Siberian rivers such as the Ob and Yenisei and by influxes of low-salinity Pacific water through the Bering Straits.

For freshwater input from Siberian rivers to have increased over the last 2 m.y., evaporation-precipitation ratios in the watershed areas would have had to decrease. However, vegetation histories from Siberia [Wolfe, 1985] suggest the opposite trend, with a change from deciduous forest to conifers to Taiga since the late Miocene. It is unlikely that this inferred decline in temperatures was accompanied by enhanced rainfall. Other geologic evidence for late Neogene aridification of northern and central Asia, ascribed to the climatic effects of Tibetan uplift, is summarized by Liu [1983], Whyte [1984], and Li [1985].

Furthermore, the subsidence of the Bering land bridge, and the initiation of flow through the Bering Straits, occurred early in the late Pliocene. A major dispersal of Pacific fauna into the Atlantic Ocean suggests that the Bering seaway was permanently established by about 3.0 Ma [Hopkins, 1967; Gladenkov, 1981]. Similarly, the formation of the Panamanian isthmus could also have significantly influenced oceanic heat transport to the high-latitude North Atlantic and potentially the Arctic Ocean [Maier-Reimer et al., this issue] but this occurred in the mid-Pliocene [Keigwin, 1982]. Thus, while long-term changes in either Siberian river discharge or ocean circulation pathways could be expected to influence the stability of the Arctic water column, either the direction or timing of the inferred variations are inadequate to explain the development of perennial ice cover after 2.5 Ma.

Some other factor or factors, such as global cooling associated with the glaciation of North America and Eurasia or tectonic movements of the Earth's crust, must have influenced the evolution of Arctic sea ice cover over the last 2.5 m.y. For instance, results of a GCM experiment which examined the effects of late Cenozoic uplift of the Tibetan and Colorado plateaus [Ruddiman and Kutzbach, 1989], predict a cooling of the Arctic region, both in winter and summer, with progressive uplift. Long-term atmospheric CO_2 changes, also driven by variations in global uplift rates [Raymo et al., 1988], could have initiated cooling in polar regions. However, it is unclear whether tectonic movements could have had a significant climatic effect over an interval as short as the last 2 m.y.

Additional geologic proxies, especially from the circum-Arctic regions and Arctic basin, are needed if we are to understand fully the climatic and paleoceanographic history of Arctic sea ice. Sensitivity experiments with interactive ocean-ice-atmosphere models will also provide more realistic information on the response of sea ice to changes in boundary conditions. In particular, the important change in North Atlantic air-sea heat exchange needs to be examined. Finally, experiments with interactive ocean/ice models are needed to evaluate factors (uplift, CO_2) that caused sea ice changes.

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