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Last Interglacial Arctic warmth confirms polar amplification of climate change

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Abstract

The warmest millennia of at least the past 250,000 years occurred during the Last Interglaciation, when global ice volumes were similar to or smaller than today and systematic variations in Earth's orbital parameters aligned to produce a strong positive summer insolation anomaly throughout the Northern Hemisphere. The average insolation during the key summer months (M, J, J) was ca 11% above present across the Northern Hemisphere between 130,000 and 127,000 years ago, with a slightly greater anomaly, 13%, over the Arctic. Greater summer insolation, early penultimate deglaciation, and intensification of the North Atlantic Drift, combined to reduce Arctic Ocean sea ice, allow expansion of boreal forest to the Arctic Ocean shore across vast regions, reduce permafrost, and melt almost all glaciers in the Northern Hemisphere. Insolation, amplified by key boundary condition feedbacks, collectively produced Last Interglacial summer temperature anomalies 4–5 °C above present over most Arctic lands, significantly above the average Northern Hemisphere anomaly. The Last Interglaciation demonstrates the strength of positive feedbacks on Arctic warming and provides a potentially conservative analogue for anticipated future greenhouse warming.

1. Introduction

Planetary warm times in the recent geological past inform the debate over Earth's response to the continuing build up of radiatively active atmospheric trace gases. Strong positive feedbacks in the Arctic are expected to amplify future greenhouse warming (Holland and Bitz, 2003), but the cumulative effect of these feedbacks remains debated (Serreze and Francis, 2006). Past planetary warm times provide a testing ground for the debate over polar amplification of climate change. During the Last Interglaciation (LIG) Earth was warmer than at any time in at least the past 250,000 years, with global temperatures 0–2 °C above present (e.g., CLIMAP Project Members, 1984; Bauch and Erlenkeuser, 2003). Circum-Arctic warming is recognized to have been substantial during the LIG (Lauritzen and Anderson, 1995), but quantitative reconstructions for this region are lacking.

Here we present quantitative estimates of circum-Arctic LIG summer air and sea-surface temperatures reconstructed from proxy records preserved in terrestrial and marine archives. These reconstructions demonstrate that Arctic summer air temperatures averaged ca 4–5 °C above present for most of the Arctic, well above the planetary LIG average, although with coherent spatial patterns in the magnitude of warmth. Arctic summers were warm enough to melt all glaciers below 5 km elevation except the Greenland Ice Sheet, which was reduced by ca 20–50% (Cuffey and Marshall, 2000; Otto-Bliesner et al., 2006). In addition, the margins of permanent Arctic Ocean sea ice retracted well into the Arctic Ocean basin and boreal forests advanced to the Arctic Ocean coast across vast regions of the Arctic currently occupied by tundra, although the central Arctic Ocean basin remained covered by permanent sea ice (Spielhagen et al., 2004). These boundary condition changes reduced Earth's albedo and altered the exchange of heat, moisture and trace gases between the land, ocean and atmosphere, collectively amplifying insolation-driven (Fig. 1) circum-Arctic interglacial warmth.

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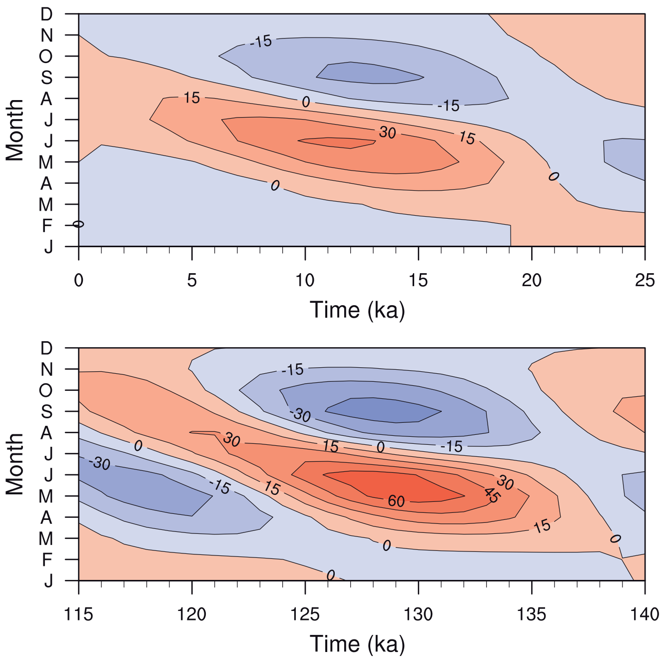

Fig. 1. Monthly insolation anomalies (deviations from the present) for the time periods from 25 ka to present (upper panel) and from 140 to 115 ka (lower panel) at 65°N, showing the larger insolation anomaly for the Last Interglaciation relative to the present interglaciation. Anomalies are expressed as W m^{-2}; time is in thousands of years before present.

2. The Last Interglaciation

The LIG was recognized more than a century ago in northern Europe (Harting, 1875). Subsequently, it was defined from pollen records as the Eemian, a time of deciduous forests bounded by periods of tundra (Jessen and Milthers, 1928). A half-century later, the Eemian was correlated with marine isotope stage (MIS) 5 (Emiliani, 1955), and eventually with the much shorter MIS 5e (Shackleton, 1969; Mangerud et al., 1979; Shackleton et al., 2002). The LIG was securely placed in an absolute time frame when correlated to coral terraces on stable platforms that were amenable to high-precision uranium-series dating (Broecker and Van Donk, 1970; Gallup et al., 2002). We define the LIG temporally as the penultimate interval of minimum global ice volume, when sea level was at or above present. It is now widely accepted that sea level reached modern levels 130$\pm$2 ka (thousands of years ago), and remained at or above this level until the inception of the last glaciation, as ice grew on the continents ca 116 ka (Stirling et al., 1998; Henderson and Slowey, 2000; Mcculloch and Esat, 2000; Gallup et al., 2002; Muhs et al., 2002). Although we use the term LIG as broadly correlative with both the Eemian and MIS 5e, we recognize that these intervals are not precisely correlative. The first half of the LIG coincides with an unusually strong Northern Hemisphere positive summer insolation anomaly, whereas summer insolation receipts are relatively low through the later half of the marine highstand (Fig. 1). Peak Eemian warmth may have occurred slightly before sea level reached present, and certainly was over long before sea level fell below present ca 116 ka (Zagwijn, 1996; Grøsfjeld et al., 1996). Despite these modest temporal differences, sea level provides a powerful tool for correlating the LIG across much of the globe, and associated coral reefs offer the most secure dating of the LIG.

The LIG has received less attention in the Arctic than at mid-latitudes, largely because of difficult access and paucity of sites. A decade ago, a circum-Arctic summary confirmed that the LIG was warmer than present (Lauritzen and Anderson, 1995), although quantitative estimates of warmth were not then possible. In the intervening decade, many new sites have been described from both marine and terrestrial archives, systematic recovery of sea-floor sediment from the Arctic Ocean has provided new evidence for the state of the polar ocean during the LIG, and new training sets, coupled to improved statistical tools, allow more precise quantification of many climate proxies (ter Braak and Juggins, 1993; Birks, 1998). For many Arctic regions, reliable transfer functions for summer temperatures are now available for pollen, chironomids, diatoms, dinoflagellates, planktonic and benthonic foraminifera, alkenones, and several ice-core parameters. Consequently, it is now possible to describe quantitatively at least some aspects of the LIG climate in the Arctic.

3. Methods

3.1. Dating the Last Interglaciation

We capitalize on the uniqueness of the Late Quaternary climate record and recent advances in geochronology to define LIG sites with reasonable certainty. Arctic sites with continuous accumulation since the LIG demonstrate that only the LIG shows evidence of summer temperatures comparable to, or higher than the Holocene. Materials suitable for high-precision U/Th analyses are rare in the Arctic, but advances in trapped-charge dating (optically stimulated luminescence, infra-red-stimulated luminescence, thermoluminescence) provide independent verification of LIG ages for many sites, although the dates often lack sufficient precision to subdivide the interglaciation (Berger and Anderson, 2000; Murray and Funder, 2003). Diagnostic tephra, especially in Alaska and the Nordic Seas, provide additional temporal constraint (Haflidason et al., 2000; Begét and Keskinen, 2003; Rasmussen et al., 2003a, b). LIG sites (Tables 1 and 2) are identified with moderate to high certainty by a combination of stratigraphic position, climate proxies, absolute dating, and relationship to stratigraphic marker horizons (e.g. diagnostic tephra).

3.2. Quantifying Last Interglacial summer temperatures

At high northern latitudes, summer temperatures exert the dominant control on glacier mass balance, unless

Table 1
Published sites with quantitative summer temperature estimates for peak LIG warmth from terrestrial sites around the circum-Arctic. Details of site location and reconstruction techniques are provided in the references cited. Some key cites with important qualitative LIG summer or winter temperature or precipitation are also included

	Proxy	ΔT (°C) Summer	ΔT (°C) Winter	Ppt	Reference	Comments
Scandinavia						
Finland	Treeline position reconstructed from pollen	Warmer			(Saarnisto et al., 1999)	Boreal forest north of present limit; birch at Arctic Ocean shore.
Russia						
European Russia	Pollen, plant macrofossils		4		(Devyatova, 1982) (Grichuk, 1984)	Boreal forest north of present limit.
West central Siberia	Pollen, plant macrofossils	6 to 8	5 to 7	Wetter	(Gudina et al., 1983) (Grichuk, 1984)	Boreal forest 800 km north of present limit.
NE Siberia	Pollen	4 to 8	4	Wetter	(Lozhkin and Anderson, 1995)	Treeline (boreal forest) ~600 km farther north and west of modern range limits. Range extensions up to 1000 km for primary tree species.
NE Siberia Lake Elgygytgyn	Pollen	⩾3	⩾4	Wetter	(Lozkhin et al., 2006)	Northern extension of treeline by at least 150 km.
Siberia	Pollen	4 to 5		Wetter	(Andreev et al., 2004)	Some plant taxa suggest even warmer summers.
Alaska						
NW Alaska Squirrel Lake, SW Brooks Range	Pollen, plant macrofossils	1 to 2	−1 to −3	Wetter	(Berger and Anderson, 2000) (Anderson, 2005)	Continuous high-resolution pollen record. Currently near treeline. Black spruce boreal forest at LIG, analogous to fully developed modern forests located ~300 km to the southeast. Unpublished pollen analogue-derived temperature estimates.
NW Alaska Ahaliorak Lake, North Slope	Pollen	1 to 2	−1 to −2	Wetter	(Brubaker et al., 1996) (Anderson and Brubaker, 1986) (Brubaker, 2005)	Continuous record. Currently in tundra of Brooks Range. In LIG forest extended north of Brooks Range (at least 50 km). Unpublished pollen-Panalogue-derived temperature estimates.
NW Alaska Imuruk Lake, Seward Pen.	Pollen	1 to 2	−1 to –2		(Shackleton, 1982) (Colinvaux, 1964)	Continuous low-resolution record. Old Crow tephra. Tundra today, boreal forest in LIG; westward expansion beyond present limits.
NW Alaska Noatak Valley	Pollen, beetles, plant macrofossils	0 to 2			(Edwards et al., 2003)	Tundra today. OC tephra reworked into LIG sediment. Treeline in Brooks Range of north present limit in LIG

Table 1 (*continued*)

	Proxy	ΔT (°C) Summer	ΔT (°C) Winter	Ppt	Reference	Comments
Interior Alaska Eva Creek	Pollen. spores, soils	0 to 2		Wetter	(Muhs, et al. 2001)	
Interior Alaska KY-11, Koyukuk River, Southern Brooks Range	Pollen, tephra	0 to −2	0 to 1		(Hamilton and Brigham-Grette, 1991) (Schweger, 2002)	Contains Old Crow tephra; partial record of LIG. Unpublished pollen analog-derived temperature estimates.
Interior Alaska Palisades of the Yukon	Pollen, spruce needles, tephra				(Begét, et al., 1991)	Contains Old Crow tephra
Western Alaska Nome	Pollen,	Warmer			(Brigham-Grette and Hopkins, 1995)	Pollen in Pelukian (LIG) marine deposits indicates boreal forest beyond Holocene limits
NW Alasaka Deering	Spruce macrofossils,	Warmer			(Brigham-Grette and Hopkins, 1995)	Macrofossils in Pelukian (LIG) marine deposits indicate spruce was beyond Holocene spruce limit.
NW Alaska Baldwin Peninsula	Spruce macrofossils	Warmer			(Brigham-Grette and Hopkins, 1995)	Macrofossils in Pelukian (LIG) marine deposits indicate spruce was beyond Holocene spruce limit.
Interior Alaska Birch Creek	Pollen	0 to −2	0 to 3	Wetter	(Edwards and McDowell, 1991) (Edwards, 2005)	Lake sediments; contains Old Crow Tephra. Unpublished pollen analog-derived temperature estimates.
Canada						
Tuktoyaktuk, NW Canada	Plant macrofossils	2			(Rampton, 1988)	Boreal forest at coast
Old Crow Basin, northern Yukon	plant macrofossils, beetles	Warmer			(Matthews et al., 1990)	Extra-limital taxa currently restricted to southern Yukon and northern British Columbia
Hudson Bay	Plant macrofossils	Warmer			(Dredge et al., 1990)	Boreal forest farther north
Banks Island, Arctic Canada	Pollen, plant macrofossils, insects	Warmer			(Matthews et al., 1986)	
Ellesmere Is	Pollen	Warmer			(Evans and Mott, 1993)	Possible LIG; organics between till
Robinson Lake, SE Baffin Island.	Pollen	5			(Miller et al., 1999) (Fréchette, 2005)	Stratified LIG lacustrine sediment below till. Correspondence analysis regression; Best modern analogue
Amarok Lake, S. Cumberland Peninsula, Baffin Is.	Pollen	5 to 6			(Fréchette et al., 2006)	Stratified LIG lacustrine sediment Correspondence analysis regression; Best modern analogue
Fog Lake, N. Cumberland Peninsula, Baffin Is.	Pollen	3 to 4			(Fréchette et al., 2006)	Stratified LIG lacustrine sediment. Correspondence analysis regression; Best modern analogue
Fog Lake, N. Cumberland Peninsula, Baffin Is.	Chironomids	7±2			(Francis et al., 2006)	Stratified LIG lacustrine sediment. Weighted averaging

Table 1 (*continued*)

	Proxy	ΔT (°C) Summer	ΔT (°C) Winter	Ppt	Reference	Comments
Brother of Fog Lake N. Cumberland Pen Baffin Is.	Pollen	4			(Fréchette et al., 2006)	Stratified LIG lacustrine sediment. Correspondence analysis regression; Best modern analogue
Brother of Fog Lake N. Cumberland Pen Baffin Is.	Chironomids	8±2			(Francis et al., 2006)	Stratified LIG lacustrine sediment. Weighted averaging
Flitaway Beds, Isortoq Beds, Central Baffin Is.	Insects, plant remains	4 to 5			(Morgan et al., 1993)	Originally thought to be LIG, then assigned to Plio-Pleistocene because estimated temperatures so warm; now likely to be LIG
Greenland						
NGRIP, Central Greenland	$\delta^{18}O$, δD	5			(North Greenland Ice Core Project members, 2004)	Ice core, Greenland Ice Sheet
Renland, E Greenland	$\delta^{18}O$, δD	5			(Johnsen et al., 2001)	Ice core, Renland Ice Cap
Jameson Land, East Greenland	Pollen, plant macrofossils, beetles, other invertebrates	5			(Bennike and Böcher, 1994) (Funder et al., 1998) (Bennike and Weidick, 2001)	Birch woodland present; none today.
Thule, NW Greenland	Pollen, chironomids	4			(Kelly et al., 1999)	
Labrador Sea	Pollen, spores				(Hillaire-Marcel et al., 2001)	ODP site 646: Pollen in LIG Lab Sea sediments indicate that southern Greenland must have been free of inland ice

accompanied by dramatic precipitation changes (Koerner, 2006). Summer temperature is also the most effective predictor for most biological processes, although seasonality and moisture availability may influence phenomena such as evergreen vs. deciduous biotic dominance (Kaplan et al., 2003). For these reasons, we focus our reconstructions on peak summer warmth during the LIG, a strategy adopted for other time periods in the Arctic (e.g the Holocene (Kaufman et al., 2004), and the past 400 years (Overpeck et al., 1997)). Terrestrial climate is reconstructed from diagnostic assemblages of biotic proxies preserved in lacustrine, peat, alluvial, and marine archives and isotopic changes preserved in ice cores and marine and lacustrine carbonates. Quantitative reconstructions of climatic departures from the present-day are derived from range extensions of individual taxa, mutual climatic range estimations based on groups of taxa, and analogue techniques. Estimated winter temperatures, and hence seasonality are well constrained for Europe, but poor for most sectors; likewise, precipitation estimates are limited to qualitative estimates in most cases, and are not available for most regions.

4. Regional summaries (Tables 1 and 2)

4.1. NW Europe

Although well outside of the Arctic, we include the LIG in NW Europe because it has the most detailed terrestrial record of the LIG (see Kaspar et al., 2005 for a review of sites), where a characteristic and remarkably uniform vegetational succession characterizes most of the region (Zagwijn, 1996; Guiter et al., 2003; Kühl and Litt, 2003). Annually laminated lacustrine sediments in Germany suggest that the LIG persisted for 10,000–12,000 years (Müller, 1974; Frenzel and Bludau, 1987; Hahne et al., 1994; Caspers, 1997). Mean July temperatures were 2–3 °C above present in the UK, and about 2 °C above present on the continent, with considerable spatial variability (Aalbersberg and Litt, 1998; Kaspar et al., 2005) . The widespread occurrence of *Hippopotamus* and the water tortoise *Emys orbicularis* in Britain supports these reconstructions (Turner, 2000). Boreal spruce forest spread north of its Holocene limit, and birch forest likely reached to the shores of the Arctic Ocean in Finish Lapland

Table 2
Published sites with quantitative summer sea surface temperature estimates for peak LIG warmth from marine sites around the circum-Arctic. Details of site location and reconstruction techniques are provided in the references cited. Some key cites with important qualitative summer SST information are also included

Site	Proxy	ΔT (°C) Smr SST	ΔT (°C) Wntr SST	Ppt	Reference	Temp. reconstruction technique	Comments
Svalbard	Mollusks, foraminifera	2 to 2.5			(Mangerud et al., 1998) (Miller et al., 1989) (Sejrup et al., 2004)	Faunal range extensions	
North Atlantic	Foraminifera				(McManus et al., 2002)		
JPC8: 61°N	Planktic forams	3 to 4			(Cortijo et al., 1999)	Modern analog	
NA87-25 52°N	Planktic forams	1 to 2			(Cortijo et al., 1999)	Modern analog	
CH69-K9 41°N	Planktic forams	-1			(Cortijo et al., 1999)	Modern analog	
SU90-03 40°N	Planktic forams	0 ± 1			(Cortijo et al., 1999)	Modern analog	
M23414 53.5°N	Planktic forams; Mg/Ca	up to 2			(Kandiano et al., 2004)	Mg/Ca & Faunal	
European Russia	Mollusk, foraminifera	3 to 4			(Funder et al., 2002) (Grøsfjeld, et al., 1996, and references in Svendsen et al., 2004)		LIG “Boreal Transgression” brought Atlantic water along Russian coast from White Sea to eastern Taimyr.
West and central Siberia	Mollusk, foraminifera,		4 to 8		(Troitsky, 1964)		
North coast Alaska	Mollusks in beaches, isotopes in mollusks, ostracodes in off shore marine sediments; Beach morphology	3			(Funder et al., 2002) (Brigham-Grette and Hopkins, 1995)		Sea ice reduction by 800 km; Atlantic water at much shallower depths (30 to 50 m) than in the Holocene (>200 m).
					(Brigham-Grette et al., 2001) (Khim et al., 2001)		
Bering Sea	Mollusks, beach morphology				(Brigham-Grette and Hopkins, 1995)		Sea ice did not occur South of Kotzebue
Arctic Ocean Gateways/Marginal Seas PS2138: 81°N PS2741: 81°N PS2471: 79°N PS2757: 81°N	Coccoliths, dinocysts, benthic forams				(Matthiessen et al., 2001a, b) (Matthiessen and Knies, 2001) (Svendsen et al., 2004)		High influx of Atlantic water through the Nordic Seas into the Arctic Ocean; At least as warm or warmer than present. Boreal mollusks along Russian Arctic coast to Taimyr.
Nordic Seas	Planktic foraminifera				(Sejrup et al., 1995)		
North Nordic Seas	Planktic forams	<0			(Bauch et al., 1999)		
Denmark Shelf	Benthic forams	1.5 to 2			(Seidenkrantz, et al., 2000) (Rasmussen et al., 2003b)		
SW Norway	Benthic forams	1 to 2 greater			(Sejrup et al., 2004)	WA-PLS	
Heat flux to Nordic Seas					(Fronval et al., 1998)		
Central Arctic Ocean	Foraminifera, nanofossils				(Spielhagen et al., 2004)		Abundant planktic fossils in Lomonosov

Table 2 (*continued*)

Site	Proxy	ΔT (°C) Smr SST	ΔT (°C) Wntr SST	Ppt	Reference	Temp. reconstruction technique	Comments
					(Jakobsson et al., 2003) (Svendsen et al., 2004)		Ridge cores imply seasonally open waters or common leads in the eastern and central Arctic Ocean.
Labrador Sea	Foramifera				(Rasmussen et al., 2003a)		Continuous ventilation through LIG
Labrador Sea	Dinocysts, foraminifera				(Hillaire-Marcel et al., 2001)		No ventilation in LIG
Thule, West Greenland	Mollusks, foraminifera	Warmer			(Funder, 1990) (Kelly et al., 1999)		Shallow-water raised marine sediment.
Jameson Land, East Greenland	Mollusks, foraminifera	2 to 3			(Funder et al., 1998)		Vigorous advection of Atlantic water into coastal regions

(Saarnisto et al., 1999). LIG January anomalies were small over most of Europe, but increased to the east and north, where temperatures 4–7 °C higher than present are reconstructed for the LIG (Kaspar et al., 2005).

4.2. Russia

Syntheses of fossil pollen data and other proxy evidence have been used to produce continental-scale reconstructions of vegetation, soils, and climate during the LIG in the Russian Federation (Grichuk, 1984; Velichko et al., 1991; Lozhkin and Anderson, 1995; Velichko et al., 1998; Zelikson et al., 1998). A difficulty with many of the sites used in these syntheses is that sedimentary deposition from LIG to present is not continuous, and a LIG age for a specific horizon is largely derived from stratigraphic position. Nevertheless, we can surmise with reasonable confidence that the northern forest ecosystems underwent dramatic poleward expansions (Fig. 2).

Forests reached the Arctic Ocean coast across most of the European sector (Grichuk, 1984; Morozova et al., 1998). In western Siberia, the boreal forest (e.g., Siberian spruce (*Picea obovata*), Siberian fir (*Abies sibirica*), Siberian stone pine (*Pinus sibirica*)) was displaced northward by as much as 400 km, and arctic tundra and forest-tundra was likely eliminated from the landscape (Arkhipov and Volkova, 1994). Southern areas of the conifer-dominated forest included broadleaf species (e.g., linden (*Tilia*), elm (*Ulmus*), oak (*Quercus*)), which are characteristic of more southern boreal forest today.

LIG sites are rare in Eastern Siberia; consequently, the character of LIG climate in this region is not well documented. Grichuk (1984) suggested that dark coniferous forest possibly extended from the Yenisei valley eastward to the middle Lena and Aldan basins during the LIG. Light coniferous forest occupied areas north of ~64 °N, extending nearly to the coast. A narrow band of coastal forest-tundra was limited to the vicinity of the Lena delta, continuing slightly to the west.

Forest dominated the vegetation of northeastern Siberia during the peak of the LIG (Lozhkin and Anderson, 1995). Light coniferous forest migrated ~600 km northward, eliminating coastal tundra. The forest also moved a similar distance eastward into areas of Chukotka that are now shrub tundra. Tree birch formed forests in parts of Chukotka, representing northward range extensions of ~1000 (*Betula ermani*) to ~600 km (*Betula cajanderi*). Like today, the interior of Northeast Siberia supported Dahurian larch (*Larix dahurica*) forest, but in contrast to the modern forest, Siberian spruce, Siberian stone pine, and tree alder (*Alnus hirsuta*) were also present. The establishment of Siberian spruce and pine in the upper Kolyma and Indigirka drainages suggest range extensions of 700–800 km. As compared to the modern vegetation, the LIG forests along the northern coast of the Okhotsk Sea contained more tree birch and temperate trees, representing range extensions of up to 1000 km.

The LIG is well represented in the pollen spectra extracted from a sediment core from Lake El'gygytgyn, northern Siberia, within a continuous record of vegetation change over the past 300,000 years (Lozkhin et al., 2006). Although larch pollen is absent in the LIG spectra, climate reconstructions based on statistical comparisons to modern pollen assemblages suggest that summer temperatures were sufficiently warm to support this tree, which is greatly under-represented in pollen records. Such an interpretation is consistent with other studies that suggest an extensive northward displacement of boreal taxa, with open *Betula* forest-tundra and *Larix*-*Pinus pumila* forest established in northeastern Chukotka (Lozhkin and Anderson, 1995).

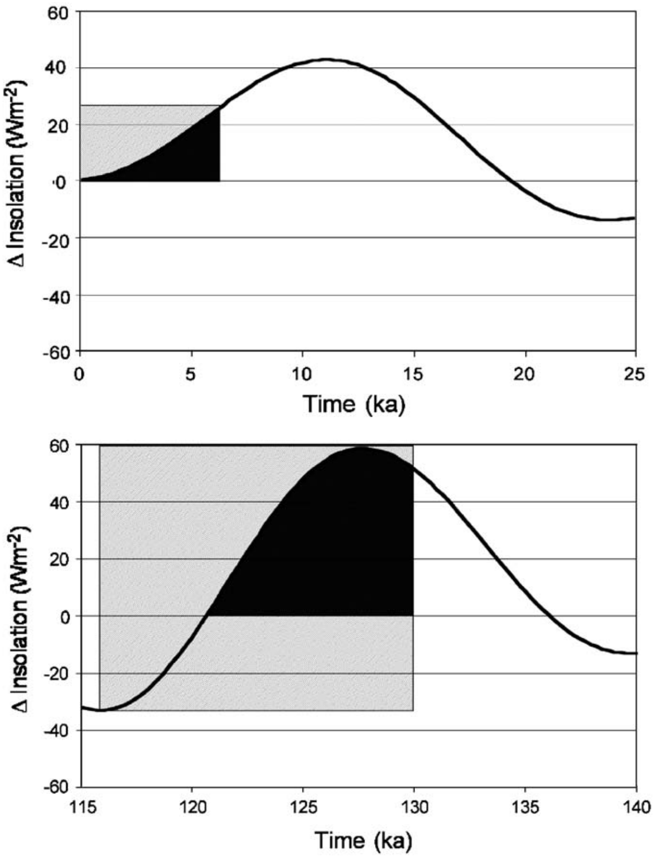

Fig. 2. Mean summer (M, J, J) insolation anomalies (W m^{-2}) expressed as the departure from present at 65°N from 25 ka to present (upper panel) and from 140–115 ka (lower panel). The region where summer insolation was above present and sea level at or above present is shown by solid black fill, whereas the duration of the interglacial, defined as the time when sea level is at or above present, is shown by the hachured box. The larger magnitude and duration of the LIG insolation anomaly relative to the Holocene anomaly results from the combination of a greater LIG insolation anomaly and early penultimate deglaciation.

Furthermore, it is likely that *Betula*, *Alnus*, and possibly *Salix* occurred in large growth forms (trees or tree-sized shrubs).

In summary, the LIG plant communities represent individual range extensions of as much as 1000 km.The western boreal forest contained considerably more birch than does the present-day boreal forest and broadleaf taxa such as hornbeam (*Carpinus*) were found north and east of their modern distributions in European Russia. Dahurian-larch dominated northern boreal forests in northeastern Siberia as is the case today, but in some regions Siberian stone pine, Siberian spruce, and tree alder were important constituents of the light coniferous forest.

Climate estimates based on these floristic differences suggest summer temperatures 0–2 °C above present along the present southern boreal forest belt, 5–8 °C higher than present across central Siberia, and 4–8 °C above present in NE Siberia. Where winter temperatures can be reconstructed, almost all sites indicate milder temperatures, despite reduced winter insolation, suggesting reduced sea ice more than compensated for lower winter insolation. The Lake El'gygytgyn pollen data suggest that maximum summer warmth was 2–4 °C above present and winters were warmer than present (Lozhkin et al., 2006). The Lake El'gygytgyn record also suggests a rapid increase in tree and shrub taxa during the early part of the LIG, where the proportion of tree and shrub pollen rises from 10% to over 90% in a period of what appears to be a few centuries based upon the current chronological control.

4.3. Alaska

Several stratified sites north and west of modern tree line in Alaska document the extension of forest into regions dominated by tundra throughout the Holocene, although LIG ecotonal shifts were not as great as for the Eurasian north. Summer warmth during the LIG is estimated to have been 0–2 °C above present, but winters were 1–3 °C below present. There is a consistent signal for greater moisture at Alaskan sites, but summer temperature increases over present were small compared with other sectors; spatial gradients between the interior and the north and west may have been shallower.

Continuous lacustrine sections through the LIG and fossils in other LIG deposits show establishment of spruce forest at sites in the north and west of Alaska that lie beyond spruce limits characteristic of the Holocene, although tree line displacements were not as great as for the Eurasian north (Brigham-Grette and Hopkins, 1995). Sedimentation at Squirrel Lake, NW Alaska, was continuous from MIS 6 through to the present. Temperature estimates using the modern analogue technique on pollen assemblages in LIG beds at Squirrel Lake indicate a modest (1–2 °C) increase in mean July temperature (Anderson, 2005). The temporal sequence indicates that warming occurred early and rapidly and that high moisture levels were attained early in the interglacial. Pollen from LIG levels in Ahaliorak Lake suggests that mean January temperatures were 1–3 °C cooler and annual precipitation slightly higher than modern (Eisner and Colinvaux, 1990). Fossil beetle assemblages from NW Alaska suggest summer temperatures similar to, or 1–2 °C warmer than present, and winter temperatures similar to, or 1–2 °C colder than present (Edwards et al., 2003). In interior Alaska, LIG deposits are related to the stratigraphic marker of the Old Crow tephra (Hamilton and Brigham-Grette, 1991). A LIG forest bed and paleosol near Fairbanks that contains spruce needles and pollen indicating summers as warm as present also contains spores and chemical weathering characteristics suggesting conditions considerably wetter than present (Péwé et al., 1997; Muhs et al., 2001).

4.4. Arctic Canada

Although tundra remained dominant over much of far northern Canada during the LIG, pollen, macrofossil and insect data indicate warmer-than-present summers in the Queen Elizabeth Islands (Matthews et al., 1986). Boreal forest advanced to the Arctic Ocean coast in the NW of mainland Canada (Rampton, 1988) and north of its present position in the Hudson Bay region (Nielsen et al., 1986; Dredge et al., 1990; Mott and Dilabio, 1990; Wyatt, 1990), changes that require only modest warming. In contrast, four stratified LIG lacustrine sequences from the eastern Canadian Arctic that capture the LIG document summer temperatures 4–8 °C above present (Table 1). For two nearby lakes (Fog, Brother-of-Fog), LIG summer temperatures are independently reconstructed from pollen and chironomid analyses from the same sediment cores. From Fog Lake, pollen indicates peak LIG summer temperatures 3–5 °C above present, whereas chironomids indicate 7 ± 2 °C above present. At Brother-of-Fog Lake, pollen indicates 3–4 °C warmer, and chironomids indicate 8 ± 2 °C warmer than present (Francis et al., 2006; Fréchette et al., 2006). Collectively, the proxy data indicate a LIG summer temperature 4–7 °C above present. Pollen data are also available from two other LIG lakes (Amarok (Fréchette et al., 2006) and Robinson (Miller et al., 1999; Fréchette, 2005) lakes), both suggesting LIG summers ca 5 °C warmer than present. A probable LIG site in central Baffin Island contains plant and insect remains indicating summer temperatures 4–5 °C above present (Morgan et al., 1993). Shallow-water marine records of probable LIG age are common along the Baffin Island coast, and contain extralimital southern invertebrates indicative of a greater flux of Atlantic Water into Baffin Bay and reduced summer sea ice (Miller, 1985), although the chronologies for the marine sites are less certain than for the terrestrial sites and SSTs have not been quantified.

The absence of any ice older than the LIG at the base of the larger Canadian ice caps indicates they melted completely during the LIG, only to reappear when climate deteriorated during the transition into the ensuing glacial period (Koerner and Fisher, 2002). Many of these ice caps survived the warm early Holocene period but have negative balances today and would again disappear if the present warm period persisted. However, the time to melt all ice in Arctic Canada (several millennia) almost certainly is longer than that which could be maintained by the effects of anthropogenically produced greenhouse gases.

4.5. Greenland

Terrestrial LIG conditions over Greenland are defined from ice-core and biological proxies. A rich and diverse assemblage of bryophytes, vascular plants, beetles and other invertebrates from terrestrial and shallow marine LIG localities along central East Greenland indicate summers 5 °C warmer than present (Bennike and Böcher, 1994). Extensive birch forests are inferred, suggesting an ice-free southern Greenland. Pollen and spores from LIG levels in a core from the Labrador Sea off SW Greenland also suggest that adjacent southern Greenland was free of inland ice, with a sub-arctic or even temperate climate (Hillaire-Marcel et al., 2001). Extralimital southern taxa found in LIG beds at Thule, NW Greenland, require summer temperatures at least 4 °C warmer than present (Bennike and Böcher, 1992; Kelly et al., 1999).

Two ice cores from Greenland capture undisturbed LIG layers, NorthGRIP (North Greenland Ice Core Project members, 2004) and Renland (Johnsen et al., 2001), although neither extends through the LIG and into the preceding glacial period. Ice-isotopes in LIG layers from both cores indicate that the LIG was 5 °C warmer than present. The impact of warmer summers on the dimensions of the Greenland Ice Sheet remains debated. The absence of a continuous record of undisturbed ice through the LIG, and the discordance between ice isotopes in the older portions of cores from DYE-3 and Camp Century relative to the benchmark Summit and NorthGRIP cores, suggest that significant reduction in the ice sheet may have occurred, with both Camp Century and DYE-3 ice free in the LIG (Koerner, 1989; Koerner and Fisher, 2002). Total gas evidence from Eemian ice in the GRIP ice core indicates that the Summit region may have been up to 500 m lower than present at some time in the LIG (Raynaud et al., 1997). Ice-sheet modeling (Cuffey and Marshall, 2000; Otto-Bliesner et al., 2006) suggests that the Greenland Ice Sheet may have been reduced by as much as half its current volume during the LIG, including the loss of its southern dome, whereas the northern dome is modeled to remain within 700 m vertically of its current elevation (Otto-Bliesner et al., 2006), implying an ice sheet with a much steeper profile.

The ice core data has also been used to make a case for a relatively stable Greenland Ice Sheet through the LIG. LIG ice is present in all deep ice cores from Greenland, and this ice has δ^{18}O values ca 3‰ heavier than typical Holocene δ^{18}O values in the same cores (except DYE-3; North Greenland Ice Core Project members, 2004). This constant isotopic offset has been used to suggest that the ice sheet was not dramatically smaller in the LIG than in the Holocene, with the southern dome remaining intact through the LIG, although somewhat thinner (North Greenland Ice Core Project members, 2004). Regardless of the exact configuration of the Greenland Ice Sheet, all interpretations of the ice core data indicate LIG temperatures 5 °C above present.

4.6. Arctic ocean

The Arctic Ocean remains the least understood ocean basin; access is difficult and many standard climate proxies are compromised by dissolution and meltwater overprinting, especially in the central Arctic Ocean basin where even the development of reliable geochronologies has been a challenge (Jakobsson et al., 2003; Backman et al., 2004). The LIG is usually identified by a combination of coccoliths, low IRD, and increased biological productivity.

Direct solar insolation, river runoff, sea level, and the intensity and pattern of North Atlantic Drift (NAD) inflow are expected to be the dominant determinants of surface water characteristics in the Arctic Ocean during the LIG. Concerted efforts over the past decade have led to the recovery of high-quality sediment cores from the Arctic Ocean, and improvements in their geochronology and the interpretation of environmental proxies extracted from the sediment. This has led the way to an emerging consensus view of the LIG, especially in the marginal seas, with significant improvements in our understanding of the central interior (Spielhagen et al., 2004).

A stronger flux of warm, salty Atlantic water into the Arctic Ocean started early in the LIG (Matthiessen et al., 2001b; Wollenburg et al., 2001). This influx led to greatly reduced sea ice in the marginal seas (Knies and Vogt, 2003), although not all core sites support this interpretation. Site-to-site differences in LIG surface water characteristics may indicate changing trajectories of Atlantic Water currents between the Holocene and the LIG, and even within the LIG; Additional uncertainties may be related to changes in carbonate dissolution rates (see discussions in Matthiessen et al., 2001b). Seasonal increases in primary productivity in ice-free areas and along extensive leads supported a more diverse and abundant nano- and micro-fauna (Gard and Backman, 1990; Matthiessen et al., 2001a; Matthiessen and Knies, 2001; Wollenburg et al., 2001). The increased flux of Atlantic water reduced the vertical stratification over many of the marginal seas, allowing Atlantic water to remain closer to the surface than in the Holocene. Atlantic surface water characterized the Arctic coast of Russia from the White Sea to eastern Taimyr, 2000 km farther east and 3–4 °C warmer than in the Holocene (Funder et al., 2002; Grøsfjeld et al., in review; Svendsen et al., 2004). Delayed winter sea ice formation along the Eurasian coast would have led to significant warming of the winter atmosphere, despite negative winter insolation anomalies at that time. This may help to explain the maintenance of boreal forest far north of its current position across Eurasia during the LIG. Conifer trees are sensitive to extreme winter cold, which would have been mitigated by reduced coastal-zone sea ice.

LIG beaches along northern Alaska contain southern extralimital mollusk taxa indicative of SST 3 °C above present, and extensive storm beaches that require sea ice reduction by 800 km in winter (Brigham-Grette and Hopkins, 1995).

Although the status of sea ice over much of the central Arctic Basin is still debated, the available data suggest that sea ice remained through the summer in the central basin, possibly as far as NE Svalbard throughout the LIG (Spielhagen et al., 2004). On the other hand, relatively high concentrations of planktonic foraminifera and the presence of coccoliths in LIG levels of cores from the Lomonosov Ridge, central Artic Ocean, suggest that seasonally open waters or extensive leads characterized the eastern and central Arctic Ocean during portions of the LIG (Svendsen et al., 2004).

4.7. Nordic seas

Identification of the LIG is more secure for the Nordic Seas than for the Arctic Ocean. Standard marine isotope stages can be recognized by diagnostic patterns of $\delta^{18}O$ in benthic foraminifera, although with some meltwater overprinting during the deglaciations (Bauch et al., 1996; Fronval et al., 1998), supported by the occurrence of Icelandic tephra during the LIG (Sjoholm et al., 1991), low levels of IRD, and increased concentrations of sub-polar faunal and floral species reflecting a northward displacement of the Polar Front. During the past 150 ka, only the Holocene and the LIG show evidence of a strong and sustained incursion of Atlantic Water into the Nordic Seas (Kellogg, 1980). Higher LIG SST characterize the Scandinavian coast well into Arctic Russia (Mangerud et al., 1981; Raukas, 1991; Velichko et al., 1991; Funder et al., 2002), suggesting warm North Atlantic current close to the coast of Norway, extending well into the Eurasian coast of the Arctic Ocean. Cores from the southwestern Nordic Seas, along the coast of E. Greenland and western Svalbard, south of Jan Mayen, and in northern Baffin Bay indicate LIG SST above those of the Holocene (Miller et al., 1989; Mangerud and Svendsen, 1992; Sejrup et al., 1995; Funder et al., 1998; Bauch et al., 1999; Kelly et al., 1999), but this difference is not reflected in many cores from the eastern Nordic Seas. An alternative position is that Eeman SSTs in the Nordic Seas were lower than during the Holocene (Bauch et al., 1999). Conflicting interpretations of SSTs in the Nordic Seas during the Eemian and Holocene could be related to the manner by which different proxies respond to changing water masses, within-interglacial variability at millennial- to submillennial-scales that complicates the interpretation of low-resolution records, or changes in the precise thread of warm, salty North Atlantic water as it moved through the Nordic Seas into the Arctic Ocean. Although some Atlantic surface water was routed through the Baltic into the White Sea early in the LIG, the capacity of this connection was not sufficient to have an impact on the general surface circulation, and isostatic emergence severed the passage ca 3000 yr into the LIG (Funder et al., 2002). The evidence from the Arctic Ocean (above) necessitates a greater dominance of Atlantic water at the surface of the Arctic Ocean. This could be accomplished by some combination of intensified NAD, stronger surface water mixing in the Arctic Ocean, or reduced freshwater runoff; an intensified flux of Atlantic water into the Arctic Ocean must have come through the Nordic Seas.

4.8. Sub-polar North Atlantic

Conditions in the sub-polar North Atlantic are relevant to the transfer of energy between low and high latitudes.

North Atlantic SSTs were stable and similar to, or up to 2 °C higher than present throughout the LIG (Bauch et al., 2000); SSTs warmer than for the Holocene are most apparent in the eastern North Atlantic at ca 50 °N. The strength of deepwater production appears to have been similar to present (Adkins et al., 1997; Hall et al., 1998; Lehman et al., 2002; McManus et al., 2002; Backman et al., 2004), although there are few reliable quantitative proxies of past deep-water production rates.

5. Discussion

5.1. The Arctic at peak LIG warmth

All sectors of the Arctic register summers warmer than present during the LIG, but the magnitude of warming exhibits spatial variability (Fig. 3). The greatest positive summer temperature anomalies occur around the currently arctic regions of the Atlantic sector, where summer warming was typically 4–6 °C. This anomaly extends into Siberia, where summer temperature anomalies were 4–8 °C, presumably due to the impact of strong insolation forcing over a large continent supplemented by the penetration of Atlantic water well into the Arctic Ocean along the Russian coast. The anomalies decrease from Siberia westward to the European sector (0–2 °C), and eastward toward Beringia (2–4 °C). The Arctic coast of Alaska indicates warmer SST (3 °C) and considerable summer sea ice reduction, but much of interior Alaska registers small anomalies (0–2 °C) that probably extend into western Canada, although quantitative data to evaluate this trend is scarce. In contrast, northeastern Canada and Greenland register summer temperature anomalies of at least 5 °C. The Arctic Ocean records greater advection of warm, salty Atlantic water through the Nordic Seas, although Atlantic water did not dominate surface waters of the central Arctic Ocean. Surface waters were warmer in the LIG than at any time in the Holocene in Baffin Bay and along the north coast of Alaska, where sea ice was also much retracted.

Precipitation and winter temperatures are more difficult to reconstruct for the LIG than are summer temperatures. In northeast Europe, the later part of the LIG was characterized by a marked increase in winter temperatures. A large positive winter temperature anomaly is seen also in Russia and western Siberia, although the timing is not as well constrained (Troitsky, 1964; Gudina et al., 1983; Funder et al., 2002). Precipitation is more difficult to quantify, but most sectors with qualitative estimates indicate wetter conditions than in the Holocene.

Despite some uncertainty in the SSTs of the Nordic Seas during the LIG, the general consensus is that there must have been a greater flux of Atlantic water through the Nordic Seas into the Arctic Ocean. The precise path this flow followed may have differed from the present route, especially early during the LIG when the marginal shelves had yet to recover from isostatic depression under the MIS 6 ice sheets. From the pattern of LIG SSTs, the eastern branch of the Norwegian Current intensified and reached

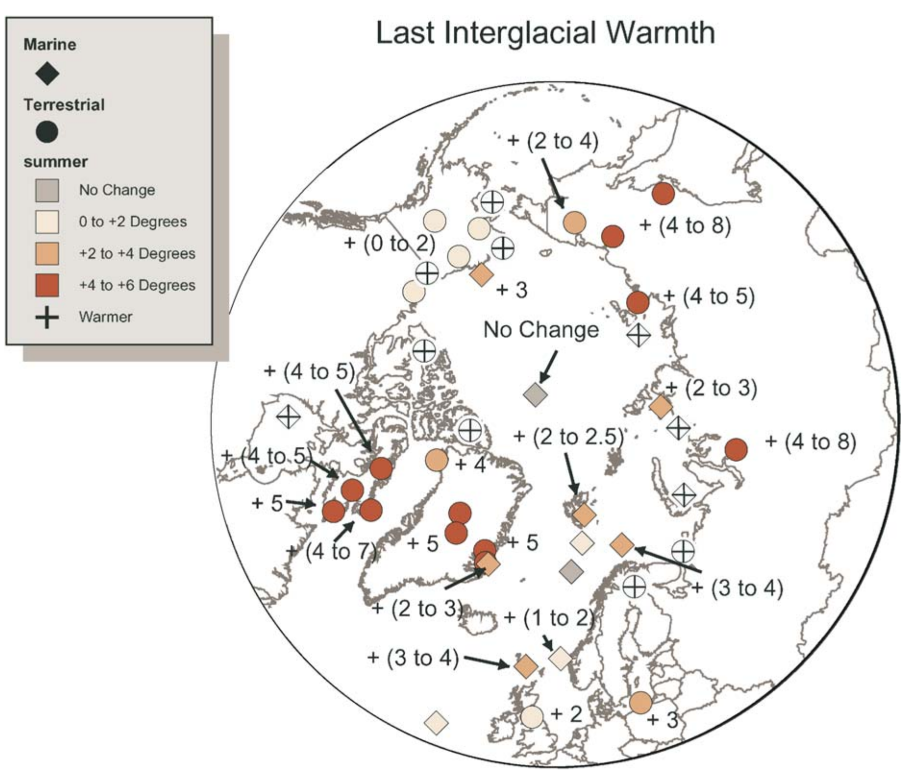

Fig. 3. Polar projection showing regional maximum LIG summer temperature anomalies relative to present derived from paleotemperature proxies (derived from Tables 1 and 2). Terrestrial sites in circles, marine sites in squares.

the White Sea region already a few centuries into the LIG (Mangerud et al., 1998; Grøsfjeld et al., in review). The western branch may have followed a path different than present, but reached the East Greenland coast early in the LIG (Funder et al., 1998). As isostatic recovery progressed, the trajectory of the NAD likely evolved through the LIG. The increased poleward transfer of Atlantic water implies a larger production of deepwater, currently estimated to be about 15 Sv ($10^6\,m^3\,s^{-1}$), but no quantitative LIG estimates are available. Alternatively, the greater influx of Atlantic water may have been countered by a greater outflow of surface water from the Arctic Ocean, which could have changed the position of the Polar Front in the Nordic Seas (Fronval et al., 1998).

Several sites suggest that the transition into the LIG was rapid, with peak warmth early in the interglaciation. Although not all Arctic LIG sites offer sufficient resolution to evaluate the pattern of warming during the LIG, those that do exhibit rapid warming to a peak temperature near the beginning of the LIG. The pattern of rapid initial warming, followed by slow cooling is mirrored in all four Antarctic ice cores that sample the LIG (Byrd, Vostok, Dome Fuji, and Dome C (Johnsen et al., 1972; Petit et al., 1999; Watanabe et al., 2003; EPICA Community Members, 2004), despite negative summer insolation forcing in the Southern Hemisphere at this time. The apparent concordance of Antarctic ice core and Northern Hemisphere summer temperature records suggests that the oceans may have transferred Northern Hemisphere warmth into the Southern Hemisphere.

5.2. *Why was the LIG in the Arctic so warm?*

Arctic summers during peak LIG warmth were 5 °C above present over most of the region, considerably higher than during the Holocene thermal maximum (Kaufman et al., 2004). The pattern and magnitude of LIG summer temperature anomalies in the Arctic is largely related to four phenomena: (1) strong summer insolation forcing, (2) optimal phasing of penultimate deglaciation in relation to the orbitally controlled insolation maximum, (3) increased meridional heat transport through the Nordic Seas into the Arctic Ocean, and (4) strong positive feedbacks associated with changes in sea ice, land ice and snow cover, and the poleward expansion of boreal forests.

A strong positive summer insolation anomaly across the Northern Hemisphere through the first half of the LIG is the dominant forcing that explains the observed Arctic warmth. Average LIG summer (M, J, J) insolation anomalies are positive in the Arctic from ca 138 until ca 121 ka, with anomalies between 131 and 127 ka much larger than at any time in the Holocene (Fig. 1). Peak LIG summer insolation at the top of the atmosphere across the Arctic (60–90 °N) was ca 13% higher than present, whereas it was ca 9% higher at the peak Holocene anomaly, 11 ka (Fig. 4). In most regions, LIG summer temperatures were well above those of the Holocene thermal maximum, consistent with greater LIG insolation forcing. However, the reconstructed thermal response of the Arctic at peak Holocene insolation forcing 11 ka (CAPE Project Members, 2001), and at the ice-volume minimum 6 ka (Bigelow et al., 2003) was, by comparison, muted. Tree line advances were modest and relatively short-lived in most sectors, and temperature anomalies at the Holocene thermal maximum were also modest (ca 1.6 °C; Kaufman et al., 2004).

The weaker thermal response of the Arctic in the Holocene is only partly explained by the insolation anomaly differences; phasing of deglaciation also must be important. A rigorous evaluation of all uranium-series dates on corals from stable platforms confirms that sea level reached modern levels (global ice volume similar to present) by 130 ± 2 ka (summarized in Overpeck et al., 2006). LIG summer (M, J ,J) insolation at the top of the atmosphere peaked between 131 and 127 ka (Fig. 1). Because the penultimate (MIS 6) ice sheets disappeared (sea level at or above present) by 130 ka, the full magnitude of the summer insolation anomaly was available to heat the Northern Hemisphere, rather than to melt residual ice sheets. This contrasts with the Holocene, when peak Northern Hemisphere insolation occurred 5 ka before the Northern Hemisphere ice sheets had melted and sea level attained modern levels (6 ka; Fig. 2). Consequently, during early Holocene peak summer insolation, much of the insolation anomaly was consumed in the melting of residual ice sheets, rather than heating the land/ocean/atmosphere, dampening the climate response to insolation forcing (Koerner, 1989; Bigelow et al., 2003; Kaplan et al., 2003; Vavrus and Harrison, 2003). By the time sea level reached present in the Holocene (6 ka), the high latitude Northern Hemisphere summer (M, J, J) insolation anomaly was ca 15 $W m^{-2}$, whereas at the comparable time in the LIG (130 ka) it was ca 45 $W m^{-2}$, three times as large.

Changes in oceanic surface currents may explain some of the LIG spatial variability across the Arctic. Although the insolation anomaly is symmetric about the Arctic, the expansion of forest ecotones north of their northernmost Holocene limits is most pronounced in Eurasia, and less so in Alaska, and central and northwestern Canada. A plausible explanation for this asymmetry is the increased flux of warm Atlantic surface waters northward around the Eurasian Arctic. Atlantic water along the Arctic coast of Russia would have limited the formation of sea ice, minimizing extreme winter cold, which is a strong deterrent to poleward expansion of evergreen trees. The flux of Atlantic water may have also influenced climate by steering relatively warm air masses into the Eurasian North.

The magnitude of LIG summer warmth observed across the Arctic is not mirrored by comparable warmth at lower Northern Hemisphere latitudes, despite similar forcing. Peak LIG summer insolation anomalies averaged only slightly below the Arctic (60–90 °N) values at low (0–30 °N; 20% less) and mid (30–60 °N; 10% less) latitudes. Although hemispheric summaries of LIG summer temperature anomalies have not been compiled, the available

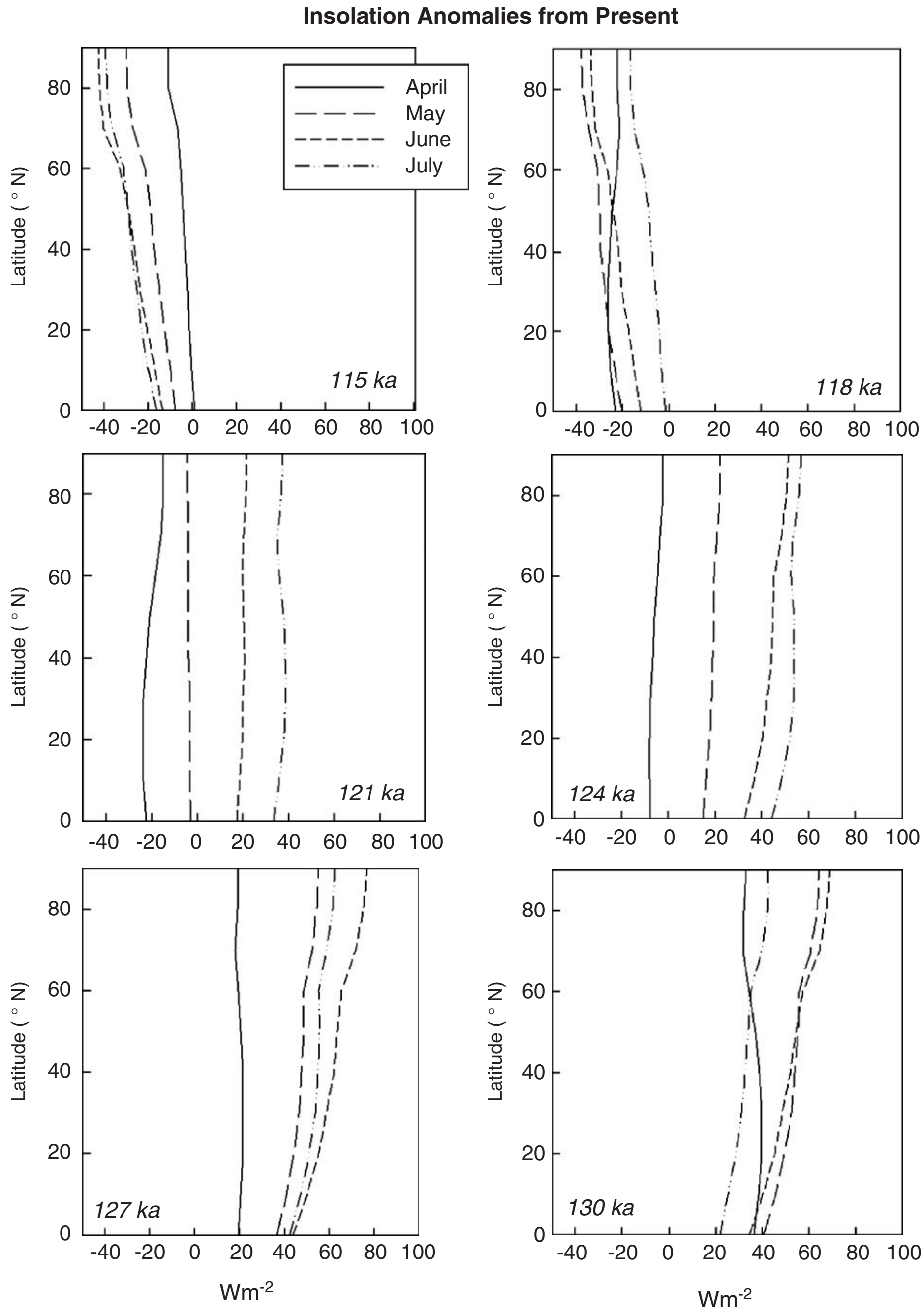

Fig. 4. Insolation anomalies expressed as a percentage deviation from present for the months of April (blue), May (red), June (yellow), and July (green), plotted against Northern Hemisphere latitude every 3 ka from 130–115 ka, showing the evolution of the insolation anomaly through the LIG, and its latitudinal gradient (data from Berger and Loutre, 1991).

records suggest that peak LIG warmth at mid- and low-latitudes in the Northern Hemisphere was between 0–2 °C above present.

The greater thermal response of the Arctic in the LIG than at lower latitudes cannot be explained easily by the modest latitudinal gradients of LIG summer insolation forcing (Fig. 4). The most likely explanation for this discrepancy is the strong positive feedbacks associated with reduced sea ice/snow cover and the expansion of boreal forests. Increased summer insolation and the poleward transport of heat by ocean currents led to reduced sea ice in the Arctic Ocean and its marginal seas. Positive feedbacks associated with reduced sea ice in both summer and winter would have amplified primary insolation forcing. The replacement of tundra with boreal forest across much of the Arctic landscape would have decreased its albedo, providing another strong positive feedback that would amplify Arctic warming (Chapin et al., 2005; Serreze and Francis, 2006).

We postulate that the cumulative effect of positive feedbacks from reduced sea ice, reduced permanent land ice, reduced seasonal snow cover and the expansion of boreal forests northward, coupled with increased meridional heat transport by oceanic surface currents and the atmosphere, amplified the modest LIG summer insolation gradient to produce the substantially stronger thermal response seen in the Arctic relative to lower latitudes in the Northern Hemisphere. Climate modeling of the LIG also underscores the importance of snow, ice and vegetation feedbacks that amplify insolation-driven Arctic warmth (Crucifix and Loutre, 2002; Otto-Bliesner et al., 2006).

6. Conclusions

Quantitative reconstructions of LIG summer temperatures suggest that much of the Arctic was 5 °C warmer during the LIG than at present, and that this warming occurred rapidly and reached peak warmth during the earliest portion of the interglaciation. Arctic sea ice was reduced to a greater extent in the LIG than during the early Holocene due to both greater summer insolation and the larger flux of relatively warm Atlantic surface water into the Arctic Ocean during the LIG than at any time in the Holocene. Changes in surface ocean characteristics influenced the planetary energy balance through ice-albedo feedback; reduced winter sea ice and increased ventilation rates impacted climate over adjacent lands, especially in winter. The magnitude and spatial gradients of LIG Arctic summer temperature anomalies are related to the timing and magnitude of the orbitally controlled summer insolation anomaly in conjunction with optimal phasing of penultimate deglaciation, increased meridional heat transport through the Nordic Seas into the Arctic Ocean, and strong positive feedbacks associated with reduced land ice and snow, retraction of sea ice and expansion of boreal forests.

Polar amplification of climate change is predicted by most climate models. However, the observational records of 20th century warming are not in perfect accord with model projections, an observation that has fed a lively debate (Serreze and Francis, 2006). During the 20th century, the planetary temperature increased 0.7 °C, whereas most regions of the Arctic record warming of 1–3 °C over the same interval (Serreze et al., 2000). But, most of the warming occurred in summer months, whereas model projections indicate winter warming should dominate. The paleoclimate record is more direct. The average planetary temperature depression during the peak of the last glacial maximum (ca 20 ka) is generally estimated to be 6 to 8 °C, whereas Arctic cooling, as measured by borehole temperatures in the Greenland Ice Sheet, was ca 25 °C (Cuffey et al., 1995; Dahl-Jensen et al., 1998). Our reconstructions of LIG Arctic warmth show similar anomalies, with summers averaging 5 °C warmer than present across most of the Arctic, whereas the Northern Hemisphere average summer temperature is estimated to be only 0–2 °C warmer than present, despite similar insolation forcing. These discrepancies are consistent with strong positive feedbacks in the Arctic that amplify primary insolation forcing.

A quantitative assessment of LIG Arctic summer temperatures informs projections of future greenhouse warming. Although LIG summer warmth is primarily the result of a peak summer insolation anomaly, the responses of land and sea ice, and vegetation, may have provided feedbacks that not only amplified summer warmth, but also contributed to winter warmth, counteracting, in part, negative forcing from negative winter insolation anomalies (Fig. 1). In contrast, projected greenhouse forcing does not exhibit seasonality and is predicted to increase both winter and summer temperatures. Warmer winters would contribute to the poleward migration of coniferous forest, and via earlier spring melting, accelerate summer snow and ice melting (Guetter and Kutzbach, 1986; Chapin et al., 2005). Thus, the pattern of warming may differ in the coming decades from those observed in the LIG. However, the strong thermal response of the Arctic to primary summer insolation forcing during the LIG suggests that future Arctic changes related to the continuing build up of greenhouse gases are likely to be much larger than at lower latitudes.

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