Climate changes relative to 0 ka

The main goals of this section are (1) to characterize simulated 21 and 14 ka climate anomalies relative to the preindustrial control simulation. Due to the boundary conditions forcing the glacial climate, our model results are broadly consistent with previous GCMs of the LGM (CLIMAP, 1981; Braconnot et al., 2007a), where climate anomalies and differences between simulations are minimal in the Southern Hemisphere. Therefore, we focus our attention to the Northern Hemisphere where greater variability is simulated.

S1 Atmospheric Anomalies

S1.1 Surface air temperature

Both 21 ka simulations show global cooling with the largest surface air temperature (SAT) anomalies (21 ka minus 0 ka) in the Northern Hemisphere where LGM ice sheets exert their influence. SAT anomalies are greatest directly over the LGM ice sheets due to the large changes in elevation and albedo, but there is also substantial downstream cooling over most of northern Asia and the northern oceans (Figure S1a).

We use the differences in elevation and the resulting SAT at each time slice to estimate summer temperature lapse rates over the LIS by regressing the change in SAT onto the change in elevation. Using this approach, we estimate summer ice-sheet lapse rates of -4.6 ^{o}C/km for 21 ka, which is consistent with the published range of -4 to -7 ^{o}C/km of Abe-Ouchi et al. (2007), but slightly lower than earlier estimates of -6 to -8 ^{o}C/km (Marshall et al., 2000; Pollard & PMIP, 2000; Charbit et al., 2002). Applying this lapse rate over the LIS compared to modern elevation accounts for 40-50% of the SAT anomalies over the LIS. The remaining change in SAT temperature is due to other ice-sheet impacts (i.e. albedo) and changes in atmospheric circulation.

To the extent that borehole measurements capture SAT from the LGM, our results are ~1 ^{o}C colder than globally integrated borehole estimates that suggest LGM cooling of 4.3±0.2 ^{o}C (Huang et al., 2008). Borehole measurements from Summit Greenland suggest LGM cooling of

14-16 ^{o}C (Cuffey et al., 1995; Cuffey and Clow, 1997), which is colder than our model anomalies of -9.9±0.7 ^{o}C (21ka-L) and -11.0±0.8 ^{o}C (21ka-5). This difference could in part be due to the smoothed topography in ModelE2-R relative to the actual elevation of the ice-core site, as well as due to differences in the temperature inversion over the ice sheet (Cuffey and Clow, 1997; Alley, 2000). An alternate estimate adjusting for possible changes in the temperature inversion over Greenland suggests that Greenland LGM cooling to be ~10 ^{o}C (Alley et al., 2010). Boreholes from Germany, Slovenia, and the Czech Republic suggest LGM cooling of 7-10 ^{o}C (Safanda & Rajver, 2001), which agree with our model anomalies of -10.1±0.7 ^{o}C (21ka-L) and -9.5±0.6 ^{o}C (21ka-5G).

For the 14 ka simulations, SAT anomalies continue to be negative (Figure S2a), with global mean SAT anomalies of -3.0±0.1 ^{o}C (14ka-L) and -2.8±0.1 ^{o}C (14ka-5G). These values are consistent with the global integrated borehole estimate of -2.9±0.6 ^{o}C for the same time period (Huang et al., 2008). Data from Summit Greenland suggest 14 ka cooling of 7-9 ^{o}C (Cuffey and Clow, 1997), which is comparable with our 14 ka anomalies at this location of -6.4±0.7 ^{o}C (14ka-L) and -6.9±0.6 ^{o}C (14ka-5G). The greatest cooling occurs directly over the ice sheets with downstream cooling by as much as 10 ^{o}C over northern Asia (Figure S2a). Again, we estimate a summer ice-sheet lapse rate of ~4.8 ^{o}C/km over the LIS (calculated by regressing change in SAT onto change in elevation between 14ka-L and 14ka-5G, as above), which is consistent with values from the 21 ka simulations.

S1.2 Precipitation minus evaporation

At 21 ka and 14 ka, precipitation is largely diminished, reflecting the decrease in atmospheric water vapor (global mean specific humidity anomalies of -20.2, -19.6, -11.9, and -11.4 g water vapor per kg dry air for 21ka-L, 21ka-5G, 14ka-L, 14ka-5G, respectively, tracking relative anomalies in global mean temperature). In general, the largest precipitation minus evaporation (P-E) anomalies occur in the tropics (Figures S1b and S2b) reflecting a southward displacement of the Inter-Tropical Convergence Zone (ITCZ) as seen by the P-E over the tropical Pacific, Atlantic, South America, and Africa. This shift in the ITCZ is consistent with models and proxy records of the glacial hydrologic cycle (e.g., Peterson et al., 2000; Thompson et al., 2000; Chiang et al., 2003; Wang et al., 2004; Broccoli et al., 2006; Braconnot et al., 2007b; Arbuszewski et al., 2013).

In all simulations, enhanced precipitation over the Indonesian archipelago and the Arafura Sea north of Australia is likely due to a too large of land-sea contrast in ModelE2, with enhanced drying over the surrounding water masses. This feature likely traces back to too many low clouds in the tropics in ModelE2. This is not unique to these simulations and exists in all the CMIP5 runs by ModelE2; newer versions of ModelE2 have addressed this issue (delGenio personal communication, 2013). The lower sea level also subaerially exposes much of this region, contributing to this enhanced precipitation.

S1.3 δ^{18}O of the atmosphere

LGM δ^{18}Oa anomalies largely reflect the reduction in SAT, with the greatest depletion in δ^{18}O occurring directly over the ice sheets and across northern Asia (Figures S1a and S1c). However, this direct coupling of SAT and δ^{18}Oa does not hold everywhere. Areas of slight enrichment occur over nearly all tropical oceans basins in regions of reduced precipitation (Figure S1b), despite globally colder SAT. In contrast, 21 ka anomalies are particularly depleted over northern Australia and into Indonesia where there is enhanced P-E.

Direct comparison of the simulated change in δ^{18}Oa relative to 0 ka with terrestrial records that span this period shows that the 21 ka simulations capture the general change in δ^{18}Oa where such proxy records exist (see colored dots in Figure S1c). However, the simulations seem to do poorly in capturing the tropical speleothem records that are heavily influenced by precipitation seasonality (Wang et al., 2001; Bar-Matthews et al., 2003; Holmgren et al., 2003; Dykoski et al., 2005; Partin et al., 2007; Cheng et al., 2012). Despite this bias, the δ^{18}Oa anomalies of 21ka-L and 21ka-5G correlate well with these data with $r=0.72$ and $r=0.67$, respectively (correlations are significant with $p\leq0.01$).

The δ^{18}Oa anomalies at 14 ka also reflect the reduction in SAT, particularly over the Northern Hemisphere ice sheets and northern Asia (Figure S2a and S2c). The enrichment of δ^{18}Oa over the tropical oceans is no longer evident, but the 14 ka depletion over northern Australia and into Indonesia still persists in concert with the enhanced P-E anomaly (Figure S2b). The comparison with proxy data again captures the general spatial trends in 14 ka δ^{18}Oa anomalies (Figure S2c). There is still some poor mismatch with some of the anomalies in the tropical records, but the 14ka-L simulation overall correlated well with the proxy measurements ($r=0.6$, $p\leq0.05$). However, the correlation of 14ka-5G is not as strong and is less significant ($r=0.4$, $p=0.19$). Unfortunately majority of proxy records are concentrated where modeled anomalies are small, thus limiting the testing of the model results. Proxy records from northern Asia would be a better test of the models ability to capture large trends in δ^{18}Oa.

S1.4 Circulation and pressure

The emplacement of the LIS and an enhanced Northern Hemisphere equator-pole temperature gradient at both 21 ka and 14 ka sets of simulations result in the strengthening and southward shift of the polar jet (Figure S1e), as in previous simulations (Kageyama et al., 1999; Arpe et al., 2011; Hofer et al, 2012). In addition, the north-south Rossby wave nature of the jet appears to be enhanced, particularly downstream of the ice sheets, reflecting greater instability in the mean state of the jet (Donohoe and Battisti, 2009). This shift in jet location influences surface winds with enhancement over much of the North Atlantic and North Pacific (Figure S1d). In the Southern Hemisphere, however, there is a weakening of the subtropical jet, relative to 0 ka, and surface wind anomalies are diminished.

At 14 ka, both simulations continue to have a southward shift in the polar jet (Figure S2e). The anomalies are more zonal than in the 21 ka simulations, suggesting that Rossby Waves in the 14 ka jets is more similar to the control simulation. The southward displacement and enhancement of the polar jet over the midlatitudes leads to elevated surface winds across much of the North Atlantic and into western Asia (Figure S2d). Again, in the Southern Hemisphere, the subtropical jet is largely reduced relative to the control simulation.

Upper level air pressure is associated with these changes in atmospheric circulation as well as changes in SAT. In general, the enhanced zonal mean meridional temperature gradient during the glacial simulation drives a lowering of 500 mb geopotential heights in the higher latitudes. However, removing the zonal mean in the 500 mb heights reveals a series of planetary waves from west to east (Figure S1f) that are reflected in the meridional wave nature of the atmospheric jet (Figure S1e). Most notably, the 21 ka simulations show a deepening of a trough immediately downstream of the LIS, as well as an enhanced trough over eastern Asia and an enhanced ridge over Beringia, immediately upstream of the LIS. Previous sensitivity experiments testing the impact of LGM ice sheet elevation on atmospheric circulation have described similar patterns in geopotential heights as the "stationary wave effect" (Broccoli and Manabe, 1987; Cook and Held, 1988; Abe-Ouchi et al., 2007). The 500 mb height anomalies in the 14 ka simulations reflect a similar pattern of stationary wave generation with deepening of troughs immediately downstream of the LIS and Scandinavian Ice Sheet and an enhanced ridge immediately upstream of the LIS (Figure S2f).

S1.5 Albedo

Surface albedo anomalies at 21 ka reflect the colder Northern Hemisphere temperatures and enhanced perennial snow cover over the ice sheets and across northern Asia (Figure S1g). In addition, the expansion of sea ice in the Nordic Seas and the Northwest Pacific leads to an increase in surface albedo (Figure S1g). Planetary albedo largely shows the same pattern as ground albedo, indicating that changes in shortwave reflectivity is primarily driven by surface cover changes instead of cloud cover (Figure S1h).

Surface albedo at 14 ka is also enhanced directly over the ice sheets as well as across northern Asia, where there continues to be greater duration of seasonal snow cover relative to 0 ka (Figure S2g). Elevated albedo over the Nordic Seas is again associated with sea ice expansion in this region, and the trends in ground albedo are largely mimicked by planetary albedo (Figure S2h), suggesting that changes in low clouds only play a minor role on the total albedo anomaly.

S2 Ocean anomalies

S2.1 Ocean temperature

Similar to SAT, global sea surface temperatures (SST) are colder in the 21 ka simulations, with particularly strong cooling over the North Pacific (Figure S3a). Some proxy reconstructions have suggested warmer than present conditions in this region (CLIMAP, 1981; Waelbroeck et al., 2009), but most recent GCM simulations do not have this feature (Braconnot et al., 2007a). The warmer than present interpretation may be a result of no-analogue issues with foraminifera transfer functions (Mix et al., 1999), or a limitation in LGM simulations.

The 21 ka simulations capture the general range of mean SST estimates from the Multiproxy Approach for the Reconstruction of the Glacial Ocean Surface (MARGO) proxy data in both the Pacific and Atlantic basins (Figure S4a and S4b; Waelbroeck et al., 2009).

At 14 ka, cold SST anomalies continue throughout the global oceans, again with particular cooling in the North Pacific (Figure S5a). The 14 ka basin-wide transects of SST show a similar pattern to 21 ka, with only slightly negative anomalies throughout most of the southern and tropical regions of the basins and larger cooling north of 40 ^{o}N (Figures S4c and S4d). Unfortunately, a SST compilation analogous to MARGO is not available for 14 ka, but the strong cooling in the North Pacific continues to suggest this region's importance, despite a limited number of SST records from this region. Three records from off the coast of Japan show mean 14 ka SST cooling of 1-3 ^{o}C (Sawada & Handa, 1998; Sun et al., 2005; Yamamoto et al., 2005). Our 14 ka simulations are consistent with cooling of 2.1±0.2 ^{o}C (14ka-L) and 2.0±0.2 ^{o}C (14ka-5G) in the same region. Model SST is too cold in the Sea of Okhotsk but within the uncertainty of proxy measurements, where alkenone records suggest SST cooling of 0-4 ^{o}C (Harada et al., 2006; Seki et al., 2007), compared with 4.3±0.2 ^{o}C (14ka-5G) and 4.7±0.2 ^{o}C (14ka-L). Additionally, one alkenone record from the southeastern Bering Sea suggests a SST cooling of 1-4 ^{o}C (Caissie et al., 2010; Dubois et al., 2009), while the 14 ka simulations indicate SST cooling of 5.2±0.5 ^{o}C (14ka-5G) and 5.6±0.4 ^{o}C (14ka-L).

S2.2 Sea surface salinity

Globally averaged sea surface salinity (SSS) anomalies at 21 ka (+0.55±0.01 psu for 21ka-L; +0.60±0.01 psu for 21ka-5G) reflect a generally more saline ocean related to the reduction in global ocean volume. However, there are prominent regional distinctions in SSS anomalies (Figure S3b), such as the increase in SSS across the Arctic Ocean that is related to the increase in sea ice formation and brine rejection, as shown in previous LGM simulations (Otto-Bliesner et al., 2006). The freshening in the Gulf of Mexico and along the Scandinavian margin reflects the increased contribution of glacial runoff, which is consistent with SSS reconstructions (de Vernal et al., 2000; Flower et al., 2004). In addition, prominent freshening in the Sea of Japan is also consistent with reconstructions (Keigwin and Gorbarenko, 1992; Tada et al., 1999; Gorbarenko and Southon, 2000) and reflects the elevated P-E anomalies (Figure S1b), as well as the isolation of the basin from the open ocean due to sea-level lowering. The reduction in 21 ka SSS in the South China and Arafura Seas is also linked to P-E anomalies in the 21 ka simulations (Figure S1b). SSS anomalies from the South China Sea have been associated with greater proximity to river outlets (Steinke et al., 2006). This may confirm that P-E anomalies are too high in this region from an elevated land-sea temperature contrast in the model (see section 3.1.2).

Average SSS anomalies at 14 ka are +0.17±0.01 psu (14ka-L) and +0.19±0.01 psu (14ka-5G), continuing to suggest a more saline surface ocean related to volumetric changes in the global ocean. The Arctic Ocean is now dominated by the freshening of surface waters relative to 0 ka (Figure S5b), as the Northern Hemisphere ice sheets supply a greater amount of meltwater and river routing to the region (not shown). In addition, the Gulf of Mexico and eastern North Atlantic reflects continued meltwater runoff from the southern LIS as inferred from proxies from the region (Flower et al., 2004) as well as sustained freshening in the Nordic Seas (de Vernal and Hillaire-Marcell, 2000). The Sea of Japan continues to have prominent fresh SSS anomalies due to enhanced P-E and limited open ocean exchange, contrary to observations (Gorbarenko and Southon, 2000).

S2.3 δ^{18}O of the surface ocean

21 ka δ^{18}Oo anomalies largely reflect the trends in SSS, with nearly global enrichment in the open ocean due to sea-level lowering from increased LGM ice-sheet volume. More localized effects include δ^{18}Oo enrichment in the Arctic Ocean concurrent with elevated sea ice formation and δ^{18}Oo depletion in the Gulf of Mexico and along the Scandinavian Ice Sheet margin where there is an increase in meltwater runoff relative to 0 ka (Figure S3c). The 21 ka δ^{18}Oo anomalies also show depleted values in the Sea of Japan.

We compare the simulated change in δ^{18}Oo relative to 0 ka using marine sediment records of foraminifera that have an independent temperature reconstruction from the same core to correct for glacial changes in SST (i.e. δ^{18}O of seawater, see section 2.5). The 21 ka simulations capture the general change in δ^{18}Oo shown from the proxy data (Figure S3c). One noticeable offset is in the Gulf of Mexico where the Ziegler et al. (2008) record in the Gulf of Florida sub-basin is not significantly influenced by LIS runoff, whereas ModelE2-R simulates a significant impact of LIS runoff consistent with other Gulf of Mexico records (Flower et al., 2004). Unfortunately, the limited number of temperature-corrected δ^{18}Oo records that meet our selection criteria (providing data from the LGM as well as the past 2000 years) are primarily confined to the tropics. We therefore include these data with our δ^{18}Oa records for a global comparison of oxygen δ^{18}O at the LGM. Despite the limitations in spatial resolution of this dataset, the 21ka-L and 21ka-5G simulations correlate moderately well with the total δ^{18}O proxy dataset, with r=0.53 and r=0.62 respectively ($p \leq 0.01$ for both).

The 14 ka δ^{18}Oo anomalies also reflect the general trends in SSS (Figure S5b and S5c). The freshening of the Arctic Ocean from glacial meltwater runoff leads to depleted δ^{18}Oo values relative to 0ka. The 14 ka simulations capture the general trends in the tropics-dominated proxy dataset, and the inclusion of these data into a total 14 ka δ^{18}O comparison result in correlations for 14ka-L and 14ka-5G of r=0.41 and r=0.41, respectively ($p \leq 0.01$).

S2.4 Sea ice

There is expansion of annually averaged sea ice in the Nordic and Labrador Seas in the 21 ka simulations relative to the control (Figure S3d), and this extent is largely the same as the maximum wintertime extent (not shown). In the Nordic Sea, the sea ice increases are primarily confined to north of Iceland. This sea ice extent is largely consistent with proxy inferences of wintertime sea ice extent from this region (Sarnthein et al., 2003), but we do not capture the summer sea ice pullback of this reconstruction. In the Labrador Sea, ice extends from just south of Hudson Strait to the southwest coast of Greenland, which is consistent with perennial sea-ice reconstructions (de Vernal and Hillaire-Marcell, 2000; de Vernal et al., 2000) and previous LGM simulations (Otto-Bliesner et al., 2006; 2007; Braconnot et al., 2007a). The other region of major sea ice expansion in the 21 ka simulations occurs in the Sea of Okhotsk, and the Bering Sea (Figure S3d), also consistent with reconstructions (Shiga and Koizumi, 2000; Katsuki et al., 2010; Caissie et al., 2010).

The 14 ka simulations also have expansion of sea ice into the Nordic Sea, Sea of Okhotsk and Bering Sea (Figure S5d). Unfortunately there is no basin-wide reconstruction of 14 ka sea ice in the North Atlantic, although it was largely reduced to the Fram Strait by the start of the Holocene (de Vernal and Hillaire-Marcell, 2008). Reconstructions from the Sea of Okhotsk suggest persistent sea ice in the region until after 6.5 ka, but the Bering Sea likely began to transition away from perennial sea ice after 17 ka (Caissie et al., 2010). However, summer sea ice extent in the Bering Sea in the 14 ka simulations is largely the same as the 0 ka simulation (not shown), consistent with this loss of perennial sea ice.

S2.5 Ocean circulation

North Atlantic Deep Water (NADW) production is enhanced in the 21 ka simulations at the mid-latitudes (up to 50° N) but diminished at higher latitudes (50°-65° N) due to a southward shift in convection sites (Figure S3e). This is associated with a deepening and strengthening of overall mean AMOC transport to 30.8±0.6 Sv (21ka-L) and 33.2±0.7 Sv (21ka-5G), relative to 28.2±0.7 in the control simulation. Below the NADW, however, is an enhanced contribution of Antarctic Bottom Water (AABW) from the south (Figure S3e). In the Pacific, the 21 ka simulations show enhanced deepwater circulation from AABW and a reduction in North Pacific Intermediate Water (NPIW; Figure S3f).

The 14 ka simulations also have enhanced NADW formation with a southward shift in convection sites (Figure S5e). The location of maximum AMOC is approximately at the same depth as the control simulation, but overall mean transport remains elevated at 30.5±0.6 Sv (14ka-L) and 32.7±0.7 Sv (14ka-5G), relative to the control (28.2±0.7 Sv). Again, below the NADW, there is an enhanced contribution of AABW. In the Pacific, Antarctic Intermediate Water (AAIW) formation is slightly enhanced and NPIW is reduced relative to the control (Figure S5f).

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Table S1. $\delta^{18}O$ proxy records used in model-data comparison of $\delta^{18}Oa$ and $\delta^{18}Oo$. Records were selected that have continuous coverage from the LGM to preindustrial. For ocean records, an additional selection criteria required the existence an associated independent temperature proxy (i.e., Mg/Ca) to calculate $\delta^{18}O_{seawater}$ from $\delta^{18}O_{calcite}$ (Bemis et al., 1998).

Reference	Proxy	Lat (° N)	Lon (° E)
$\delta^{18}Oa$ records			
Bar-Matthews et al., 2003	Speleothem	31.5	35.0
Cheng et al., 2012	Speleothem	42.9	81.8
Cruz et al., 2005	Speleothem	-27.2	-49.2
Dykoski et al., 2005	Speleothem	-27.2	-49.2
Holmgren et al., 2003	Speleothem	-24.0	29.2
Partin et al., 2007	Speleothem	4.0	114.0
Wang et al., 2001	Speleothem	25.3	108.1
Wang et al., 2007	Speleothem	-27.2	-49.2
Williams et al., 2005	Speleothem	-42.0	172.0
Dansgaard et al., 1993	GRIP ice core	72.6	-37.6
Grootes et al., 1993	GISP2 ice core	72.6	-38.5
Svensson et al., 2008	NGRIP ice core	75.1	-42.3
Thompson et al., 1998	Bolivia ice core	-18.0	-69.0
Kohn and McKay, 2010	Megafauna teeth (LGM only)	45.0	-108.0
$\delta^{18}Oo$ records			
Benway et al., 2006	Marine sediment core	7.9	-83.6
Carlson et al., 2008b	Marine sediment core	33.7	-57.6
Carlson et al., 2008b	Marine sediment core	32.8	-76.3
Carlson et al., 2008b	Marine sediment core	-27.5	-46.5
Chen et al., 2010	Marine sediment core	26.6	125.8
Govil and Naidu, 2010	Marine sediment core	14.5	72.7
Klinkhammer et al., 2009	Marine sediment core	7.9	83.6

Koutavas et al., 2002	Marine sediment core	-1.2	-89.7
Lea et al., 2000	Marine sediment core	2.3	-91.0
Lea et al., 2003	Marine sediment core	10.7	-64.9
Lea et al., 2006	Marine sediment core	0.5	-92.4
Mohtadi et al., 2010	Marine sediment core	-1.5	100.1
Mohtadi et al., 2010	Marine sediment core	-5.9	103.2
Oppo and Sun, 2005	Marine sediment core	19.6	117.6
Pahnke et al., 2003	Marine sediment core	-45.5	174.9
Sagawa et al., 2006	Marine sediment core	36.1	141.8
Skinner and Shackleton, 2004	Marine sediment core	37.8	-10.2
Steinke et al., 2006	Marine sediment core	6.6	113.4
Steinke et al., 2008	Marine sediment core	6.6	113.4
Steinke et al., 2011	Marine sediment core	19.5	116.3
Stott et al., 2007	Marine sediment core	-10.6	125.4
Stott et al., 2007	Marine sediment core	-5.0	133.4
Stott et al., 2007	Marine sediment core	6.3	125.8
Visser et al., 2003	Marine sediment core	-4.7	117.9
Weldeab et al., 2005	Marine sediment core	2.5	9.4
Weldeab et al., 2006	Marine sediment core	-4.6	-36.6
Weldeab et al., 2007	Marine sediment core	2.5	9.4
Ziegler et al., 2008	Marine sediment core	29.0	-87.1
Xu et al. 2008	Marine sediment core	-13.1	121.8

Figure S1. 21 ka anomalies (21ka-L minus 0 ka), for the following annually averaged atmospheric variables: (a) Surface Air Temperature (^{o}C); (b) Precipiation minus Evaporation (mm/day); (c) δ^{18}Oa (‰) with anomalies from proxy records (see Table S1) plotted as circles with the same colorbar; (d) Surface Wind Speed (m/s); (e) Atmospheric Jet Speed (definied as the wind speed at 250 hPa; m/s); (f) Geopotential height at the 500 mb level, with zonal mean removed (meters); (g) Ground Albedo (%); (h) Planetary Albedo (%). Ice sheet extents outlined in bold black line.

Figure S2. Same as Figure S1, but for 14 ka anomalies (14ka-L minus 0ka).

Figure S3. 21 ka anomalies (21ka-L minus 0 ka), for the following annually averaged ocean variables: (a) Sea Surface Temperature (^{o}C); (b) Sea Surface Salinity (psu); (c) δ^{18}Oo (‰) with anomalies from proxy records (see Table S1) plotted as circles with the same colorbar; (d) Sea Ice Fraction (%); (e) Atlantic Ocean overturning stream function (Sv); (f) Pacific Ocean overturning stream function (Sv).

Figure S4. Longitudinal transects of SST anomalies averaged across the Atlantic and Pacific basins for the 21 ka simulations, with comparison to published MARGO data and uncertainties (Waelbroeck et al., 2009). MARGO data is separated by categories of proxies used in the reconstruction of SST as follows: blue x's, dinoflagellates; black squares, foraminifera; red diamonds, Mg/Ca; and green circles, alkenones ($U^{K'}{}_{37}$). A similar compilation of SST records does not exist for 14 ka.

Figure S5. Same as Figure S3, but for 14 ka anomalies (14ka-L minus 0ka).