Ocean circulation exerts a fundamental control on heat transport, marine biological productivity, and ocean-atmosphere CO₂ exchange (1). Today, the North Atlantic and North Pacific have contrasting circulation regimes, with vigorous deep-water formation in the North Atlantic, but only limited local ventilation in the North Pacific (2–4). This drives distinct patterns of CO₂ exchange, nutrient supply, and heat transport in each basin (5) and represents a first-order asymmetry in global biogeochemistry and climate (Fig. 1). It has been proposed that the circulation of the North Pacific was markedly different in past cold climates, but no consensus on this change exists. Proxies for ocean ventilation suggest enhanced intermediate water formation under glacial conditions (6–11) and local deep-water formation during Heinrich Stadal 1 (12, 13). However, paired nutrient utilization and productivity proxies indicate reduced nutrient supply to the surface during cold climates (14, 15), interpreted as the result of more stratified conditions with reduced ventilation. As the Pacific contains around half of the water in the global ocean and 30 times more carbon than the atmosphere, the lack of consensus on its behavior in different climate states represents a major gap in our understanding of the global climate system. Understanding the glacial climate of this region is also of interest for understanding the conditions under which early humans first migrated from Asia to North America around the end of the last ice age (16). Here, we use a compilation of sediment core proxy data for ocean ventilation, biological productivity, temperature, and salinity (see the Supplementary Materials for details), paired with Earth system modeling experiments, to shed new light on the circulation of the glacial North Pacific and its regional and global impact.

OCEANOGRAPHY

Overturning circulation, nutrient limitation, and warming in the Glacial North Pacific

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Although the Pacific Ocean is a major reservoir of heat and CO₂, and thus an important component of the global climate system, its circulation under different climatic conditions is poorly understood. Here, we present evidence that during the Last Glacial Maximum (LGM), the North Pacific was better ventilated at intermediate depths and had surface waters with lower nutrients, higher salinity, and warmer temperatures compared to today. Modeling shows that this pattern is well explained by enhanced Pacific meridional overturning circulation (PMOC), which brings warm, salty, and nutrient-poor subtropical waters to high latitudes. Enhanced PMOC at the LGM would have lowered atmospheric CO₂— in part through synergy with the Southern Ocean—and supported an equable regional climate, which may have aided human habitability in Beringia, and migration from Asia to North America.

INTRODUCTION

RESULTS AND DISCUSSION

Data compilation

To examine past changes in ventilation, we use proxies including δ¹³C in benthic foraminifera (6, 17), ¹⁴C offsets between benthic

Fig. 1. Modern hydrography and nutrient content of the northern Pacific and Atlantic Oceans. (A) Sea surface salinity, with sea surface temperature contours for 5° and 10°C. (B) Surface phosphate concentration, with a sea surface height contour to denote the subpolar gyre boundary. (C) Subsurface phosphate concentration along a zonal section indicated by the dashed line in (B), with pre-bomb radiocarbon age contours. Note the relative isolation of the subpolar gyre in the North Pacific, which allows pooling of cold, fresh water in the surface, and old, nutrient-rich waters to upwell from below. In contrast, the active overturning circulation of the North Atlantic flushes warm, salty, nutrient-poor water from the subtropics through the upper reaches of this well-ventilated basin. Salinity, temperature, and phosphate data are gridded annual averages from WOA09; radiocarbon data from Global Ocean Data Analysis Project version 2 [GLODAPv2 (101)]; and sea surface height from (159).
and planktic foraminifera (18, 19), and various redox tracers (7). These data indicate that waters below ~2000 m are poorly ventilated at the Last Glacial Maximum (LGM), with lower δ^{13}C, δ^{14}C, and oxygen (Figs. 2A and 3). This is consistent with previous interpretations of the deep glacial Pacific as a likely store of glacial CO2 (1, 20). In contrast, waters above ~2000 m show enhanced ventilation (Figs. 2A and 3). This is seen across all tracers, with 73% of the 72 sites above 2000 m indicating increased ventilation, 17% showing no change outside of 1σ uncertainty, and only 10% indicating reduced ventilation. This suggests a substantial increase in North Pacific Intermediate Water (NPIW) formation at the LGM (6). This result is further supported by profiles of δ^{18}O in benthic foraminifera (6, 11): In contrast to the modern, where a smooth and modest increase in δ^{18}O with depth suggests a dominantly diffusive regime, LGM profiles (6, 11) exhibit a notable transition at ~2000-m water depth (Fig. 3A). This indicates a sharp transition in temperature and/or δ^{18}O_{sw}, which, as conservative tracers, provide evidence of a physical water mass boundary (21), with increased advection of NPIW overcoming the smoothing influence of diffusion. A local source of intermediate waters in the glacial North Pacific is also supported by meridional sections of benthic δ^{14}C age and δ^{15}C (Fig. 3 and fig. S3), which show the spread of well-ventilated waters from the surface into the ocean’s interior.

To examine changes in sea surface temperatures (SSTs), we use Mg/Ca in planktic foraminifera and the alkenone saturation index UK′_{37}. SSTs in the subtropical gyre are cooler at the LGM, as expected for glacial conditions, and there is a large midlatitude cooling due to the southward expansion of the subpolar gyre (Fig. 2C and fig. S4) (22). In contrast, LGM temperatures in the subpolar gyre are either similar to or warmer than today (80% of records show warmer temperatures or no significant change), with ~2°C warming observed in the western North Pacific (Fig. 2C and figs. S1C and S4) (23, 24). These warm SSTs require a circulation change to counter lower CO2 and greater ice coverage (25) at the LGM and are consistent with enhanced heat transport by an invigorated Pacific Meridional Overturning Circulation (PMOC).

Sea surface salinity change at the LGM is estimated using δ^{18}O in planktic foraminifera, corrected for temperature using Mg/Ca and for whole-ocean δ^{18}O/salinity change. Increased salinity is indicated throughout the North Pacific subpolar gyre (Fig. 2D and fig. S4), which would have aided intermediate water formation.

To reconstruct changes in export productivity, we use sedimentary opal and biogenic barium contents (26). All data poleward of 40°N indicate a reduction in glacial export productivity (Fig. 2B and fig. S1B). However, tracers of nutrient utilization (e.g., δ^{15}N) suggest that biological nutrient drawdown was more complete (15). For more complete nutrient drawdown to be achieved despite reduced export productivity, the pool of available major nutrients (phosphate, nitrate, and silicic acid) must have been smaller (1, 15). This is further supported by considering that micronutrient (e.g., iron) availability, which partially limits productivity in the North Pacific today.

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**Fig. 2. Reconstructions of changes in ventilation, export productivity, temperature, and salinity at the LGM relative to the Holocene.** Red indicates an increase at the LGM relative to the Holocene, blue a decrease, and white no clear change outside of 1σ uncertainty (see the Supplementary Materials for details of data compilation). (A) Ventilation proxies, including δ^{13}C (circles), radiocarbon (diamonds), and a variety of redox tracers (squares), taken from sites above 20°N. (B) Export productivity proxies, including sediment core opal (circles) and biogenic barium (squares) contents. (C) Sea surface temperature proxies, including planktic foraminiferal Mg/Ca (circles) and alkenone saturation index UK′_{37} (diamonds). (D) Change in sea surface salinity, derived from paired δ^{18}O and Mg/Ca data on planktic foraminifera, corrected for whole-ocean δ^{18}O and salinity changes due to ice volume. At the LGM, North Pacific intermediate waters are better ventilated, and productivity in the subpolar North Pacific is lower. An increase in salinity is seen throughout the North Pacific, and although the subtropical gyre cools, the subpolar North Pacific shows an anomalous warming, despite peak glacial conditions. See fig. S1 for an alternative presentation of these data.
was enhanced because of higher dust fluxes (27). The decrease in export productivity in the LGM North Pacific thus requires a decrease in major nutrient supply.

Mechanisms for glacial nutrient limitation

Previous studies have suggested that nutrient limitation in the glacial North Pacific was driven by enhanced freshwater stratification in surface waters under a colder climate (14, 15). However, this suggestion is difficult to reconcile with our evidence for enhanced intermediate water formation (Figs. 2A and 3) and saltier, warmer surface waters (Fig. 2, C and D, and fig. S4). An alternative model is thus required for large-scale changes in nutrients and circulation in the northern subpolar basins.

To explore controls on circulation and nutrient supply in the North Pacific, we first consider modern phosphate concentrations in an adaptation of Warren’s box model [(2, 3); the Supplementary Materials], which was previously used to demonstrate that the low salinity that currently stratifies the subpolar North Pacific is a result of low evaporation rates relative to precipitation, as well as minimal exchange with the saltier subtropical gyre (3–4). Isolation of the subpolar gyre also contributes to high surface nutrients in the modern North Pacific, by limiting the input of low-nutrient water from the subtropics (fig. S5). Although salinity stratification limits the depth of convection, subsurface waters are still brought to the surface via wind and tidally driven mixing, wintertime convection, and Ekman suction (5). As these subsurface waters have been long isolated in the deep ocean, they are extremely rich in nutrients acquired through remineralization, with little input of low-nutrient subtropical water or newly ventilated intermediate water to dilute this potent subsurface nutrient reservoir (Fig. 1C). This contrasts with the North Atlantic, where a less zonal wind-stress pattern and an active overturning circulation supplies warm, salty, and nutrient-poor subtropical water to high latitudes (Fig. 1). Subduction of this water creates nutrient-poor intermediate and deep waters and effectively flushes nutrients from the upper reaches of the modern North Atlantic.
Thus, today we have two realizations of basin-scale nutrient dynamics and circulation at high northern latitudes: The Pacific is stratified and poorly ventilated, with cold, fresh, and nutrient-rich surface waters, supplied by upwelling of old, nutrient-rich waters from below, while the Atlantic is well ventilated, with a vigorous overturning circulation that flushes warm, salty, and nutrient-poor waters from the sub tropics through its upper reaches. At the LGM, our data show that the North Pacific was better ventilated, had lower nutrients, and was saltier and warmer. This is consistent with a more Atlantic-like circulation regime, with enhanced overturning down to \( \sim 2000 \)-m depth.

**Modeling experiments**

The ability of enhanced overturning to reduce nutrient supply to the surface of the glacial North Pacific is further supported by a series of experiments with an Earth system model (cGENIE) and tests with a simple box model (see the Supplementary Materials). We simulate Pacific overturning in cGENIE by reducing the prescribed transport of fresh water from the North Atlantic to the North Pacific. As the North Pacific gets saltier, its meridional overturning circulation increases (fig. S6), driving a decrease in surface and subsurface nutrients (Fig. 4 and figs. S5 and S7) due to increasing input of nutrient-poor subtropical waters. Although convective mixing increases and stratification is reduced (Fig. 4C), the dilution of the subsurface nutrient reservoir by freshly ventilated waters has a much larger effect (Fig. 4 and figs. S5 and S7; see Materials and Methods). This pattern is observed across different model base states and is also seen in the higher-resolution LOVECLIM and UVic Earth system models (fig. S8) (28). Recent work using the Paleoclimate Modelling Intercomparison Project (PMIP3) ensemble shows that Ekman suction in the glacial North Pacific was around 60% higher than today (29). Given this increase in upwelling, a substantial decrease in the nutrient content of the upwelled water is required to result in lower surface nutrient concentrations; our modeling results show this can be effectively achieved by enhanced overturning circulation.

**Pacific overturning and biogeochemistry during other climate states**

The presence or absence of a PMOC may help explain biogeochemical transitions in other time intervals. For example, the peak in nutrient supply during the Bølling-Allerød warm interval [14.7 to 12.9 thousand years ago (ka)] appears to be associated with an abrupt decrease in PMOC (12, 29), while the decrease in export productivity and increase in SST during intensification of Northern Hemisphere

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**Fig. 4.** cGENIE Earth system model experiments illustrating the impact of changes in North Pacific overturning on nutrient concentrations. Each symbol (A to C) represents a 5000-year-long model experiment under preindustrial (orange triangles) or glacial (blue diamonds) boundary conditions. Experiments are forced by decreasing the atmospheric freshwater flux into the North Pacific, which increases surface salinity and enhances PMOC (fig. S6). Output is plotted against the maximum meridional stream function in the North Pacific below 280 m. Phosphate concentrations are shown for (A) the surface layer (top 80 m) and (B) intermediate depths (928 to 1158 m) in the subpolar gyre (grid cells spanning 43°N to 66°N and 115°W to 225°W). Convection frequency (C), a count of convection events throughout the water column per model time step, is shown for the same region. Meridional sections of phosphate concentration (shading) at 165°W and Pacific meridional overturning stream function (contours) are shown in (D), for an experiment with a salinity forcing of \(-0.19\) Sv and a PMOC of 8 Sv (indicated by a dashed line in (A) to (C)) that produces the best fit to our LGM data (Fig. 5 and fig. S10); anomalies of phosphate and overturning from the glacial base state are shown in (E); and the spread of North Pacific water at 1000-m depth, predominantly on the basin's western boundary, is shown in (F).
glaciation ~2.7 million years ago (14) could be explained by increased PMOC, analogous to the change from modern to LGM conditions. We note that, as a transient condition, the onset of deep-water formation can lead to a short-lived peak in surface nutrients and CO₂, as convection initially taps into high-nutrient subsurface waters before they have been replaced with well-ventilated low-nutrient waters (fig. S9); this remains a likely explanation for North Pacific pH, δ¹³C, and ¹⁴C excursions (12) during Heinrich Stadial 1 (HS1; ~17.5 to 16.0 ka). However, the typical pattern under steady-state conditions—both in paleo records and our modeling—appears to be that increased overturning drives a decrease in nutrient supply (Fig. 4).

The dynamics of a glacial PMOC

Increased ventilation of the North Pacific at the LGM would be facilitated by an increase in the salinity of subpolar surface waters (2, 3), as implied by the available data (Fig. 2D and fig. S4). Higher salinity in the glacial North Pacific may be driven by reduced net precipitation (P−E), due to a weakened hydrological cycle under globally cooler conditions (30), and a southward shift in storm track precipitation (31). Increased Ekman suction in the presence of the Laurentide ice sheet (29) would also bring up salt from subsurface waters. Salinity may be further elevated by increased exchange with the saltier subtropical gyre, due to an intensification of the westerly winds in response to the Laurentide ice sheet (22, 29), or as a Stommel feedback to an initial increase in overturning (32). Although an increase in SST would, in isolation, decrease the density of surface waters, this may be overcome by an increase in salinity, in part resulting from enhanced evaporation and divergent moisture transport (2); the combination of increased SST and increased North Pacific ventilation is indeed found in ocean circulation models (figs. S7 and S11). Last, it is possible that seasonal sea ice formation and brine rejection play a role (8, 25), in particular, in helping to set the local sites of downwelling, which are likely to be in the marginal seas (10). Enhanced sea ice production may be reconciled with warmer SSTs either by a difference in seasonality or if sea ice production occurs predominantly in coastal settings (8). Although basin geometry and a closed Bering Strait may, in part, limit salinity in the North Pacific (4), our proxy reconstructions show that local ventilation to intermediate depths is indeed possible, a result also found in the higher-resolution NCAR CCSM3 and GFDL CM2Mc models under some instances of colder climate conditions (figs. S11 and S12).

Comparing reconstructed δ¹³C profiles and other ventilation data with cGENIE output (Fig. 5 and fig. S10) suggests a PMOC of around 8 Sv at the LGM, which is also consistent with the changes seen in SST (fig. S7). Note, however, that the PMOC cell remains largely above 2000 m and within the Pacific basin in our cGENIE experiments (fig. S7), in contrast with Atlantic Meridional Overturning Circulation (AMOC) at similar overturning rates. δ¹³C values in the intermediate North Pacific also remain lower than those in the glacial North Atlantic (fig. S13). These observations may, in part, be explained by the much greater volume of the Pacific, which increases the residence time of northern-sourced waters within the basin, and allows remineralization and mixing with southern-sourced waters to exert greater influence. While glacial AMOC may have remained higher than PMOC, the LGM’s overturning circulation, with northern-fed cells in both basins to ~2000 m overlying southern-sourced waters, was substantially more symmetrical than today. It is possible that this more symmetrical glacial circulation is the most common configuration of the overturning circulation during the late Pleistocene, with today’s asymmetric mode limited to peak interglacial conditions.

Impact of PMOC on CO₂

Enhanced PMOC would have increased the North Pacific’s influence on the glacial carbon cycle and climate (Fig. 6). Today, the North Pacific is a High Nutrient Low Chlorophyll zone, with an
Fig. 6. Schematic of ocean circulation and productivity in the modern and LGM North Pacific. The modern North Pacific (left) lacks vigorous local ventilation (thin black arrows), because of low salinity in surface waters of the subpolar gyre (SPG). This is a result of high net precipitation (P-E) and minimal exchange with the saltier waters of the subtropical gyre (STG). The modern subpolar North Pacific is thus dominated by upwelling of nutrient- and CO₂-rich subsurface waters (gray vertical arrow), driving relatively high export productivity (green wavy arrows) and CO₂ outgassing. During the LGM (right), our data compilation suggests that ventilation at intermediate depths was enhanced, export productivity was reduced, and subpolar surface waters were saltier and warmer (red wavy arrows). This is consistent with an invigorated meridional overturning circulation, with enhanced formation of intermediate waters and advection of warm, salty, and nutrient-depleted subtropical waters to high latitudes, analogous to a shallower version of the overturning circulation seen in the modern North Atlantic.

inefficient biological pump and high CO₂ outgassing. By reducing the carbon content of the high-latitude North Pacific, an invigorated PMOC would have helped stem this CO₂ leak, as indicated by boron isotope data, which show LGM surface ocean CO₂ close to equilibrium with the atmosphere (29). Furthermore, these changes in the surface ocean would have had a greater impact on the net ocean CO₂ storage, because enhanced intermediate water formation would lower the preformed nutrient content of the ocean's interior. Combining our cGENIE results with a simple scaling between preformed phosphate and atmospheric CO₂ (33) shows that atmospheric CO₂ is relatively sensitive to modest increases in PMOC and suggests CO₂ drawdown on the order of ~16 parts per million (ppm) (see Materials and Methods and fig. S14). CO₂ drawdown may be further enhanced by PMOC's impact on biogeochemistry in the Southern Ocean (34). Today, carbon-rich waters recirculated from the Pacific upwell in the Southern Ocean and contribute to high surface nutrients and CO₂ outgassing in this region. Replacing a portion of this upwelling water with well-ventilated glacial NPIW reduces the carbon and nutrient supply to the Southern Ocean surface (see Earth system model results in fig. S8), thus increasing the efficiency of the Southern Ocean biological pump and further lowering atmospheric CO₂. This process may help explain the similarity in nutrient utilization records from the North Pacific and Southern Oceans [the "polar twins" (35)]. If enhanced glacial PMOC results from increased sea ice and ice sheet size, then this would provide a feedback between Northern Hemisphere insolation and atmospheric CO₂, linking the orbital pacing of the ice ages to global glacial climate change.

Regional warming in Beringia due to enhanced PMOC

The relatively mild regional climate supported by heat transport from an enhanced PMOC may have implications for the first human migration to the Americas. Recent genetic studies suggest that the founding population of early Native North and South Americans were isolated from other, future Asian populations during the LGM, before dispersing into and through the American continents (16). This isolation is thought to have been located in northeastern Siberia or on the now submerged Bering Sea shelf, lasting between 7000 and 15,000 years (36) during what is known as the “Beringian standstill.” Some terrestrial records support the interpretation that this region was an ecological refugium, with spruce, birch, and alder pollen and fossil insects providing evidence of a relatively mild and maritime climate (37). However, given the extremely cold conditions further inland (38), it has remained unclear how habitable conditions could have been maintained. Our data and modeling suggest that an enhanced overturning circulation helped warm this region at the LGM, countering the influence of lower greenhouse gasses and higher albedo, to bring Pacific Beringian temperatures up to, or even warmer than, modern (Fig. 2C and figs. S1C and S4). Warming may have been particularly enhanced during LGM winters, as seen in CCSM3 (fig. S11B), which shows that for a mean annual temperature change of ~2°C in the glacial northwest Pacific (similar to our reconstructions), winter warming may have been around 6°C. An enhanced Pacific overturning circulation could have thus aided human habitation in Beringia during a period of refuge before migration to North America.

MATERIALS AND METHODS

Ventilation data

Sediment core proxy data for ventilation at the LGM includes δ¹³C data on benthic foraminifera, tracers of sedimentary redox, and radiocarbon age offsets between benthic and planktic foraminifera, taken from sites above 20°N in the North Pacific (with a version above 40°N shown in fig. S1A). Stable carbon isotope data (δ¹³C) (6, 9, 11, 39–53) on benthic foraminifera of the genus Cibicides are largely taken from the recent synthesis by Peterson et al. (17), supplemented with recent studies (9, 53). Where possible, we compare foraminiferal data for the LGM (19 to 23 ka) and late Holocene (0 to 6 ka). At sites where Holocene foraminiferal data are not available (24 of 62), we use seawater δ¹³C data taken from the PACIFIC ocean Interior Carbon (PACIFICA) database (52), with values interpolated for these core sites using the three-dimensional estimation tool in Ocean Data View [ODV, (54)], except at Keigwin’s (6) depth transect in the NW Pacific and Max’s (9, 53) sites in the Bering Sea, where we use the accompanying water column measurements. Both approaches show an increase in δ¹³C at intermediate depths. To account for whole-ocean change in δ¹³C at the LGM (attributed to a smaller terrestrial biosphere), and thus better reveal patterns of change in circulation, we correct LGM δ¹³C values by +0.34‰o (17). We then calculate the LGM minus Holocene δ¹³C difference, with positive values indicating higher δ¹³C and increased ventilation at the LGM. We use a threshold value of ±0.1‰ for a significant increase/decrease in δ¹³C when comparing LGM and Holocene foraminiferal data, based on 1 SD of replicate samples of benthic foraminifera from the NW Pacific (6). When comparing LGM foram δ¹³C to modern water δ¹³C, we use a threshold value of 0.2‰o, to account for calibration error [e.g., (42)] and the sparseness of water column δ¹³C data in this region.

Redox reconstructions (55–79) follow the approach and compilation of Jaccard and Galbraith (7), who estimated whether oxygen
concentrations increased or decreased at the LGM (defined in their study to be 20 to 22 ka) compared to Holocene (5 to 10 ka). Redox proxies include trace metals, sediment laminations, assemblages of benthic foraminifera, and nitrogen isotopes from regions of modern denitrification. Following Jaccard and Galbraith (7), redox changes based on nitrogen isotopes are plotted at 400 m to reflect the depth of water column denitrification. We use the published thresholds for significant change in redox, quoted as approximately 1 SD.

Radiocarbon data (12, 20, 53, 70, 76, 80–100) are taken largely from syntheses by (13, 18–20). Glacial data shown in Fig. 2A consist of $^{14}$C age offsets between benthic and planktic foraminifera from the same sample and include data with calendar ages in the window 18 to 23 ka. Holocene benthic–planktic $^{14}$C data are rare in this region because of poor carbonate preservation, so we compare LGM benthic–planktic offsets to deep-surface offsets in water column $^{14}$C. These are calculated from nearby deep-water $^{14}$C data from GLODAPv2 (101) and a surface water age of 400 $^{14}$Cyr south of the subpolar gyre boundary (43°N) and 700 $^{14}$Cyr within the subpolar gyre. The threshold for a significant change at the LGM is the quadratic combination of $^{14}$Cyr 1-SE measurement uncertainties on LGM benthic and planktic $^{14}$C data and a 200 $^{14}$Cyr uncertainty on modern deep-surface offsets, and averages 289 $^{14}$Cyr. The section view in Fig. 3C uses the datasets listed above with benthic data shown as age offsets from the contemporaneous atmosphere (102). Note that despite their varying influences and sources of uncertainty, each set of ventilation proxy data shows the same patterns of change at the LGM, with enhanced ventilation at intermediate depths and reduced ventilation in the abyss.

Productivity data
Biogenic opal and barium data (26, 59, 74, 76, 87, 103–118) were taken from the compilation of Kohfeld and Chase (26) and supplemented with recent data from (12, 15, 87, 119–122). LGM and Holocene mean values and SEs were calculated from data in the time windows 0 to 10 ka (Holocene) and 19 to 23 ka (LGM). A significant change is noted when the difference between the LGM and Holocene means exceeds the quadratic sum of their 1-SE values. Sedimentaryopal (percent) and barium (ppm) content are shown in Fig. 2B. Where available, published mass accumulation rate estimates of these biogenic components are shown in fig. S1B, largely based on sediment core age models (from $\delta^{18}$O and $^{14}$C) and dry bulk density (either measured or estimated); fluxes based on $^{230}$Th are indicated with bold symbol outlines (fig. S1B). Each of these treatments of the data indicates reduced export productivity in the LGM subpolar North Pacific.

Temperature and salinity data
Published records of planktic foraminiferal Mg/Ca and $\delta^{18}$O calcite (24, 29, 82, 87, 123–127), $^{18}$Osw ($\delta^{14}$O, $^{14}$C, and dry bulk density) (either measured or estimated); fluxes based on $^{230}$Th are indicated with bold symbol outlines (fig. S1B). Each of these treatments of the data indicates reduced export productivity in the LGM subpolar North Pacific.

LGM warming is again likely to be conservative. An increase/decrease is indicated by a change of $\pm 1^\circ C$ for Mg/Ca, $\pm 1.2^\circ C$ for $^{13}C_{\text{UP}}$, and $\pm 0.25\%$ for $^{18}O_{\text{sw-ivc}}$. These thresholds correspond to the uncertainties associated with the temperature proxies (136, 141, 142) and the equivalent in $^{18}O$ (144).

**Time windows**

Our data compilations take advantage of available syntheses of data from the North Pacific where available, supplemented with recently published data. Time slices selected for the LGM in these syntheses, as detailed above, are fairly consistent. Holocene time slices show more variety although, in each case, aim to most accurately represent postglacial Holocene conditions. For $^{13}C$, this requires that the glacial-interglacial whole-ocean/atmosphere/biosphere change is complete, which was only achieved in the late Holocene. For some redox proxies, core-top values are liable to disturbance, so an early Holocene slice may be more appropriate. Late Holocene carbonate preservation in the North Pacific is poor, limiting Holocene proxy records based on planktic foraminifera. As a result, LGM benthic-planktic $^{14}C$ data are compared to modern values from the water column. Last, as noted above, the use of early Holocene data within our LGM to Holocene temperature comparison is likely to underestimate the magnitude of LGM warming compared to pre-industrial, given the warmth of the early Holocene (147), so the LGM warming signal seen in Fig. 2C and figs. S1C and S4 is likely to be conservative. We also provide a $^{13}C$ profile for a time slice ~3000 years older than the original LGM pick in fig. S2, which demonstrates that the enhanced ventilation signal is robust to age model uncertainties, and cannot be attributed to aliasing of an HSI signal.

**Box model**

We use an adaptation of Warren (2) and Emile-Geay et al.’s (3) salinity budget for the North Pacific subpolar gyre to explore controls on salinity and nutrients in this region. The model considers a mass balance for water in a subpolar gyre box, representing latitudes 43°N to 63°N, longitudes 140°E to 120°W, and water depth 0 to 200 m, with inputs from the subtropical gyre ($V_{\text{STG}}$), upwelling ($V_{\text{UP}}$), net precipitation ($P-E$), and riverine runoff ($R$). Salinity of the subpolar gyre ($S_{\text{SPG}}$) is calculated from the ratio of inputs of salt to inputs of water

$$S_{\text{SPG}} = \frac{S_{\text{UP}} V_{\text{UP}} + S_{\text{STG}} V_{\text{STG}}}{V_{\text{UP}} + V_{\text{STG}} + (P-E) + R}$$

and depends on the salinity of subtropical input ($S_{\text{STG}}$) and upwelling water ($S_{\text{UP}}$). For the derivation of this equation, see (2) and (3). We follow the same approach to model phosphate concentration in the subpolar gyre ($PO_{4}^{3-}_{\text{SPG}}$)

$$PO_{4}^{3-}_{\text{SPG}} = \frac{PO_{4}^{3-}_{\text{UP}} V_{\text{UP}} + PO_{4}^{3-}_{\text{STG}} V_{\text{STG}}}{V_{\text{UP}} + V_{\text{STG}} + (P-E) + R}$$

using the volume transport terms and box definitions as described above and substituting salinity for phosphate concentrations in the subtropics ($PO_{4}^{3-}_{\text{STG}}$) and in upwelled waters ($PO_{4}^{3-}_{\text{UP}}$). Note that this formulation considers only phosphate supply by ocean advection and does not attempt to model removal by productivity. As a result, our modeled phosphate concentrations are equivalent to winter values before significant drawdown in the spring bloom.

We use Emile-Geay’s (3) estimates of advective transport from the subtropical gyre and upwelling, Wills and Schneider’s (148) estimates of $P-E$ and $R$ (taken from ERA-Interim reanalysis), and salinity and phosphate concentrations for subtropical and upwelled waters from World Ocean Atlas, computed using the box averaging tool in Ocean Data View (149). Following Warren (2), we consider only advection and do not attempt to account for the influence of diffusion or eddy flux, which have large uncertainties for salinity and are poorly constrained for phosphate. We checked this assumption with an estimate of phosphate eddy flux, based on the eddy transport terms determined by Emile-Geay (3) for salinity and using regional phosphate gradients. This made little difference (<5%) to our results, as the resulting values were small, and the supply of phosphate by vertical eddy flux is countered by the influence of nutrient-poor water from horizontal eddy flux. We use Warren’s approach (2) to obtain subtropical and upwelling salinity and phosphate values with minimal influence from eddy exchange with water from the subpolar gyre, by excluding values immediately adjacent to the subpolar gyre. Our subtropical surface values use a box spanning 24°N to 40°N, 135°E to 115°W, and depths 0 to 200 m, and our upwelling values are taken from 2000 m, with the same latitude and longitude bounds as the surface subpolar box. All model input values are given in table S2.

The model accurately reproduces the salinity of the modern North Pacific subpolar gyre (33.0). The model gives phosphate in the subpolar gyre of 2.0 μmol/kg, slightly higher than annual average values of 1.8 μmol/kg in the modern North Pacific, but matching February (prespring bloom) values of 2.0 μmol/kg. The influence of possible glacial conditions is shown in fig. S5 and table S1. Subpolar salinity increases when (i) net precipitation is reduced, (ii) upwelling is increased, and (iii) exchange with the subtropics is enhanced, particularly when subtropical salinity is elevated. Phosphate concentrations in the subpolar gyre are reduced by exchange with the subtropics and by decreasing the phosphate concentrations of upwelled waters and are increased by increasing the upwelling flux. However, the increase in the upwelling flux has less impact on surface phosphate than the reduction of the phosphate concentration of the upwelled water; indeed, the net effect of doubling vertical exchange while halving subsurface nutrients is similar to that of halving subsurface nutrients alone (fig. S5). Because the wind- and tidally driven transfer of water from the subsurface to the surface North Pacific is high, vertical nutrient gradients are relatively low, so further increase in vertical exchange (e.g., with reduced stratification) has less influence than decreasing the nutrient content of the upwelled water, as a result further substantiated by Earth system modeling (see below).

**cGENIE Earth system model**

To explore the impact of enhanced North Pacific overturning on nutrients and tracers in more detail, we ran a series of sensitivity tests with the “muffin” release of the cGENIE Earth system model (150). cGENIE consists of a three-dimensional frictional geostrophic ocean circulation model, two-dimensional sea ice and energy-moisture balance atmospheric models and incorporates representations of the marine geochemical cycling of carbon and other biologically mediated tracers (151). We use cGENIE with a 36 × 36 equal-area horizontal resolution and 16 vertical levels, with highest vertical resolution toward the ocean surface. Ocean physics and carbon cycle configuration are as described and evaluated in (150), with the addition of iron colimitation on marine productivity (152). Despite its coarse resolution, cGENIE successfully reproduces...
many of the large-scale features of marine biogeochemical cycles in the present-day (150, 151). Of particular relevance to this study, cGENIE predicts nutrient (PO$_4^{3-}$) distributions in the North Pacific that are in agreement with present-day observations (151). The coarse resolution also allows cGENIE to be run efficiently and a broad parameter space to be explored: here, we show 28 × 5000–year runs for two sets of boundary conditions.

To test the potential influence of background climate state on our results, we ran experiments under preindustrial and glacial boundary conditions. To simulate a glacial climate state, we reduced the radiative forcing consistent with concentrations of the major greenhouse gases (CO$_2$, CH$_4$, and N$_2$O of 191 ppm, 350 parts per billion (ppb), and 216 ppb, respectively; 153–155). In addition, we increased the zonally averaged planetary albedo profile in the Northern Hemisphere and increased average ocean salinity by ~1 PSU, similar to that used in (12). Our experiments start from the end of 20,000-year equilibrium spin-ups carried out with prescribed atmospheric CO$_2$ of 278 ppm, $\delta^{13}$C of ~6.50‰, and $\Delta^{14}$C of 0‰. The same prescribed atmospheric CO$_2$ concentration and isotopic compositions are used for both preindustrial and glacial boundary conditions, so that the size of the ocean-atmosphere carbon inventory is similar across our experiments. Radiative forcing by CO$_2$ is set independently of the actual concentration of atmospheric CO$_2$ calculated by the ocean-atmosphere biogeochemistry modules and is held constant at either preindustrial or glacial levels as described above.

To simulate an increase in North Pacific ventilation, we reduced the prescribed transfer of atmospheric fresh water from the Atlantic to the Pacific, which implicitly accounts for the net moisture transport between these basins (156, 157). We ran 28 total experiments under preindustrial and glacial boundary conditions, with a range of freshwater transport anomalies from 0.04 to ~0.38 Sv. Experiments were run for 5000 years, which allows ocean circulation and nutrient concentrations to stabilize, with <2% change in mean global surface PO$_4^{3-}$ over the last 500 years of the simulations.

The decrease in atmospheric freshwater transport from the Atlantic to the Pacific increases PMOC and reduces AMOC. The distribution of surface salinity also changes deep-water formation in the Southern Ocean, in turn changing biological pump efficiency. Each of these processes may change atmospheric CO$_2$, and deconvolving their individual influences is beyond the scope of this study. To place broad constraints on changes in pCO$_2$ due to enhanced PMOC, we calculate changes in the ocean’s preformed phosphate inventory resulting from increased input of NPIW and use the theoretical scaling of Ito and Follows (33) to estimate the pCO$_2$ impact. We use a dye tracer to calculate the volumetric contribution of North Pacific waters to the ocean’s interior and use the end-member preformed phosphate composition of the grid box with the maximum contribution of NPIW. We assume that this NPIW replaces water in the Pacific’s interior that has a preformed phosphate of 1.4 µmol/kg (158) and calculate the resulting difference in preformed phosphate inventory by mass balance. The results of this mass balance calculation are shown in fig. S14, with contours showing the general relationship between NPIW volume, preformed phosphate, and pCO$_2$ change, and symbols showing the realizations found in our cGENIE simulations. With enhanced PMOC, the preformed phosphate content of NPIW decreases, as nutrients are flushed from the subpolar gyre, and this low-phosphate water occupies a greater ocean volume, thus increasing the net efficiency of the biological pump and decreasing atmospheric CO$_2$. To estimate the cGENIE scenario most representative of LGM conditions in the North Pacific, we compare modeled $\delta^{13}$C profiles in the NW Pacific to the data (Fig. 5 and fig. S10). An overturning rate of 8 Sv, found with an Atlantic to Pacific freshwater forcing anomaly of ~0.19 Sv, best matches the shape of the LGM $\delta^{13}$C profile (Fig. 5) and also shows a good match to observed changes in $\delta^{13}$C, redox, and $^{14}$C across the basin (fig. S10). We thus estimate a change in pCO$_2$ of ~16 ppm as a result of enhanced PMOC at the LGM. However, we note that the influence of changes in ocean circulation and preformed nutrient inventory on pCO$_2$ may vary depending on circulation regime in the Southern Ocean (34), and that other processes, including sea ice and carbonate compensation, will also influence CO$_2$ but are not captured by this treatment. Further work is required to constrain the impact of PMOC on pCO$_2$ in more detail.

**SUPPLEMENTARY MATERIALS**

Supplemental material for this article is available at http://advances.sciencemag.org/cgi/content/full/6/50/eabd1654/DC1

**REFERENCES AND NOTES**


123. L. Rodríguez-Sanz, P. G. Mortyn, J. C. Herguera, R. Zahn, Hydrographic changes in the tropical and extratropical Pacific during the last deglaciation. Paleoceanography 28, 529–538 (2013).


131. M. Inagaki, M. Yamamoto, Y. Isgarash, K. Ikehara, Biomarker records from core GH02-1030 off Tokachi in the northwestern Pacific during the last 23,000 years: Environmental changes during the last deglaciation. J. Oceanogr. 65, 847–858 (2009).


**Acknowledgments:** The ideas in this manuscript benefited from spirited discussions on the role of the North Pacific in past climates, beginning with a Leopoldina Symposium organized by M. Sarnthein and G. Haug. We thank L. Menviel and E. Galbraith for sharing model output. This work was substantially improved by the comments of several anonymous reviewers.

**Funding:** This work was funded by NERC grant NE/N011716/1 to J.W.B.R., a NERC studentship to B.T., and NSF grant OPP 1643445 to I.E. A.R. acknowledges support from NSF grant 1736771.


**Competing interests:** The authors declare that they have no competing interests. **Data and materials availability:** Data presented in this paper are available as a supplementary data file accompanying this paper online, and on Pangaea, NCDC, and the NERC National Geoscience Data Centre. The cGENIE.muffin model code is hosted on GitHub and can be obtained by cloning: https://github.com/derpycode/cgenie.muffin. Configuration files for the specific experiments focused on in the paper can be found in the subdirectory: genie-userconfigs/MS/raetal.2020. Details of these experiments, plus the command line needed to run each one, are given in the readme.txt file in that directory. All other configuration files and boundary conditions are provided as part of the release. The complete ensemble of model experiments can be obtained by contacting the first author. Note that the specific version of the cGENIE.muffin code used to produce the model results in this paper is git revision: 8e340e5b171b7d86d3e47b0c163200fab623aa67.

Submitted 4 June 2020
Accepted 21 October 2020
Published 9 December 2020
10.1126/sciadv.abd1654

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Sci Adv 6 (50), eabd1654.
DOI: 10.1126/sciadv.abd1654

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DOI: 10.1126/sciadv.abd1654

The PDF file includes:

- Figs. S1 to S14
- Tables S1 and S2
- References

Other Supplementary Material for this manuscript includes the following:

(available at advances.sciencemag.org/cgi/content/full/6/50/eabd1654/DC1)

- Data file S1
Figure S1: Further detail of changes in ventilation (a), export productivity (b), temperature (c), and salinity (d) at the LGM relative to the Holocene, following Figure 2 in the main text. Red indicates an increase at the LGM relative to the Holocene, blue a decrease, and white no clear change outside of 1σ uncertainty. Ventilation proxies (δ¹³C, circles; benthic-planktic radiocarbon offsets, diamonds; redox tracers, squares) are shown here only for sites more northerly than 40 °N (c.f. 40 °N in Figure 2). Productivity data (opal, circles; biogenic barium, squares) are shown here as mass accumulation rate of biogenic material into sediment. In most cases this is calculated using δ¹⁸O or ¹⁴C derived sediment core age models and dry bulk density, either measured or estimated from the compilation of Kohfeld & Chase (26); ²³⁰Th-normalisation is used at a subset of sites, indicated with bold symbol outlines. Temperature reconstructions (planktic foraminiferal Mg/Ca, circles; alkenone saturation index U^K'37, diamonds) and change in salinity (δ¹⁸O on planktic foraminifera corrected for temperature using Mg/Ca and for whole ocean δ¹⁸O and salinity changes due to ice volume) are shown using a graduated scale to illustrate the magnitude of change. This alternative presentation of the data in Figure 2 of the main text supports the conclusion that the subpolar LGM North Pacific was better ventilated at intermediate depths, with lower productivity, and relatively warm and salty surface waters.
Figure S2: Further detail on δ¹³C depth profiles, highlighting minimal influence of age model uncertainties (a) and regional differences (b). (a) Benthic foraminiferal δ¹³C data from the northwest Pacific from the Holocene (open symbols), LGM (closed blue symbols), and a timeslice 3000 years older than the LGM timeslice (closed yellow symbols) (6). Cibicides spp. are shown in circles, Uvigerina in squares; although Uvigerina are not typically used for estimates of bottom water δ¹³C, they are indistinguishable from the Cibicides data at the LGM, and allow extension of the profile deeper in time (as Cibicides data do not extend beyond the LGM). LGM and LGM+3ka δ¹³C data have been corrected for a whole ocean δ¹³C change of 0.34 ‰ (17). The LGM and Holocene foraminiferal δ¹³C are fit with a general additive model. (b) Holocene (open symbols) and LGM (closed symbols) foraminiferal δ¹³C data from across the basin, as used in the ventilation proxy compilation (see Methods). Data from west of the dateline is shown in pink, east in green, and each profile has been fit with a general additive model. Water column δ¹³C for the North Pacific (black dotted line) and North Atlantic (red dotted line) are taken from the recent compilation of (162), and include all values between 40-65 °N in each basin; here a generalized additive model fit to the data is shown. The Pacific depth profile of δ¹³C is notably different at the LGM, with elevated values found at intermediate depths. The glacial increase in intermediate depth δ¹³C is more apparent in the west of the basin relative to the east; this is suggestive of a source of local ventilation in west of the basin.
Figure S3: Meridional $\delta^{13}C$ sections from benthic foraminifera in the Western (a, b) and Eastern (c, d) Pacific in the late Holocene (a, c) and at the LGM (b, d). Data are taken from the compilation of (17), with glacial data corrected for a whole ocean $\delta^{13}C$ change of 0.34 ‰, and are plotted using Ocean Data View (54). The location of the data making up these sections are shown in the inset maps. Areas of poor data coverage are shown in grey. At the LGM there is a substantial increase in $\delta^{13}C$ of intermediate waters in the North Pacific, indicating enhanced ventilation. The largest change is observed in the West of the basin, as expected from the formation of a deep western boundary current.
Figure S4: Latitudinally binned deglacial changes in SST (a) and $\delta^{18}$O of seawater (b). (a) At each site Mg/Ca and/or U$^{137}$ SSTs were converted to a difference from Holocene (0-10 ka, light grey box) and binned by latitude. (b) $\delta^{18}$O of seawater, derived from paired measurements of Mg/Ca and $\delta^{18}$O$_{calcite}$ and corrected for changes in global ice volume, was converted to a difference from Holocene and binned by latitude. The latitudinally binned SST and $\delta^{18}$O$_{seawater}$ data were modelled as a function of time using a generalised additive model (GAM), with the 68% and 95% Bayesian credible intervals shown (161, 163). For each latitudinal SST and $\delta^{18}$O$_{seawater}$ bin, the LGM (19-21 ka, dark grey box) value is given with the 95% confidence interval.
Figure S5: Model sensitivity tests, showing the salinity (a) and phosphate concentration (b) of surface waters in the subpolar North Pacific, as a function of exchange with the subtropical gyre ($V_{STG}$) under different boundary conditions. Base state values are given in Table S2; P-E+R is net precipitation plus riverine run-off; $S_{STG}$ is the salinity of the subtropical gyre; $V_{UP}$ is the upwelling flux; $[PO_4^-]_{UP}$ is the phosphate concentration of upwelled water. Subpolar salinity increases when (i) net precipitation is reduced, (ii) upwelling is increased, and (iii) exchange with the subtropics is enhanced, particularly when subtropical salinity is elevated. Phosphate concentrations in the subpolar gyre are reduced by exchange with the subtropics and by decreasing the phosphate concentrations of upwelled waters, and are increased by increasing the upwelling flux. Note that the net effect of doubling vertical exchange while halving subsurface nutrients is similar to that of halving subsurface nutrients alone (Fig. S5). Because the wind- and tidally-driven transfer of water from the subsurface to the surface North Pacific is high, vertical nutrient gradients are relatively low, so further increase in vertical exchange – for instance due to convective mixing – has less influence than decreasing the nutrient content of the upwelled water.
Figure S6: The relationship between Atlantic to Pacific freshwater forcing and maximum PMOC in cGENIE. Forcing values are shown as anomalies relative to the prescribed Atlantic to Pacific freshwater flux, which has a default value of 0.23 Sv (i.e. the default flux of 0.23 Sv plots as 0 on the figure above).
Figure S7: Representative cGENIE experiments spanning a range of PMOC states. Top row shows phosphate and overturning anomalies, illustrating the flushing of phosphate from the upper reaches of the North Pacific as overturning increases. Bottom row shows the North Pacific water fraction at 1000 m, based on a dye tracer tagging surface waters north of 40 °N in the North Pacific. These experiments were run under glacial boundary conditions. Newly formed NPIW travels south as a western boundary current, while the eastern basin feels relatively little influence of NPIW at low latitudes. Even at high overturning rates, North Pacific deep water is largely confined to the Pacific basin. Note that the central experiment (-0.19 Sv forcing and 8 Sv PMOC) shows the best fit to the data in Figures 5, S10.
Figure S8: Phosphate and Dissolved Inorganic Carbon (DIC) anomalies under enhanced PMOC in simulations with the LOVECLIM (1.1) and UVic (v2.9) Earth System Models by Menviel et al. (28). These simulations were forced with North Atlantic freshwater hosing of 0.1 Sv for 2000 years (fNA in the rubric of the original publication), which produces a strong PMOC in response (stream function contours are shown for LOVECLIM). As in eGENIE, the establishment of an active overturning circulation in these higher resolution models flushes nutrients from the upper reaches of the North Pacific.
Figure S9: Transient response of North Pacific surface nutrients at the onset of enhanced overturning in cGENIE. Data are shown for the NW of the subpolar gyre in simulations spanning a range of overturning states (as in Figure S7). For the first several hundred years after North Pacific salinity is increased, surface phosphate shows a transient increase (top panel), as convective mixing increases (bottom panel) and subsurface nutrients remain relatively high (middle panel). As overturning becomes established, nutrient-rich subsurface waters are flushed out, reducing surface nutrient supply despite the increase in convective mixing, as described in main text (e.g. Figure 4).
Figure S10: Simulated changes in δ¹³C (a), oxygen concentration (b), and Δ¹⁴C (c) compared to the changes seen in our LGM data compilation. The simulated section is from 165 °W for the experiment run with an Atlantic to Pacific freshwater forcing of -0.18 Sv under glacial boundary conditions and is shown as an anomaly from conditions prior to anomalous salinity forcing. This experiment shows the best match to LGM δ¹³C profile data (Figure 5). Note that as the abyssal Pacific likely remains ventilated by Southern Ocean waters at depths greater than 2000 m, and as we have not made any changes to the Southern Ocean in our simulations, some offsets between the experiments and the data are to be expected at depth.
Figure S11: Annual mean meridional overturning circulation streamfunction (a), surface temperature (b), and sea surface salinity (c) in the North Pacific in simulations of preindustrial and LGM climate with the Community Climate System Model version 3 (CCSM3). The North Pacific streamfunction is diagnosed from the difference between the Eulerian mean streamfunction of the global ocean and the Atlantic, both of which are output directly from the model. Positive values denote a clockwise circulation. Preindustrial properties are averaged over model years 470-489 of the preindustrial control simulation described by Otto-Bliesner et al.\cite{164}, and LGM properties over model years 380-399 of the simulation described by Otto-Bliesner et al.\cite{164, 165}. LGM conditions include greenhouse gas forcings and insolation at 21 ka values, and ice sheets based on the 21 ka reconstruction from ICE-5G\cite{166}. Note the enhanced intermediate-depth overturning, warming of NW Pacific waters, and increased salinity of surface waters in the North Pacific under glacial conditions.
Figure S12: Volumetric contribution of North Pacific waters to the global ocean in the CM2Mc model under a range of boundary conditions. Simulations are from (167) and were designed to ensure equilibration of the deep ocean. North Pacific waters were tagged using a tracer in surface waters >30 °N in the Pacific basin. The fraction of North Pacific waters filling the global ocean systematically increases with decreasing CO₂, supporting the idea that a colder climate enhances the formation of NPIW. We note that in these simulations NPIW formation is decreased in the presence of LGM ice sheets, although the geometry of the ice sheets used in these simulations is now thought to have some significant inaccuracies compared to more recent reconstructions (168). CM2Mc also has a cold bias in the North Pacific, which leads to relatively high NPIW formation under modern conditions (167). Nevertheless, these simulations illustrate the potential for enhanced North Pacific overturning in a state-of-the-art Earth system model under cold climate conditions.
Figure S13: Intermediate water δ¹³C from cGENIE over a range of overturning states in the North Atlantic (orange circles and dashed line) and North Pacific (blue crosses and solid line). Values are taken from 1000 m at a central position in each Northern basin (40°N 165°W in the Pacific, 48°N 35°W in the Atlantic). Intermediate depth δ¹³C is consistently lower in the North Pacific than in the North Atlantic under similar rates of overturning.
Figure S14: Change in the inventory of preformed phosphate (a) and atmospheric CO$_2$ (b) as a function of changes in NPIW’s global volume fraction, preformed phosphate composition, and PMOC (c). Shaded contours show the result of mass balance calculations, assuming NPIW replaces water with preformed phosphate of 1.4 (169), with CO$_2$ change then calculated using the scaling of Ito & Follows (33). Symbols in (a) and (b) show the NPIW volume and preformed phosphate found in our cGENIE experiments, and (c) shows the calculated change in CO$_2$ using these values and the scaling of (33) against the maximum PMOC for each experiment. Note the sensitivity of CO$_2$ to changes in PMOC of ~6-12 Sv (c), driven by the increase in the volume of the global ocean occupied by NIPW (b).
Table S1:

Results of calculations based on Warren’s box model. Input values include net precipitation (P-E) and riverine runoff (R), the salinity of the subtropical gyre (S\textsubscript{STG}), advection from the subtropical gyre (V\textsubscript{STG}), and the phosphate concentration of upwelled waters (PO\textsubscript{4}\textsuperscript{UP}). Output values are the salinity (S\textsubscript{SPG}) and phosphate concentration (PO\textsubscript{4}\textsuperscript{SPG}) of the top 200 m of the subpolar gyre.

<table>
<thead>
<tr>
<th>Model inputs</th>
<th>Model outputs</th>
</tr>
</thead>
<tbody>
<tr>
<td>P-E + R</td>
<td>S\textsubscript{STG}</td>
</tr>
<tr>
<td>Sv</td>
<td>psu</td>
</tr>
<tr>
<td>Modern Pacific - observed</td>
<td>0.21 + 0.07</td>
</tr>
<tr>
<td>Modern Pacific - model</td>
<td>0.21 + 0.07</td>
</tr>
<tr>
<td>Subtropical salinity + 1 psu</td>
<td>0.21 + 0.07</td>
</tr>
<tr>
<td>Gyre exchange + 2 Sv</td>
<td>0.21 + 0.07</td>
</tr>
<tr>
<td>Subtropical salinity &amp; Gyre exchange combined</td>
<td>0.21 + 0.07</td>
</tr>
<tr>
<td>Net rainfall (P-E) - 30%</td>
<td>0.15 + 0.05</td>
</tr>
<tr>
<td>Subtropical salinity, Gyre exchange, P-E combined</td>
<td>0.15 + 0.05</td>
</tr>
<tr>
<td>As above with 2x upwelling &amp; 0.5x upwelled [PO\textsubscript{4}]</td>
<td>0.15 + 0.05</td>
</tr>
<tr>
<td>Modern Atlantic - observed</td>
<td>0.10 + 0.06</td>
</tr>
</tbody>
</table>

Table S2:

Model input parameters for modern base state. P-E and R are from ERA-Interim reanalysis from (148), V\textsubscript{STG} and V\textsubscript{UP} are from (3) and salinity and phosphate concentrations are from World Ocean Atlas, computed using the box averaging tool in Ocean Data View (149).
REFERENCES AND NOTES


54. R. Schlitzer, Ocean Data View (2010); http://odv.awi.de.


100. C. M. Lindsay, S. J. Lehman, T. M. Marchitto, J. D. Carriquiry, J. D. Ortiz, New constraints on
deglacial marine radiocarbon anomalies from a depth transect near Baja California.
Tanhua, M. Hoppema, S. Jutterström, The Global Ocean Data Analysis Project version 2
(GLODAPv2)—an internally consistent data product for the world ocean. Earth Syst. Sci. Data 8,
(2016).
Manning, M. Niu, R. W. Reimer, D. A. Richards, E. M. Scott, J. R. Southon, R. A. Staff, C. S. M.
Turney, J. van der Plicht, IntCal13 and Marine13 radiocarbon age calibration curves 0–50,000 years
103. H. Kawahata, K.-i. Ohkushi, Y. Hatakeyama, Comparative Late Pleistocene paleoceanographic
104. L. Maeda, H. Kawahata, M. Nohara, Fluctuation of biogenic and abiogenic sedimentation on the
Shatsky Rise in the western North Pacific during the late Quaternary. Mar. Geol. 189, 197–214
(2002).
105. I. L. Hendy, T. F. Pedersen, J. P. Kennett, R. Tada, Intermittent existence of a southern Californian
upwelling cell during submillennial climate change of the last 60 kyr. Paleoceanography 19 (2004).
107. R. S. Ganeshram, T. F. Pedersen, Glacial-interglacial variability in upwelling and bioproductivity
108. L. D. Keigwin, G. A. Jones, P. N. Froelich, A 15,000 year paleoenvironmental record from Meiji
109. S. A. Gorbarenko, Stable isotope and lithologic evidence of late-glacial and Holocene
110. B. G. Brunelle, D. M. Sigman, S. L. Jaccard, L. D. Keigwin, B. Plessen, G. Schettler, M. S. Cook,
G. H. Haug, Glacial/interglacial changes in nutrient supply and stratification in the western subarctic


