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Synchronous subsurface ocean warming in western boundary regions of both hemispheres during Heinrich Stadial 1

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Determining whether both hemispheres warmed synchronously during the last deglaciation is important for identifying mechanisms for ice-age terminations. Here we show that intermediate waters in the southwestern Pacific Ocean warmed simultaneously with some of the upper ocean waters in the Atlantic during Heinrich Stadial 1. In both regions, proxy records from benthic foraminifera show warming to interglacial temperatures beginning ~18,000 years ago and concluding by ~15,000 years ago. Our observations require a mechanism capable of producing rapid warming in these nearly antipodal ocean basins. We propose that some of the oceanic warming signal co-registered between these sites can be explained by a shift of the Southern Hemisphere Westerly Winds, which increased the temperature of intermediate water masses produced in the Southern Ocean. Taken together with results from Earth-system modeling experiments, we suggest that the Southern Hemisphere climate system played a role in regulating interhemispheric oceanic heat transport at millennial timescales.

Introduction:

The last deglaciation (18–11 calibrated kiloyears before present, cal kyr BP) was marked by melting ice sheets, rising oceanic and atmospheric temperatures, and increasing atmospheric carbon dioxide (CO₂) concentrations. What caused this major climatic transition represents an outstanding paleoclimatology problem. Warming of the last deglaciation was punctuated by the Heinrich Stadial 1 (HS1, 17.8–14.7 kyr BP), Antarctic Cold Reversal (ACR, 14.7–12.9 kyr BP), and Younger Dryas (YD, 12.9–11.6 kyr BP) millennial-scale climate oscillations¹. These events were characterized by abrupt, widespread climate change^{2,3,4,5,6}, but underlying causes remain unclear. One topic of ongoing research is the phasing of warming between the Northern and Southern Hemispheres during these episodes^{7,8,9}, which is important for constraining potential mechanisms of climate change. Here, we evaluate this problem by presenting a new record of subsurface temperatures in the southwest Pacific that spans the last glacial termination. We then compare our record to complementary records in the Atlantic. On the basis of this comparison, we evaluate possible mechanisms for interhemispheric paleoceanographic changes during the last termination.

We present a record of water temperature developed from a sediment core collected from north of New Zealand at the depth of modern Antarctic Intermediate Water (AAIW) (Fig. 1). We measured Mg/Ca and Mg/Li on two benthic foraminifera species, *Uvigerina peregrina* and *Hoeglundina elegans*, respectively, to provide a record of AAIW temperature^{10,11} from the Last

Glacial Period (LGP) through the mid-Holocene. This dataset contributes to the relatively sparse collection of AAIW trace element records at this resolution (an average of ~5–6 samples per kyr) and offers a new perspective on this water mass during the deglaciation.

Today, AAIW is primarily identified as a cool (4–5 °C) and relatively fresh (34.2–34.6 psu) layer that exists between the depths of ~800 and 1500 m, depending on location. AAIW forms around the Subantarctic Front (SAF)¹² in the Southern Ocean, primarily in the southeast Pacific (Fig. 1a). From this formation region, AAIW travels eastward through the Drake Passage into the Atlantic, or northward and into the south Pacific subtropical gyre. The influence of AAIW extends throughout the tropics and subtropics of the Southern Hemisphere, and into the subtropical North Atlantic as well, indicated by intermediate-depth salinity minima, silica maxima, and CFC profiles^{13,14,15,16}.

Core RR0503-79JPC (hereafter 79JPC) was collected in the southwest Pacific Ocean (36.959°S, 176.593°E, Fig. 1a, b) at 1163 m water depth in New Zealand's Bay of Plenty, within the low-salinity core of modern Antarctic Intermediate Water (AAIW; Fig. 1c, d). This part of the southwest Pacific contains multiple varieties of AAIW. These include AAIW from the Pacific subtropical gyre, AAIW sourced more directly from the Southern Ocean south of New Zealand, and AAIW that has circulated through the Tasman Sea¹⁷. These circulation patterns and origins yield minor differences in salinity, nutrients, temperature, and other properties within the bounds that define AAIW^{18,19}. For example, the Tasman AAIW dominant at the location of 79JPC is slightly warmer and saltier than other varieties in the southwest Pacific (Fig. 1c).

Because of its rapid renewal rate (decades rather than centuries)¹⁹, AAIW in the modern ocean has been a research focus for tracking the influence of anthropogenic CO₂ and global warming in the Southern Ocean and beyond^{20,21,22,23}. The quick cycling of AAIW relative to deeper water masses makes paleo records of AAIW excellent candidates for capturing signatures of rapid deglacial climate change. Here we interpret our downcore observations in relation to novel output from the GFDL Earth System Model 2M (ESM2M)^{24,25,26} and other upper ocean water temperature records of the last glacial termination.

Results and Discussion:

Deglacial warming in the southwest Pacific

Results are shown in Figure 2. The average reconstructed glacial (24.3–17.8 cal kyr BP) temperature at this site is 1.8 ± 0.3 °C (temperature uncertainty is 2 standard error, 'se', unless otherwise noted), while the average early Holocene (beginning of Holocene to end of record, 11.6–6.5 cal kyr BP) temperature is 4.7 ± 0.4 °C, constituting a total warming of ~2.9 °C. Abrupt warming begins 18.5 ± 0.7 cal kyr BP (age uncertainty is 1 standard deviation, σ , unless otherwise noted), coinciding with or slightly preceding the beginning of HS1 (Fig. 2a). During HS1, AAIW temperatures average 3.9 ± 0.6 °C, while during the latter part of HS1 (16.1–14.7 cal kyr BP), AAIW temperatures average 4.9 ± 0.5 °C, indistinguishable from Holocene temperatures. These temperatures, combined with breakpoint analysis (see Supplementary Methods, Supplementary Fig. 1), suggest that full LGP-Holocene warming at this site was complete as early as 16.1 ± 0.9 cal kyr BP, followed by a transient peak in temperature during the YD (Fig. 2a).

These results are consistent with sea surface temperature (SST) records in the southwest Pacific that show both warming and a southward shift in the Subtropical Front (STF) at ~ 18 cal kyr BP^{27,28}. The STF is a boundary that separates warm subtropical water from cool subpolar water, and it can be identified by a strong SST gradient across the front^{29,30}. The STF's position is in some places pinned to bathymetric features such as the Chatham Rise, but in others (such as in the Tasman Sea), it is free to move dynamically in response to overlying wind systems. Latitudinal transects reconstructing sea-surface temperature south of Tasmania show a southward shift of the STF during the last deglaciation, from $\sim 44^\circ\text{S}$ at ~ 19 cal kyr BP to $\sim 45 - 46^\circ\text{S}$ by ~ 16 cal kyr BP²⁸. Likewise, influx of *G. ruber* south of Australia during early HS1 indicates that the STF had shifted south, allowing for the Leeuwin Current to transport tropical water masses south of Australia³¹. To the south and west of New Zealand, similar transects indicate rapid regional ocean warming during the early deglaciation, from $\sim 19 - 16$ cal kyr BP (*i.e.*, before and during early HS1), coupled to a southward shift of the STF²⁷. North and east of New Zealand, studies also show deglacial warming during HS1, though the STF position was constrained by the Chatham Rise to $\sim 44^\circ\text{S}$ ^{32,33,34}. The southward STF shift south of Australia and in the Tasman Sea and warming in the southwest Pacific, combined with independent evidence for rapid glacial recession in New Zealand's Southern Alps^{35,36}, indicates that a major southward shift in atmospheric and oceanic circulation occurred in the Southern Hemisphere during HS1.

Altogether, these temperature records indicate that both the atmosphere and upper ocean at Southern Hemisphere mid-latitudes experienced a rapid warming to near-interglacial values during the first half of the deglaciation, synchronous with a southward shift of wind and frontal systems^{37,38} (Fig. 3). While some warming trends begin sooner, starting as early as $\sim 19 - 20$ cal kyr BP (RR0503-87JPC³³, MD97-2121³⁹, RS147-GC07²⁸), southwest Pacific surface-temperature reconstructions and mountain glacier chronologies^{35,36} indicate the sharp onset of rapid warming at approximately the same time as the initial increase in atmospheric CO_2 ($\sim 17.7 \pm 0.2$ cal kyr BP, 1σ). Most glacial-interglacial warming was nearly complete by the end of HS1 (~ 14.7 kyr BP; RR0503-79JPC, this study; RR0503-87JPC³³, MD97-2120³⁹, RS147-GC07²⁸; Fig. 3), whereas CO_2 had risen by only ~ 40 ppm by that time ($\sim 40\%$ of the full deglacial change). As a greenhouse gas, CO_2 must have provided a feedback on shallow ocean warming during the deglaciation. However, surface-ocean warming commenced ~ 2 ky before CO_2 began its deglacial increase (Fig. 3), and SW Pacific surface-ocean warming and glacial recession achieved interglacial values several millennia before CO_2 rose to interglacial levels. Together these observations indicate that the initial rise in atmospheric CO_2 was a consequence of the process of deglacial warming, and therefore would have been a positive feedback on, but not a primary driver of, the transient warming of the middle-latitude surface ocean and atmosphere during the last deglaciation. Our results are consistent with the existence of an additional, complementary mechanism driving the transition from a glacial to an interglacial mode.

Of the temperature records examined here, both 79JPC and MD97-2120 (located south of Chatham Rise) show a warming onset indistinguishable from the onset of the southward migration of the SHW at 18.6 ± 0.2 cal kyr BP³⁸ (Fig. 3), suggesting a link between movement of the SHW and warming of the Southern Ocean. These records, along with others from the southwest Pacific, also show a temperature maximum late in the deglaciation (Younger Dryas to early Holocene) followed by a cooling toward mid-late Holocene temperatures (Fig. 3). The SHW reached their southernmost position at the same time as the temperature maxima, from $\sim 12 - 10$ cal kyr BP³⁸ (Fig.

3), further suggesting a causal relationship between the SHW and Southern Ocean temperatures. The alignment of these temperature trends with SHW position reconstructions suggests a potential dynamical link between the two, which we investigated through Earth system modeling.

A role for the Southern Hemisphere westerlies?

The atmospheric warming^{35,36}, sea-surface warming^{32,39,28,27,33}, and subsurface warming (this study) observed at these locations in the southwest Pacific are consistent with predictions for a rapid southward shift in the Southern Hemisphere westerlies (SHW) as simulated in ESM2M, a coupled Earth System Model (ESM) from NOAA's Geophysical Fluid Dynamics Laboratory^{24,41,26}. These experiments include a "preindustrial" (control) and an ensemble of 30 "wind" (perturbation) simulations, run for a 200-year integration time. The perturbation simulation approximates poleward-shifted and strengthened SHW via a momentum-flux adjustment, with no other changes with respect to the control run²⁶. These experiments are not intended to fully simulate Holocene or LGM conditions; rather, they are designed to change a single parameter and thereby enable a straightforward assessment of the impact of a southward shift of the westerlies on the rest of the ocean-climate system. In the poleward-shifted winds scenario, not only do ocean temperatures increase at the locations discussed above, but other model predictions are also matched by observations, including a decrease in the water mass age, which supports observations of decreased water-mass radiocarbon ages^{42,43} (Supplementary Fig. 2) and increased $\delta^{13}\text{C}$ values⁴⁴ of intermediate water in the southwest Pacific. The EMS2M wind-perturbation experiment also predicts a warming of $\sim 1^\circ\text{C}$ at the depth and location of 79JPC, along with broader regional warming of AAIW (Figure 3e), without accounting for the different boundary conditions associated with LGM climate state. Therefore, a shift in the SHW alone has the potential to explain at least one third of the total AAIW temperature increase across HS1 at site 79JPC, and within the necessary time frame. This supports the hypothesis that changes in atmospheric circulation played an important role in early deglacial warming of the mid-latitudes.

What could have caused rapid shifts in the SHW and upper ocean during HS1? Our observed subsurface warming was preceded by rising SSTs nearby in the Bay of Plenty³³ (Fig. 3), where temperatures began to increase by $\sim 20.4 \pm 0.9$ cal kyr BP, earlier than in most other ocean basins^{3,6,45,46}. This was followed by warming at the southern end of the East Australian Current system at $\sim 19.3 \pm 0.8$ cal kyr BP (RS147-GC07²⁸; Fig. 3), and then by warming south of the modern STF³⁹. Early warming in the subtropics of the southwest Pacific could reflect expansion of the Western Pacific Warm Pool (WPWP), increased transport of warm waters from the tropics, and/or a shift in austral summer duration⁴⁷ that affected wind position and strength²⁶. A warmer WPWP is suggested by SST records from the tropics of the western Pacific, indicating that warming began in these regions before the beginning of HS1 and before the onset of atmospheric CO_2 rise, perhaps as soon as 20 – 22 cal kyr BP^{48,49,50,51,52}. Warming and expansion of the tropics/low-latitude troposphere is inherently and dynamically linked to southward-shifted and intensified SHW^{53,54}, a phenomenon that has the power to drive warming in both the terrestrial and marine realms.

Comparison with subsurface water signals from the central and North Atlantic

Three intermediate water temperature records from the central Atlantic also exhibit a signature of rapid HS1 warming similar to that observed at 79JPC (Fig. 4). Like 79JPC, these additional temperature records are reconstructed using Mg/Li of benthic foraminifer *H. elegans*,

and they have similar resolutions and suitable age control for direct comparisons. Intermediate water temperatures reconstructed from ~550 m depth in the Florida Straits (KNR166-2-26JPC⁵⁵) indicate warming beginning at $\sim 18.3 \pm 0.6$ cal kyr BP (Fig. 4). Interglacial temperature values were first achieved by ~ 16.2 cal. kyr BP, and water temperature peaked during the YD⁵⁵.

A strikingly similar pattern and magnitude of warming is observed from a core at 950 meters water depth on the Demerara Rise, central Atlantic (KNR197-3-46CDH⁵⁶): deglacial warming began at $\sim 17.9 \pm 0.8$ cal kyr BP, and warming was complete by the end of HS1, with maximum temperatures occurring late in the glacial termination, centered around the YD⁵⁶ (Fig. 4). Unlike sites 79JPC and 26JPC, 46CDH shows little to no net LGP-Holocene temperature change, yet the temperature signal still exhibits deglacial warming during HS1, with maximum temperatures achieved at the same time as the other two core sites (Fig. 4).

On the Brazil Margin at 1268 m depth, (KNR159-5-36GGC¹¹), intermediate waters show deglacial warming beginning at $\sim 17.6 \pm 0.5$ (1 σ) cal. kyr BP with deglacial temperatures achieved by $\sim 14.6 \pm 0.4$ cal. kyr BP, coinciding with the end of HS1 (Fig. 4). In contrast to the sites discussed above, no YD temperature maximum was detected in 36GGC, which may be due to its greater water depth and/or lower temporal resolution during the late glacial period.

Records of ice sheet and alpine glacier change

The signatures of upper ocean warming discussed above echo patterns of atmospheric warming inferred from terrestrial records of glacier recession at middle latitudes in both hemispheres. Mountain glaciers in the Southern Alps of New Zealand exhibited extensive recession to near-interglacial conditions during the early phase of HS1 between ~ 18 and 16 cal ky BP, representing as much as $\sim 90\%$ of full-glacial-to-early Holocene snowline rise^{26,35,36}. Icefields in southern South America likewise receded during that time^{57,58}. At opposing latitudes in the Northern Hemisphere, deglaciation was underway in the North Atlantic region during HS1. Near the western North Atlantic, the southwestern margin of the Laurentide Ice Sheet had receded from eastern Maine between 17 and 15 cal kyr BP^{59,60}. In the eastern North Atlantic, glacier recession from western Ireland had taken place by as early as ~ 17.4 cal kyr BP^{61,62}, and the southwestern margin of the Scandinavian Ice Sheet had receded to its late-glacial position by as early as ~ 16.5 cal kyr BP, i.e., during the early part of HS1⁹.

This pattern of summer warming is also seen in some SST records from the North Atlantic. While many records around the North Atlantic indicate cooling during HS1 and the YD, particularly along the Iberian Margin⁶³, some sites show a pattern more consistent with summertime warming during HS1. For sites located in open-ocean settings, within the region of the modern North Atlantic Current system, records reveal HS1 temperature increases without any major cooling phases^{3,63}. The emerging picture of North Atlantic conditions suggests a complex, regionally variable climate during these millennial-scale deglacial events.

One explanation may lie in the proximity to meltwater sources. Warm summers and enhanced meltwater fluxes into the North Atlantic from the southern sector of the Scandinavian Ice Sheet during HS1^{64,65} are suggested to have freshened and preconditioned the surface ocean for the spread of winter sea ice and intensified seasonality^{66,67}. A derivative impact of summer-warmth-induced meltwater fluxes was cold, freshened surface waters recorded in locations monitoring ice melt⁶⁸ and mean-annual conditions recorded in Greenland ice cores^{7,69}. Variability

in temperature records around the North Atlantic may therefore reflect seasonal changes, or regional ocean circulation changes resulting from meltwater.

Also noteworthy is the coincidence of cooling during the Bølling-Allerød (B-A), synchronous with the ACR in the Southern Hemisphere. The episode of upper-ocean cooling registered in the North Atlantic intermediate water sites 26JPC and 46CDH during the B-A/ACR was accompanied by glacier resurgence in the British Isles⁷⁰ and in Scandinavia⁹, further indicating a pattern of deglacial climate change in the North Atlantic region resembling the Southern Hemisphere signature. This suggests that the upper ocean and atmosphere of the southwestern Pacific and the central and northern Atlantic exhibited simultaneous deglacial warming during HS1. Any mechanism for what drove the last global deglaciation must account for synchronous, rapid oceanic and atmospheric warming to near-interglacial conditions at these antipodal sectors of the planet.

Mechanisms of rapid global change

What caused the observed upper ocean and atmospheric warming during the glacial termination? Relevant processes include wind shifts, low-latitude temperature changes, and variations in ocean circulation, all of which are interlinked and therefore not mutually exclusive^{71,72,73}. We explore some possibilities below.

The SHW were located northward, closer to the equator during the LGP than they are today^{38,74,76,76}. During the last deglaciation, they moved southward (poleward), and, where not constrained by bathymetry, the Southern Ocean fronts moved as well^{5,6,27,28,38,77,78}. Coupled wind and frontal migrations expand subtropical gyres and the West Pacific Warm Pool, allowing more heat export to higher latitudes^{26,38}. Notably, a southward shift of the SHW has potential to influence warm ocean currents in multiple locations that affect the Indian, Pacific, and Atlantic basins: It has been posited that simultaneous warming in multiple regions of the Indo-Pacific and Atlantic could be explained by the concurrent opening of pathways for Agulhas Leakage into the Atlantic and the Leeuwin Current south of Australia as a result of shifting winds and surface current systems³¹. Results from model simulations suggest that the Tasman Sea may act as another important area for a shift in the routing of warm, subtropical waters to new regions²⁶.

A shift in the SHW also has implications for the balance of air-sea gas exchange in the Southern Ocean^{25,79}, offering another explanation for the correlation between rising temperatures and atmospheric CO₂ concentrations during HS1^{76,80,81}: namely, that a southward shift in the SHW could have been responsible for both adding heat to the Southern Ocean and for increasing the outgassing of CO₂ from the deep ocean through an increase in upwelling^{4,37,38}. In this scenario, initial warming precedes the onset of CO₂ rise, but greenhouse gas forcing subsequently reinforces global temperature increases and helps fuel the process of deglaciation^{9,53,82}.

The heat trapped by CO₂ and also redistributed by shifting ocean circulation in the ocean surface could reach intermediate depths via different mechanisms. Results from the ESM2M model^{24,26,41} show that sites 26JPC, 46CDH, and 36GGC, along with most of the subsurface Atlantic, are expected to warm following a poleward shift in the SHW (Fig. 5 g, h), indicating that at least part of the warming in these sites could be the direct result of wind forcing, as seen at site 79JPC. But by what mechanisms could such a wind shift accomplish this? Here we consider two possibilities: 1) temperature change of a shared water source, and 2) a parallel downward mixing of surface heat in both hemispheres.

The central Atlantic sites are not bathed by pure AAIW, which is modified and mixed with other water masses during its journey northward across the equatorial latitudes. Still, these core sites are influenced by AAIW today, demonstrated by profiles of salinity and nutrients such as silicic acid^{13,14,15,83}. Additionally, a subsurface layer with locally elevated CFC-11 and relatively low $\Delta^{14}\text{C}$ in WOCE transects were identified in the western central Atlantic as a tongue of AAIW¹⁶. However, like 79JPC, these sites are also located along the western boundary of the basin, and they are situated in a position to capture a combination of low- and high-latitude changes, including those originating in the Southern Ocean and transported northward. Given these observations, it is possible that these disparate core sites were influenced by subsurface waters with a common southern origin. Warming within the Southern Ocean would increase temperatures at sites influenced by AAIW, even if it is unlikely to account for the full magnitude of warming during HS1.

Alternatively, considering that AAIW is mixed with shallower water masses at all of the sites considered, the overlying subtropical waters could be another source of warming. As mentioned above, the variety of AAIW at site 79JPC is warmer and saltier as a result of mixing in the Tasman Sea¹⁷, while in the Atlantic, the amount of AAIW influence north of the equator is small compared to local sources. It is therefore possible that the central Atlantic and south Pacific temperature records are synchronized by a wind-driven mechanism that simply drives downward heat mixing in both hemispheres, without requiring the sites to be bathed by the same water mass. Specifically, a poleward shift of the westerly winds in both hemispheres would expand the Subtropical Gyres, whose circulation would briefly slow as a result of the expansion but ultimately spin up due to the stronger wind stress^{84,85}. This intensification would deepen and thicken the depths of SAMW, which would then provide a larger proportional contribution to AAIW^{86,87}. In the central Atlantic, a poleward shift in the South Atlantic subtropical gyre may have also led to a strengthened North Brazil Undercurrent, as is seen under warming conditions today⁸⁸. Similar changes are suggested to have occurred during HS1, evidenced by sortable silt records of flow speed at intermediate water depths⁸⁹. These changes would have implications for the amount of water transferred from the South to the North Atlantic basin. A greater amount of heat could have thus been transported from the surface to intermediate water depths in both hemispheres without needing to invoke a common water source bathing all of the sites but rather through atmospheric forcing.

Changes in Atlantic meridional overturning circulation (AMOC) may also have contributed to the degree of downward mixing occurring at these sites. Evidence suggests that AMOC experienced variations during the last glacial termination⁹⁰. It has been proposed that millennial-scale deglacial changes in AMOC could have enhanced mixing between thermocline waters and intermediate waters below⁹¹, which is another mechanism for warming intermediate waters during the deglaciation. A change in AMOC strength does not exclude other mechanisms: it could be dynamically linked to the inferred wind shifts, as surface wind stress ultimately contributes to overturning circulation^{72,92}.

Conventional views of reduced AMOC during HS1 and the YD present challenges when considered alongside recent observations. Models suggest that reduced AMOC during HS1 would result in colder atmospheric temperatures⁹³ in the North Atlantic as a result of lessened northward heat transport, whereas melting circum-North Atlantic glaciers^{9,61,64} indicate that warming was occurring in the atmosphere, at least during the summer months. Another consideration is new evidence suggesting that factors other than ventilation age strongly affect $^{231}\text{Pa}/^{230}\text{Th}$ ⁹⁴, and that production of deep waters in the North Atlantic may have been altered rather than shut down^{95,96,97}.

While proxy records indicate that important changes were occurring in Atlantic circulation, mounting evidence suggests that early interpretations may need to be revisited. Additional work is needed to probe the full complexity and nuance of AMOC changes during deglaciation and their links to millennial-scale climate change globally.

With the available data, we can not yet definitively identify which of these mechanisms explain the warming seen at intermediate water sites during HS1 and the subsequent peak temperatures during the YD and early Holocene. However, the indistinguishable timing of rapid mid-depth warming in the Southwest Pacific and in the central Atlantic during HS1 is consistent with the suggestion that oceanographic changes in the two regions were tied to a common cause. Further, the mechanism must be capable of changing rapidly, and capable of influencing both the southwest Pacific and the central Atlantic simultaneously. Considering the results of our model experiments and available paleorecords, we posit that changes in the Southern wind system played a key role in the last glacial termination.

Implications

As growing numbers of high-resolution, well-dated paleoclimate records are developed, a clearer picture is emerging regarding the timing and mechanisms of deglacial climate change^{98,99}. Rapid, synchronous warming to near-interglacial conditions, as registered by mountain glaciers and by upper-ocean temperature proxies at sites in both hemispheres, outpaced the full deglacial rise in atmospheric CO₂, pointing to an additional driver of early deglacial warming. The rapid and global nature of change suggests a mechanism capable of producing rapid (i.e., within centuries) changes in both hemispheres. Observed LGP to Holocene terrestrial and marine temperature changes are broadly consistent with the differences between those modeled in ‘control’ and ‘wind-shifted’ runs utilizing ESM2M, suggesting that a portion of warming observed at mid-latitudes during HS1 was due to changes in the SHW. Additionally, the ESM2SM experiment that shifted the SHW did not include any additional forcing, and therefore may not represent the full deglacial perturbation. For example, if the WPWP was warming or expanding prior to the SHW shift⁴⁸, the warm waters routed southward by a change in the SHW would bring even more heat than was demonstrated in the model. We propose that reconstructing and modeling changes in the position and strength of the SHW as a mechanism of driving or moderating global climate change should continue to be a priority for future work, particularly in relation to the propagation of changes across latitudes.

Understanding how the atmosphere interacts with the upper ocean and with the rest of the climate system is critical for predicting changes that may occur in the near-term future. During the past four decades, the SHW have been moving poleward^{100,101}. Impacts on Southern Ocean heat and carbon uptake are being monitored and modeled^{102,103} to gain insight into possible trajectories for this critical ocean region. Paleo observations make it clear that large, transient atmosphere-ocean shifts have occurred in the past on a global scale, with strong indications that shifts in the strength and position of the westerlies play a key role in global climate change.

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Author contributions

KAA and ELS conceived of the project and secured funding; CRS and KAA carried out laboratory work; KAA served as primary advisor to CRS during her doctoral dissertation work; ELS and AEP served on CRS's graduate advisory committee; JLR performed all Earth system modeling and provided visualizations of model output; TPG provided guidance on age models and their uncertainties. All authors provided input on the manuscript and participated in discussion and interpretation of results.

Competing interests

The authors declare no competing interests.

Data availability

The data generated in this study are available in the Pangaea repository (<https://www.pangaea.de/>)
Correspondence and requests for materials should be addressed to cassandra.stirpe@gmail.com.
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Methods

Trace element analyses:

For each sample, a total of 6-12 specimens of *Uvigerina peregrina* and *Hoeglundina elegans* were selected from the >250 μm size fraction, with the number of individuals depending on both size and availability. Specimens were cleaned following the methods of Boyle and Keigwin¹⁰⁴ and Rosenthal et al.^{105,106}. Samples were analyzed using a Thermo Scientific Element XR ICP-MS located at Rutgers University. Standard gravimetrically prepared solutions were analyzed throughout the run and used to convert counts-per-second (cps) ratios into molar ratios following Rosenthal et al.¹⁰⁷. Matrix standards were run to account for effects due to differing concentrations of calcium in sample solutions. The average matrix correction was ~4%. Samples were assessed for contamination using Ti/Ca, Mn/Ca, and Al/Ca, with thresholds set to 100, 100, and 500 $\mu\text{mol/mol}$, respectively, for *U. peregrina*, and 2, 2, and 50 $\mu\text{mol/mol}$ for aragonitic species *H. elegans*. Points were also assessed to confirm physically plausible temperatures for this water mass (i.e. above freezing, and below 10°C).

Paleotemperature reconstruction:

Antarctic Intermediate Water (AAIW) temperatures were calculated as follows: for *U. peregrina* samples, temperature was calculated using Mg/Ca and the equation of Stirpe et al.¹⁰: $\text{Mg/Ca (mmol/mol)} = 0.073(\pm 0.037) \cdot \text{BWT} + 0.90(\pm 0.04)$, where BWT is bottom water temperature in degrees Celsius. For *H. elegans* samples, temperature was calculated using Mg/Li and the quadratic equation of Umling et al.¹¹: $\text{Mg/Li} = 0.160(\pm 0.010) + 0.0211(\pm 0.0023) \cdot \text{BWT} - 0.0003(\pm 0.0001) \cdot \text{BWT}^2$. A Monte Carlo simulation with 10,000 iterations was performed with random sampling within the reported $\pm 2\sigma$ uncertainty for calibration coefficients and analytical error for trace element values to determine the mean and 95% confidence interval for the temperature record (see Supplementary Methods). Because *U. peregrina* and *H. elegans* prefer different environmental conditions, they are not often present in the core during the same time intervals. However, where their records do overlap, temperature reconstructions are within error (2σ) and show lines no apparent offset (Fig. 2a).

Stable isotope analyses:

Samples of 1-3 individual foraminifera were selected from the $>250 \mu\text{m}$ fraction, rinsed with ultrapure water, and briefly (<3 seconds) sonicated to remove foreign material. Carbon and oxygen stable isotopes were analyzed at the University of Florida in the Light Stable Isotope Mass Spectrometry Laboratory in the Department of Geological Sciences. Samples were analyzed using a Kiel III carbonate preparation device coupled with a Finnigan-Mat 252 isotope ratio mass spectrometer. Each batch of samples was measured with eight NBS-19 and two NBS-18 standards. Analytical precision averaged 0.02 for $\delta^{13}\text{C}$ and 0.07 for $\delta^{18}\text{O}$. For a composite $\delta^{18}\text{O}$ suitable for age-point analysis (see Supplementary Methods), *H. elegans* values were aligned to *U. peregrina* values by subtracting 0.54, which corresponds to the observed by Jöhnck et al.¹⁰⁸. Although there is an offset between the two species in our composite record, as expected based on previous studies^{108,109} both species follow the same pattern of gradual isotopic depletion that spans the deglaciation (Fig. 2b).

Age models:

The age model for core RR0503-79JPC is based on 5 geochemically fingerprinted tephra layers^{42,110,111} as well as a sixth tephra layer identified using a peak in the percent coarse fraction and visual confirmation of tephra presence under a light microscope (Supplementary Fig. 3). This sixth layer could only be one of two eruptions that occurred between the identified layers above and below. One possible eruption, the Rotorua (~ 15.7 cal kyr BP), was small and unlikely to have left notable deposits in the Bay of Plenty due to its size and documented distribution pattern toward the northwest of the Taupo Volcanic Zone¹¹². In contrast, the Rerewhakaaitu (~ 17.2 cal kyr BP) was a larger eruption with a wider dispersal, and it has been previously identified in several marine sediment cores in the Bay of Plenty and elsewhere north and east of New Zealand's North Island¹¹³. Because of this, the sixth tephra layer was determined to be the result of the Rerewhakaaitu eruption. Using the stratigraphy provided with all six identified tephra layers, the R package "Bacon"¹¹⁴ was used to calibrate ages using SHCal20¹¹⁵. The same approach was applied to published ages for other marine sediment cores discussed in the text, using Marine20 and local reservoir ages as determined using Calib.org and its marine database.

Earth system modeling:

The Earth system modeling referenced in the text is described fully in Denton et al.²⁶. The model compares preindustrial conditions with a perturbation experiment in which the Southern Hemisphere Westerly winds are effectively shifted to the south, creating a more poleward peak wind stress. The results discussed in this text primarily focus on temperatures at 1000 m and 500 m water depth, corresponding to AAIW and other intermediate waters.

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Figure 1. Site locations. (a) Core locations: RR0503-79JPC (this study), KNR197-3-46CDH⁵⁶, and KNR166-2-26JPC⁵⁵. Antarctic Circumpolar Current (ACC) fronts are marked from north to south¹¹⁷: Subantarctic Front (SAF, green), Polar Front (PF, blue), and Southern ACC Front (SACCF, dark blue), with the area between the Northern and Southern ACC boundaries shaded in gray. Blue shading south of the SAF in the southeast Pacific highlights an important location of intermediate water formation and export¹⁷. (b) New Zealand region with local bathymetry¹¹⁸. Study site RR0503-79JPC lies north of the North Island in the Bay of Plenty. Sites from Figure 3 are also marked, as is the approximate location of the modern Subtropical Front (STF). (c) A zonal cross-section with water properties from $\sim 30^{\circ}\text{S}$ ¹¹⁹ and bathymetry from 35°S ¹¹⁸, with color shading representing salinity. AAIW appears as a dark blue salinity minimum centered at ~ 1000 m depth. The salinity contour of 34.5 psu (containing the core of AAIW) is marked with white dashed lines. (d) A meridional cross-section with water properties from $\sim 170^{\circ}\text{W}$ ¹¹⁹ and bathymetry from 175°E ¹¹⁸. Shading as in (c). Made using Cartopy¹²⁰ and with data from Natural Earth (naturalearthdata.com).

Figure 2. Reconstructed temperature and $\delta^{18}\text{O}_b$ records from 79JPC. (a) Intermediate water temperature (IWT) reconstructed using Mg/Ca (*U. peregrina*, blue squares) and Mg/Li (*H. elegans*, teal circles). Shaded bands indicated 1σ and 2σ uncertainties for the smoothed record. Vertical gray bars represent the intervals Heinrich Stadial 1 (18-14.7 cal kyr BP) and the Younger Dryas (12.9-11.6 cal kyr BP). Late Holocene temperatures determined from coretop measurements at a nearby multicore¹⁰ are indicated near the left axis. Gray triangles indicate tephra layers used for developing the age model^{110,111} at the site. Curves on the left axis show mean glacial (green, 24-18 cal kyr BP) and Holocene (red, 11.5-6 cal kyr BP) temperatures as normal distributions. Blue normal curves show results of breakpoint analyses. Empty symbols indicate points excluded as outliers for the purposes of breakpoint analyses and LGM and Holocene averages (see Supplementary Methods). (b) $\delta^{18}\text{O}_c$, with *H. elegans* values aligned to *U. peregrina* values by subtracting 0.54 from each data point¹⁰⁹. All other features as in (a).

Figure 3. Forcing of Southwest Pacific upper ocean temperatures. (a) Changes in the latitude of the SHW (black) through the deglaciation relative to its modern position³⁸. The record shown is the change reconstructed for the Indo-Pacific, with the SHW located approximately 4° north of their modern position at the end of the LGP. Also shown are changes in atmospheric CO_2 ⁴, and the breakpoint for 79JPC (teal; 18.5 ± 0.7 cal kyr BP). Dashed vertical black line marks the estimated onset of deglacial change for the SHW. (b) AAIW temperature record 79JPC along with SST records from RR0503-87JPC³³ (green diamonds; Bay of Plenty), MD97-2120³⁹ (purple squares; South Chatham Rise), and RS147-GC07²⁸ (blue pentagons; south of Tasmania). Dashed vertical teal line marks the estimated onset of deglacial change for 79JPC. Breakpoints were calculated for all records as they were for 79JPC (Supplementary Methods).

Figure 4. Deglacial records of intermediate water temperature (IWT) discussed in the text. Site RR0503-79JPC (this study) is indicated with a teal line and diamonds. Florida Straits site KNR166-2-26JPC⁵⁵ is indicated with a red line and circles. Demerara Rise site KNR197-3-46CDH⁵⁶ is indicated with an orange line and triangles. Brazil Margin site KNR159-5-36GGC¹¹ is indicated with a violet line and hexagons. Hollow shapes indicate points excluded for the purposes of breakpoint analyses (see Supplementary Methods). Normal curves along the bottom axis show the changepoint for the records that indicates the beginning of deglacial warming.

Vertical gray bars highlight Heinrich Stadial 1 and the Younger Dryas.

Figure 5. Model simulations of modern (“control”) and southward-shifted SHW (“winds”) scenarios. (a), (c), (e), (g), and (i): results from 1000 m water depth; (b), (d), (f), (h), and (j): results from 500 m water depth. In panels (e), (f), (g), (h), (i), and (j), site 79JPC (1165 m, this study) is indicated with a teal diamond, site 46CDH (950 m, Oppo et al., 2023) is indicated with an orange triangle, and site 26JPC⁵⁵ (546 m) is indicated with a red circle. Under the wind perturbation scenario, waters at 1000 m at site 79JPC warm by ~ 1 °C, with similar and greater warming occurring at 500 m in the region where AAIW is subducted in the southeast Pacific. Most of the Atlantic, including sites 26JPC and 46CDH, warms under the wind scenario compared to the control scenario.

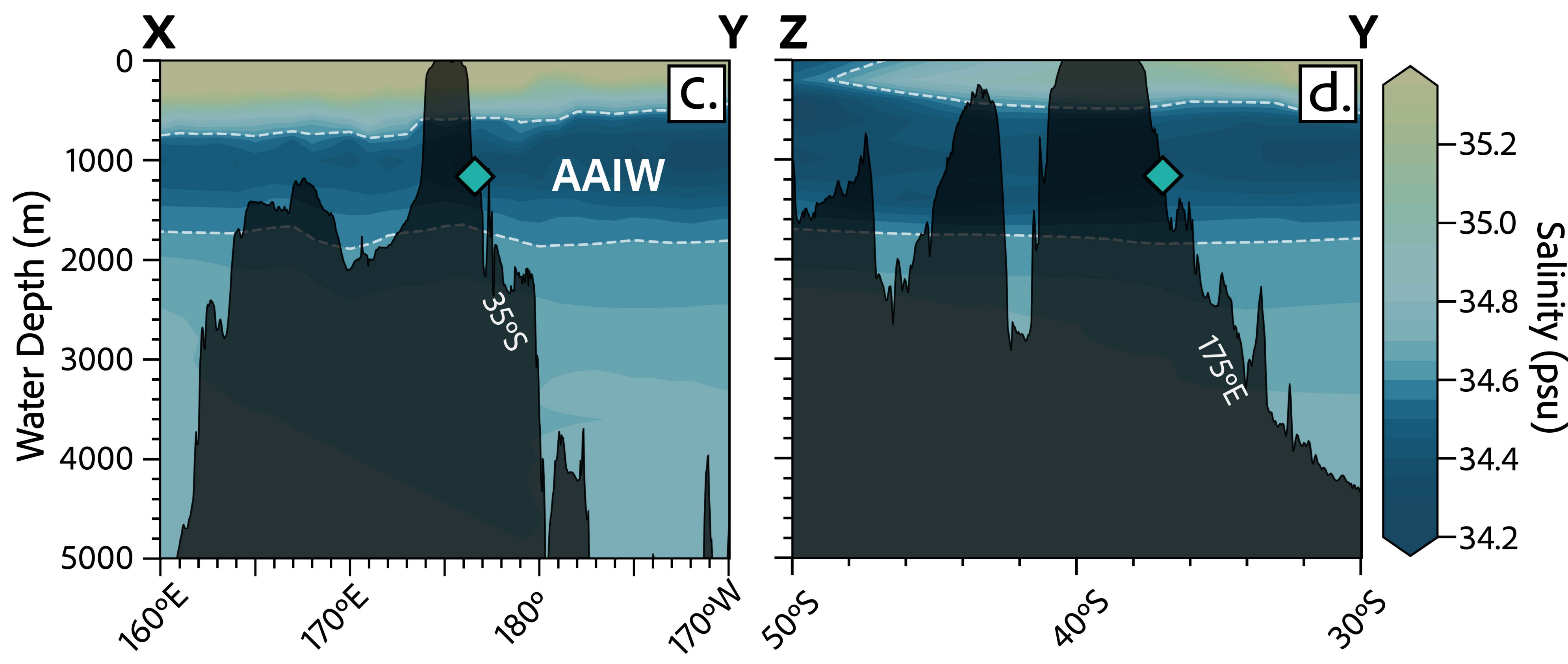
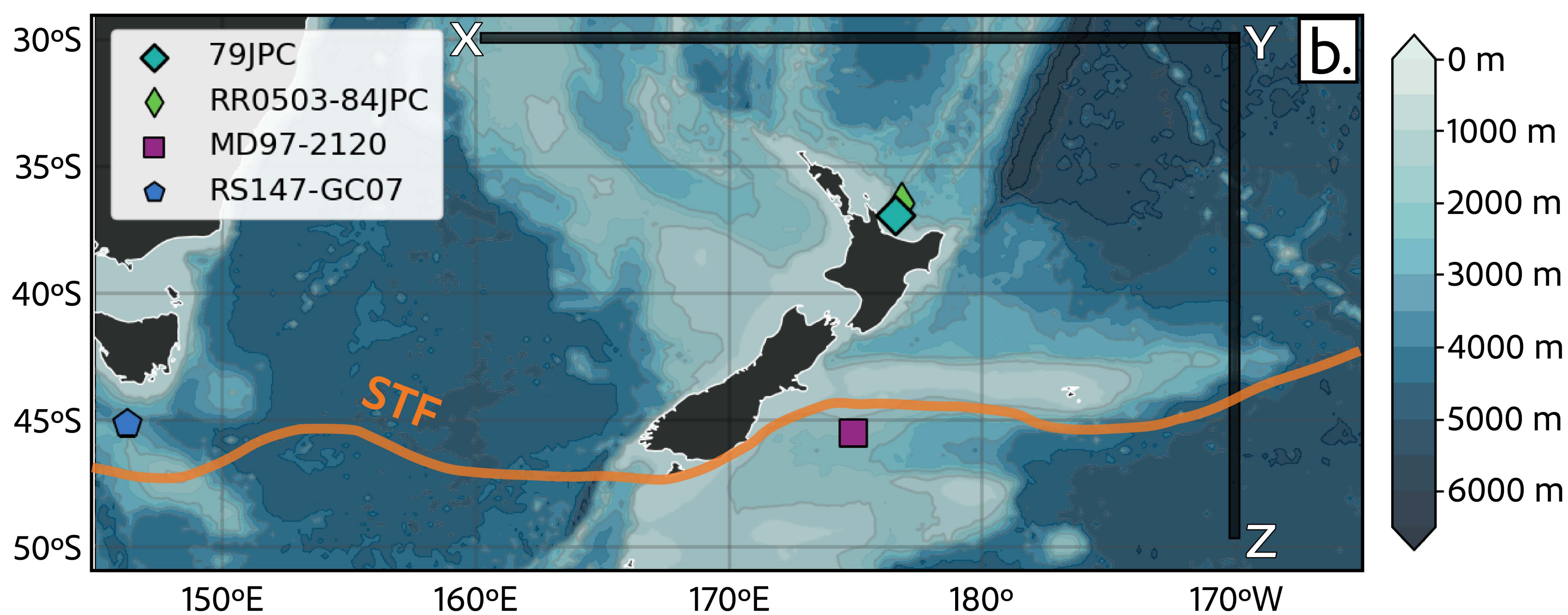
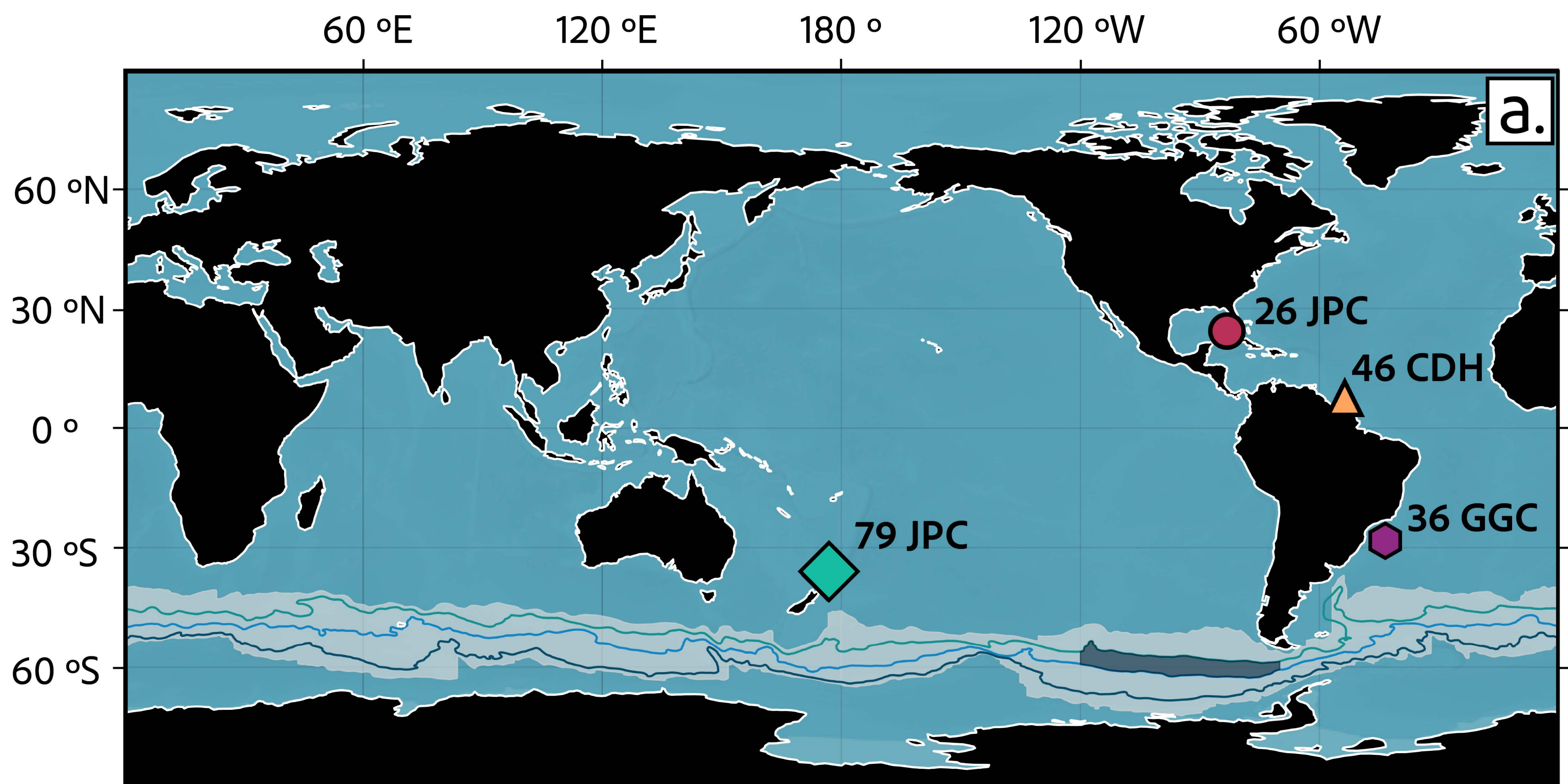
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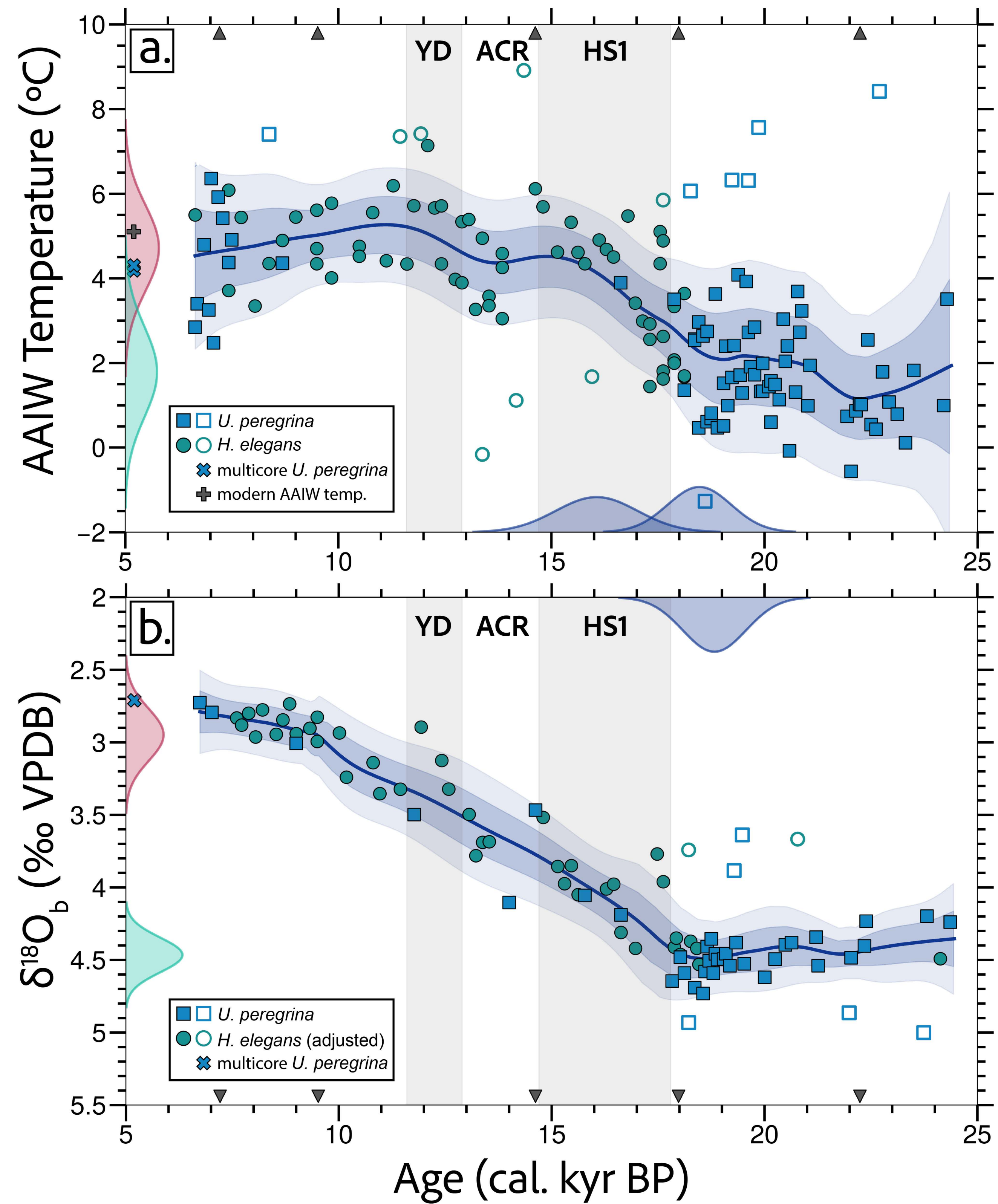
Intermediate waters in the southwestern Pacific Ocean warmed simultaneously with upper ocean waters in the Atlantic during Heinrich Stadial 1, possibly due to a southward shift of the Southern Hemisphere Westerly Winds, according to analyses of benthic foraminifera proxy records.

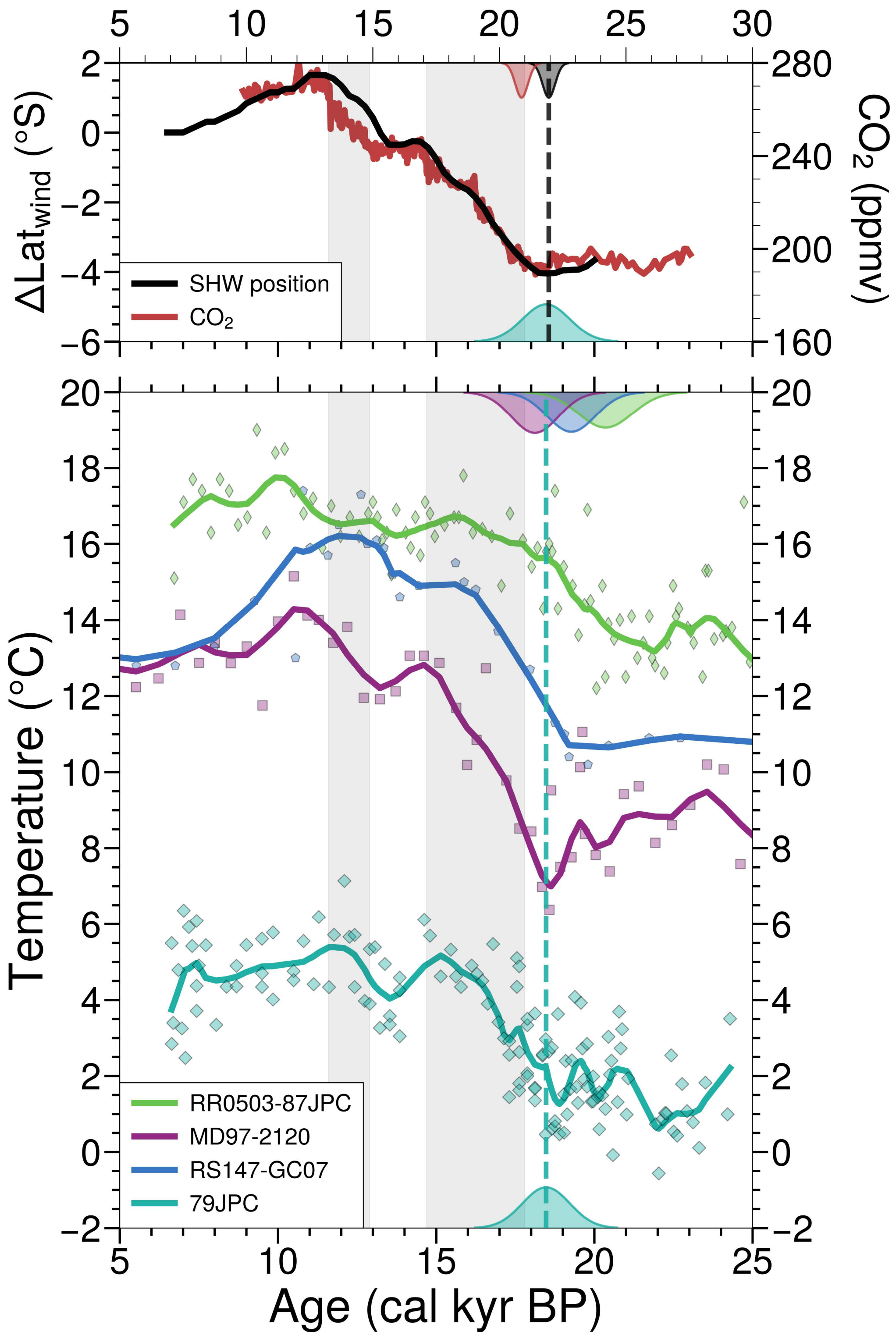
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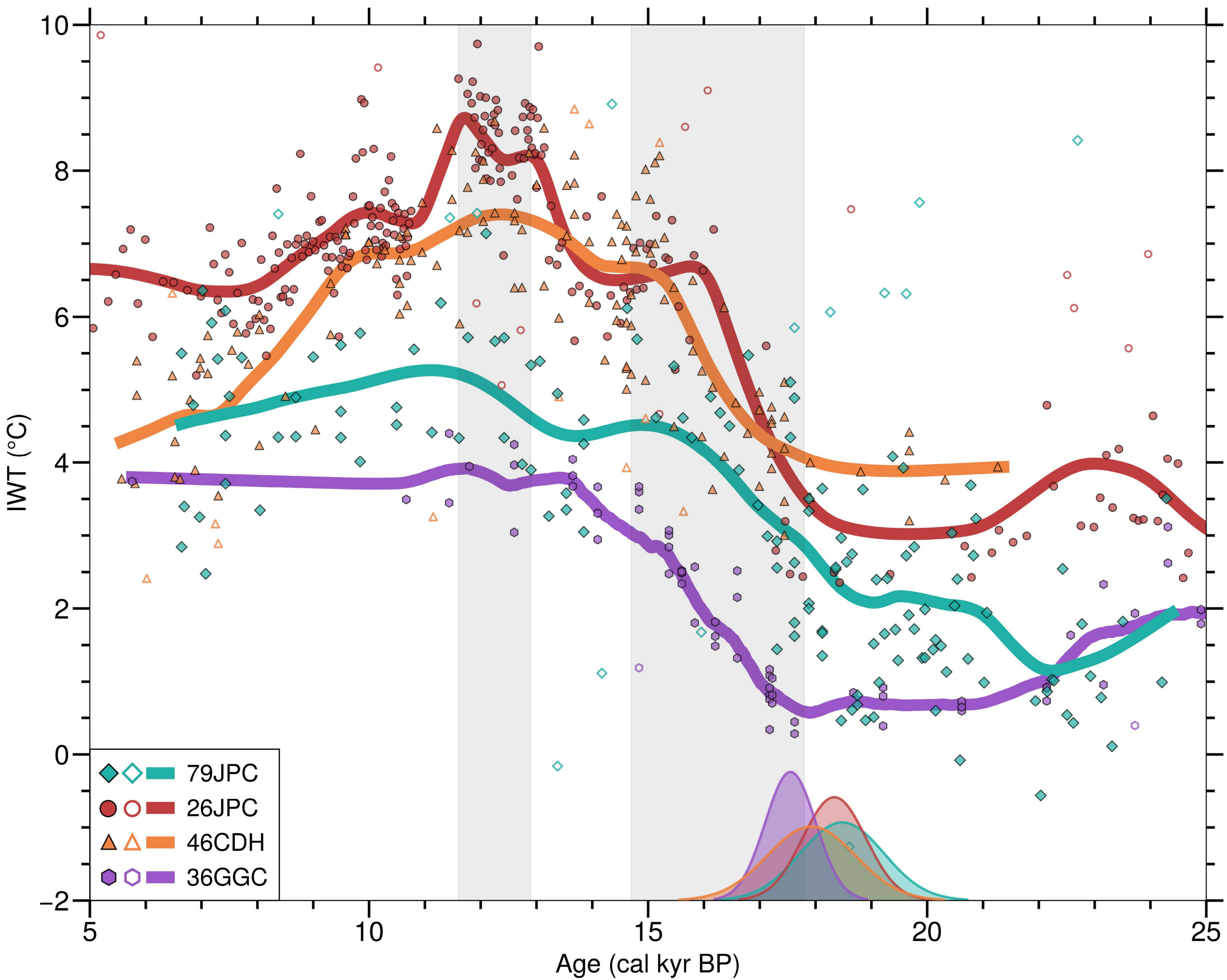
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1000 m depth

500 m depth

