Persistently well ventilated intermediate depth ocean through the last deglaciation

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During the last deglaciation (~18 to 11 ka), existing radiocarbon (^{14}C) reconstructions of intermediate waters in the mid to low latitude oceans show widely diverging trends, with some broadly tracking the atmosphere and others suggesting extreme depletions. These discrepancies cloud our understanding of the deglacial carbon cycle because of the diversity of hypotheses needed to explain these diverging records, e.g., injections of ^{14}C-dead geological carbon, mixing of extremely isolated waters from the abyssal ocean, or changes in sites of

deep water ventilation. Here we present absolutely dated deglacial deep-sea coral ^{14}C records of intermediate waters from the Galápagos Platform – in close proximity to the largest reported deglacial ^{14}C depletions – together with data from the low latitude Atlantic. Our records indicate coherent, well-equilibrated intermediate water ^{14}C-ventilation in both oceans relative to the atmosphere throughout the deglaciation. The observed overall trend toward ^{14}C-enriched signatures in our records is largely due to enhanced air-sea carbon isotope exchange efficiency under increasing atmospheric $p\mathrm{CO}_2$. These results suggest that the ^{14}C-depleted signatures from foraminifera are likely sedimentary rather than water mass features, and provide tight ^{14}C constraints for modelling changes in circulation and carbon cycle during the last deglaciation.

Natural radiocarbon is produced in the upper atmosphere, enters the surface ocean via air-sea gas exchange, and is transported to depth by the ocean's meridional overturning circulation. Since ^{14}C decays away with a 5730 year half-life, the ocean's ^{14}C concentration deficit relative to the atmosphere can be used as a chronometer of ocean circulation and a metric of air-sea gas exchange efficiency. Today, the most ^{14}C-depleted signature is found in the deep Northeast Pacific with Δ^{14}C difference from the pre-bomb atmosphere of ~240 ‰[1]. During the Last Glacial Maximum (LGM, ~22-19 ka), ^{14}C records suggest that large parts of the Pacific and Atlantic were much more ^{14}C-depleted than today, indicative of reduced exchange of carbon between the deep ocean and the atmosphere[2]. This carbon isolation in the deep ocean is thought to account for much of the drawdown of atmospheric CO_2 at the LGM compared to the preindustrial[3], and results from coupled changes in circulation and biological sequestration of carbon in the deep ocean[4]. The widely held view is that deglacial reinvigoration of ocean overturning brought ^{14}C from the surface to the abyss and at the same time released the ^{14}C-depleted "old" carbon from the deep ocean to the upper ocean and atmosphere.

However, the impact of deglacial overturning circulation on upper ocean ^{14}C ventilation is highly controversial. The presence of anomalously old ^{14}C excursions at intermediate depths during deglaciation has been used as evidence for reconnection of an extremely isolated deep carbon reservoir with the upper ocean (i.e., above the main thermocline) and the atmosphere via Southern Ocean upwelling and Antarctic intermediate waters (AAIW) advection[5]. Other suggested pathways of abyssal carbon release have included ventilation in the North Pacific[6-8] or from the Arctic into the North Atlantic[9]. Alternatively, it has also been proposed that ^{14}C-free carbon from sources such as sub-marine volcanism or methane clathrates were injected into intermediate depths of the ocean, for example in the Eastern Equatorial Pacific (EEP), and may have contributed to the atmospheric CO_2 rise during the last deglaciation[10,11]. Robust reconstructions of ^{14}C at intermediate depths are thus crucial in constraining the nature of deep ocean carbon release to the atmosphere, and potential links to changes in ocean circulation and climate.

Deglacial intermediate water ^{14}C reconstructions

Currently, the available deglacial ^{14}C records from the intermediate depth ocean show a much larger range of variability than deep-ocean or atmospheric records. While some records from benthic foraminifera show anomalies of more than 8000 years from the atmosphere[10], others do not contain any discernible episodes of large ^{14}C-depletion[12-14]. Reliable interpretation of ^{14}C ventilation history for sediment-based records (mainly benthic foraminifera) is critically dependent on age model assumptions. As an example, the same set of benthic ^{14}C data could indicate either a constant modern-like ^{14}C ventilation of deglacial AAIW[12] or a significant decrease in benthic $\Delta^{14}C$ by more than 150‰ during HS1 depending on the choice of age model[15]. Notably, the reported ^{14}C data discrepancies are particularly pronounced in the intermediate waters of the EEP (Figs. 1 and 2, Extended Data Fig. 1), yielding different numbers of occurrences, magnitudes, and durations of ^{14}C-depletion episodes over the last deglaciation. Whilst more recent detrital wood-based chronologies avoid the difficulties of variable surface reservoir ages or uncertainties in age correlation to ice-core/speleothem records, wood-based studies from the EEP also show different magnitudes of ^{14}C-depletions[14,16]. These results make it challenging to obtain a consistent picture of upper ocean ventilation and carbon cycle changes.

As an archive of past seawater chemistry, deep-sea corals can alleviate some of the complications from sedimentary processes, such as bioturbation and diagenesis within restricted pore water environments, because they live above the sediment-water interface. In addition, uranium series ages, which can be determined from their aragonite skeletons, are not influenced by radiocarbon surface reservoir ages or tie-point correlation uncertainties. We present uranium series dated ^{14}C records from two sites at ~600m in the EEP and ~1100 m in the low latitude Atlantic. The deep-sea coral based EEP record largely follows the trajectory of the atmosphere during the deglaciation at the resolution of the data (Fig. 2), and the low latitude Atlantic sample set follows a similar pattern (Extended Data Fig. 1b). Specifically, the coral-based $\Delta^{14}C$ reconstruction of the EEP shows that $\Delta^{14}C$ decreased from 170‰ in early part of the Heinrich Stadial 1 (HS1) to 20‰ in the early Holocene. The proportional offset of $\Delta^{14}C$ relative to the contemporaneous atmosphere ($\Delta\Delta^{14}C_{corr,}$ equivalent to previously reported $\Delta^{14}C_{0,adj}$[17], Extended Data Fig. 1, Methods) reveals that our data show no major ^{14}C excursions compared to other published records, some of which show large and variable ^{14}C offsets. When converted to a ^{14}C 'age', the B-Atm age (sample ^{14}C age offset from the contemporary atmosphere) of EEP intermediate waters decreased from ~1300 ^{14}C years in early part of HS1 to ~850 ^{14}C years in the Holocene (Fig. 3e). Much of this change follows the trend expected due to the deglacial increase in atmospheric pCO_2 (Fig. 3a, Methods), which increases the rate of carbon isotope exchange between the atmosphere and the surface ocean[18]. Data from equatorial Atlantic intermediate waters exhibit a broadly similar trend to the EEP, but are consistently better ventilated throughout the last deglaciation by approximately 20-30‰. This difference is small and mostly within the uncertainties of the data (Extended Data Fig. 1b).

Our EEP data are in marked contrast with an existing foraminifera-based sediment record also from the Galápagos platform, at virtually the same depth (617 m) as our coral-based data (Figs. 1 and 2). The foraminifera-based record shows $\Delta\Delta^{14}C_{corr}$ as low as -670‰[10] (Extended Data Fig. 1d). Given that this excursion is too large to be explained by any recorded deep-water signal, upwelling of carbon from abyssal waters was ruled out, and the release of ^{14}C-free carbon from clathrates or nearby volcanic provinces put forward as a possible mechanism[10]. Due to their close proximity and similar water depths, our data provide strong evidence that the hypothesized geological carbon release did not control the ^{14}C content of intermediate waters near Galápagos. More widely, other foraminifera-based records from the low-latitude eastern Pacific near Baja California show variable degrees of ^{14}C-depletion during the deglaciation, though never reaching values as low as the foraminifera from Galápagos. It hence seems plausible to invoke geologic or aged sedimentary-carbon release into pore waters to explain the foraminiferal ^{14}C anomalies at both Galápagos and Baja California. By contrast, some Baja California benthic ^{14}C records (Extended Data Fig. 1c) are aligned with our deglacial deep-sea coral ^{14}C evolution, further arguing against regionally significant deglacial water mass ^{14}C depletions and in favour of localised offsets in pore waters (Methods). Taking a global view, the absence of any discernible episodes of severe ^{14}C-depletion in our precisely-dated intermediate-water records suggests that basin-scale ^{14}C-depletion of upper ocean water masses is unlikely to have been prevalent. The relatively well-equilibrated intermediate ^{14}C signatures are in agreement with model predictions[19] and imply short residence time for carbon in the upper ocean due to global air-sea gas exchange similar to the present[19,20].

Well-equilibrated intermediate water ^{14}C ventilation

Today, the ^{14}C signature of the equatorial Atlantic at ~1100 m is similar to the EEP at around ~600 m (Fig. 1). Both sites are fed partly by AAIW, however, a fraction of ^{14}C-enriched NADW entrains into the Equatorial Atlantic intermediate depths, while ^{14}C-depleted North Pacific water contributes a higher proportion to EEP intermediate depths. Modelling and proxy records from the subarctic North Pacific indicate that unlike the modern day, North Pacific convection may have reached below ~2 km in the early deglaciation[6] between 17 and 16 ka, when AMOC was close to a collapsed state[7,21] (Fig. 3b). Our data from the EEP does not exclude the possibility of short-lived deep convection and deep water formation in the North Pacific[7] that brings ^{14}C enriched waters to the abyssal ocean and subsequently to the intermediate EEP. For example, the intermediate water ^{14}C contents of EEP were enriched and were similar to the equatorial Atlantic at ~16.5 ka (Fig. 3e, Extended Fig. 5). Nevertheless, the generally better ^{14}C ventilated signature in the Atlantic records is observed through most of the deglaciation where we have coral-based data.

The coherent signals in the low-latitude coral records from both oceans suggest that our data are representative of the mean ^{14}C evolution of upper ocean waters, after their long-distance advection and mixing from their high latitude sources[19]. Theory and modelling[18,19] suggest that the deglacial pCO_2 increase would have resulted in a greater overall air-sea CO_2 exchange and more complete $^{14}C/C$ equilibration, with an inverse scaling between surface water ^{14}C reservoir ages and pCO_2. The offsets of our data from this simple scaling (baseline in Fig. 3e, Extended Data Fig. 5) are small, suggesting this equilibration mechanism could explain most of the overall deglacial decline in intermediate water B-Atm age without invoking circulation change. In addition to changes in air-sea disequilibrium in the surface ocean, these small B-Atm ^{14}C age excursions can be interpreted as the signature of changes in the overturning of the deep ocean and its influence on the intermediate waters, with the impact of changing atmospheric $\Delta^{14}C$ propagated into the interior ocean being an additional complicating factor (Methods). For example, the decrease of atmospheric $\Delta^{14}C$ during HS1 (Fig. 2) would have left the deep water reservoirs relatively more ^{14}C enriched at the time of their formation (all else being equal) and thus should result in lower B-Atm ages. On the contrary, our record shows a higher B-Atm age than expected from the pCO_2 corrected baseline for the well-resolved equatorial Atlantic (Fig. 3E, Extended Data Fig. 5) from the early (~18 ka) to late (~15-16 ka) HS1, thus pointing toward increased connection of the intermediate waters with isolated deep waters, following reduced NADW formation[21] with increased North Atlantic surface reservoir ages[22,23], enhanced upwelling of circumpolar deep waters[24] associated with intensified Southern Ocean convection and Southern Hemisphere (SH) westerlies[25], and/or progressively deeper upwelling and ventilation[26]. Subsequently, at the HS1 to B-A transition (i.e., 15-14.6 ka), atmospheric $\Delta^{14}C$ declined more rapidly than the intermediate water records, yielding a trend toward better-ventilated oceanic signatures at low latitudes (Fig. 2a) and possibly recording the influence of enhanced formation of ^{14}C-rich NADW at this time[13,27]. Dedicated ocean modelling work (such as applying the transit-time distribution technique[28]) will be needed to precisely deconvolve the ^{14}C effects of physical circulation change, variable surface ocean disequilibrium, and variable atmosphere ^{14}C propagated into the interior ocean. Our finding that the observed intermediate water $\Delta^{14}C$ in the Atlantic and Pacific largely track atmospheric $\Delta^{14}C$ with a predictable pCO_2-dependent offset will be crucially important in these efforts.

Implications for deglacial carbon cycling

Our study provides unique constraints on oceanic carbon cycle dynamics during the last deglaciation. First, we argue that episodes of anomalous intermediate depth ^{14}C-depletion recorded by benthic foraminifera (Fig. 2, Extended Data Fig. 1) are very likely to be sedimentary rather than global or basin-scale water mass features. The impact of geological carbon on deglacial carbon cycle must therefore be re-evaluated in climate models in light of the new data obtained in this study. Second, due to the precise age control on deep-sea corals, we can resolve a minor excursion in the well-equilibrated intermediate water ^{14}C records during HS1 (i.e., from ~18 ka to 15-16 ka). This signature is indicative of enhanced mixing of relatively ^{14}C-depleted, presumably carbon-enriched deep reservoirs with the upper ocean during the early deglaciation, consistent with CO_2 outgassing of low latitude upwelling zones[29,30], improved ventilation in the deep Southern Ocean[31] (Fig. 3c), and resulting in atmosphere pCO_2 rise (Fig. 3a). Finally, surface ^{14}C reservoir age variability in the mid-low latitude surface oceans is a consequential and much debated source of uncertainty in dating marine sediment cores based on planktonic foraminifera ^{14}C measurements. Given that surface and thermocline waters are linked by high latitude upwelling and that surface waters have more carbon isotope exchange with the atmosphere than the intermediate waters at any time, the limited apparent ventilation age changes in our coral-based data would require that surface ^{14}C reservoir age variability in the mid-low latitudes should also be limited during the last deglaciation, unless there were substantial changes in local subsurface upwelling. Such understanding is consistent with recent surface (0-100 m) reservoir age simulations which show local variability of less than 500 years in the mid-low latitudes (e.g., 40° S- 40° N) over the last 20 kyr[32]. Overall, our precise reconstruction of intermediate water ^{14}C changes yields powerful constraints on mixing between the deep and upper ocean as well as the ocean carbon cycle at the end of the last ice age.

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Competing interests The authors declare no competing interests.

Figure captions

Fig. 1 Map of sample locations and estimated distribution of pre-bomb ^{14}C concentration. (a) Map showing the selected Pacific-Atlantic transection. (b) Estimated pre-bomb ^{14}C activity of the modern time at water depth of ~700 m[1]. (c) Estimated pre-bomb ^{14}C activity of the Pacific-Atlantic transection[1]. Symbols are sites of this study and existing ^{14}C reconstructions in the mid-low latitudes. Solid arrows indicate circulation of the major global water masses[33]. PDW: Pacific Deep Water; SAMW/AAIW: Subantarctic Mode Water/Antarctic Intermediate Water; UCDW: Upper Circumpolar Deep Water; NADW: North Atlantic Deep Water; AABW: Antarctic Bottom Water.

Fig. 2 The Δ^{14}C records of the eastern Pacific over the last 20 ky[5,10,12-14,34-38]. IntCal13 shows the atmosphere Δ^{14}C evolution with ±2σ uncertainty[39]. Symbols are the same as in Fig.1. 2σ error ellipses of published data and detailed data comparison can be found in Extended Data Fig. 1. YD: Younger Drayes; B-A: Bølling-Allerød; HS1: Henrich Stadial 1.

Fig. 3 The evolution of B-Atm age of our coral records together with other paleoclimate reconstructions. (a) Atmosphere CO_2 concentration[40]; (b) $^{231}Pa/^{230}Th$ ratios of sediment core OCE326-GGC5 from the deep subtropical North Atlantic[21]; (c) Sedimentary authigenic U concentrations as a deep water oxygenation proxy from a deep Southern Ocean sediment core (TN057-13PC)[31]; (d) B-Atm age evolution of UCDW recorded by deep-sea corals of Drake Passage[13,41]; (e) B-Atm age evolution of low latitude intermediate waters (this study). Dashed green and pink lines represent the scenario with atmosphere pCO_2 as the only factor affecting ^{14}C reservoir age for the EEP and equatorial Atlantic intermediate waters, respectively (Methods). Ellipses and bars show the 2σ uncertainties of the data points. Uncertainties are not shown in case they are smaller than the symbols. B-Atm age: the ^{14}C age difference between sample and the contemporary atmosphere. UCDW: Upper Circumpolar Deep Water.

METHODS

Materials and analytical methods. The deep-sea corals were collected from the Galápagos platform in the Eastern Equatorial Pacific by both dredging and Remotely Operated Vehicle (ROV) and the low latitude Equatorial Atlantic using by ROV[13,42,43] (Extended Data Figs. 2, 3, 4). Under modern circulation, the studied two sites are in part fed by intermediate/mode water originating from the Southern Ocean, with some inputs from the North Atlantic and North Pacific, so they are well suited to testing hypotheses that link the various pathways of upper-deep ocean mixing and mean state of the upper ocean ventilation. Aragonite deep-sea corals are insensitive to sedimentary processes because they live above the sediment-water interface. Therefore, deep-sea corals can provide a well-constrained ^{14}C activity of bottom water on a precisely dated absolute age scale. In previous deep-sea coral ^{14}C studies of the Southern Ocean[13,41] for example, the generation of ^{14}C temporal evolution is by linking corals growing on different locations and a range of depths, which could potentially incorporate spatial ^{14}C variability in the temporal evolution records. Nevertheless, an obvious strength in this study is that we reconstruct the deglacial ^{14}C ventilation histories using samples essentially growing at the same locations/depths by a single dredge or ROV dive during sample recovery for both oceans. One set of deglacial samples are from the Galápagos Platform recovered from water depth of ~627 m (Fig. 1, Extended Data Figs. 2, 3 and 4, Extended Data Table 1). In addition, we present new data (Extended Data Tables 1 and 2, Extended Data Fig. 6) that fill in a critical gap (late HS1) of an existing ^{14}C record of equatorial Atlantic intermediate waters at water depths of ~1080 m[13]. This composite record is dominated by samples from a single ROV dive at the eastern Atlantic Carter Seamount (EBA), with only a few data points from similar water depths at other seamounts[13] without influencing our data interpretation (Extended Data Fig. 3). Therefore, our samples yield two well resolved, location-bias free, ^{14}C records since the LGM. In both the Pacific and Atlantic, the intermediate waters sampled by our coral-based datasets are part of the upper ocean circulation system and should therefore provide reliable constraints on the release of ^{14}C-depleted carbon from either the deep ocean or geologic sources, as well as on the dissipation of that carbon through the upper ocean and atmosphere.

The deep-sea coral from the Galápagos platform are mostly large colonial coral fragments with chunky dense structures. No sign of boring holes or alterations were observed after the samples were cut and cleaned. U-Th ages of the EEP deep-sea corals have been analyzed previously[44]. The intermediate water ^{14}C record of the Equatorial Atlantic is partly based on the published data[13] (i.e., group B of the coral samples in that study). In our study, we have analyzed more samples mainly from the late HS1 based on the age screening results using the LA-MC-ICPMS method developed in the University of Bristol[45]. For U-Th dating of the new coral samples of the equatorial Atlantic, we have followed the method described previously in the same lab[13] and will not be reiterated here. One advantage of the U-Th ages of the coral samples published[13] and new ones in this study is that they were processed in the same way, and therefore no systematic error is expected between different sets of samples. The ^{14}C data of this study were analyzed by the new AMS facility recently installed in the University of Bristol, while coral ^{14}C data published in Chen et al.[13] were measured in the AMS lab of the University of California, Irvine (UCI). We have re-measured some coral samples that were previously analyzed in UCI. The results (Extended Data Fig. 6, Extended Data Table 3) show that the ^{14}C ages analyzed in Bristol reproduce quite satisfactorily even if coral samples might themselves contain some inhomogeneity. The pretreatment of the coral samples for ^{14}C measurement in Bristol essentially followed the method of UCI. Each coral sample with weight of approximately 15-20 mg was put into a glass tube for acid leach. We have leached the sample to ~10 mg with hot 0.1 N HCl prior to graphitization. After the samples were dried, the samples will react with concentrated phosphoric acid to produce CO_2 which will be transferred into the gas line with helium as the carrier gas to an automated graphitization device. After graphitization, the targets were measured by the MICADAS AMS with acceleration potential of 200 kV. The fossil coral the with ages much older than 50 ka graphitized by the automated device typically give blank ^{14}C ages of 46-50 ka. All data are reported after blank correction with 2 sigma error given in Extended Data Table 1.

Radiocarbon data report. There are three ways to present ^{14}C data in this study. (1) the known-age radiocarbon correction $\Delta^{14}C$, which is expressed as: $\Delta^{14}C_{coral}$ = (Fm × $e^{(\text{calendar age}/8267)}$-1)×1000. (2) To allow direct comparison for the changing atmosphere

^{14}C inventory, deep water ^{14}C is often reported as the B-Atmosphere age, which equals $R_{coral} – R_{atmosphere}$. R_{coral} is the ^{14}C age of corals and $R_{atmosphere}$ is the ^{14}C age of the contemporaneous atmosphere. Error propagation of the uncertainties follows those described previously[13] with a Monte Carlo technique. (3) Offset of deep water $\Delta^{14}C$ from the contemporary atmosphere, which is simply as $\Delta\Delta^{14}C=\Delta^{14}C_{coral}$ - $\Delta^{14}C_{atmosphere}$. However, $\Delta\Delta^{14}C$ will change with a changing atmosphere ^{14}C pool, without the true ^{14}C ventilation change. One useful way to apply ^{14}C as a geochemical tracer is to calculate inventory corrected $\Delta\Delta^{14}C_{corr}$[17] which can be expressed as: $\Delta\Delta^{14}C_{corr} = (\exp(\lambda_{Libby}\times(R_{atmosphere} - R_{coral}))-1)*1000$, where λ is the ^{14}C decay constant calculated from the Libby half-life of 5568 years, $R_{atmosphere}$ and R_{coral} are the ^{14}C age of contemporaraneous atmosphere and coral sample, respectively. Note this metric is functionally the same as what was defined elsewhere in the literature such as $\Delta^{14}C_{atm\ normalized}$[46] and ^{14}C[47]. $\Delta\Delta^{14}C_{corr}$ has not corrected the impact of variable atmospheric $\Delta^{14}C$ propagated into the interior ocean. Atmosphere ^{14}C evolution is taken from the IntCal13 calibration curve[39].

Possible causes of EEP ^{14}C depletions. While there is growing consensus on a more ^{14}C-depleted deep oceans during LGM than today[2,41,47-54], large data scatter is observed at different depths during this period[2]. Radiocarbon data of deep thermocline foraminifera species of the equatorial Atlantic[55] closely track our records, while data of buried deep-sea coral from Brazilian margin show larger scatter[56] (Extended Data Fig. 1d). Deglacial intermediate water ^{14}C data in the mid-low latitude eastern Pacific published over the last 3 decades[5,12,14,34-38,57] show most pronounced variability compared with the deep ocean records (see a recent compilation[58]). It is not yet fully clear what has caused the observed large $\Delta^{14}C$ differences between different records apart from age model uncertainties and bioturbation especially in the intermediate waters. There is indeed pockmark evidence for deglacial releases of clathrates to the overlying water[59], but there is no evidence to support their distinct role in deglacial carbon cycle. Compilation of global occurrences of seep carbonate formation over the last 2 glacial cycles instead implied that enhanced clathrates release occurred during warm high-sea level stands[60]. Regarding hydrothermal carbon contribution to the EEP region, it is likely that diffusion of old carbon from depth did not inject in large quantities to the overlying water mass but remained in the pore waters and at sediment-seawater interfaces, causing large excursions in the benthic foraminifera records without greatly affecting bottom waters. Indeed, negative deglacial excursions of $\delta^{13}C$ are observed in benthic foraminifera records of the EEP with different durations and magnitudes likely linked to pore water chemistry[11,61]. However, it is not possible to reconstruct seawater $\delta^{13}C$ based on the stable carbon isotopes of scleractinian corals because they are strongly regulated by biological vital effects[62]. Therefore, our study is unable to provide independent evaluation from the coral stable isotope perspective. Other possibilities such as diagenetic overprint[63] or species effect[38,64] are also worth investigating, but they appear not to be the fundamental causes for benthic foraminifera ^{14}C depletions in the EEP[10,11].

pCO_2 effect. pCO_2 effect describes the phenomenon that atmospheric carbon isotopes exchange more slowly with the seawater when the atmosphere CO_2 concentration is lower[18]. This might result in an increase of surface ^{14}C reservoir age by ~250 years, which is then rapidly propagated into the intermediate waters, during the LGM compared to the modern, even when ocean circulation remains invariant. It is important to take this effect into account for our precisely dated coral samples that aim to track nuanced changes in the ocean circulation induced ^{14}C variability of intermediate waters. Rather than introducing new metrics of the ^{14}C to consider the pCO_2 effect, we simply construct two curves as shown in Fig. 3e with Holocene B-Atm ages of 700 and 900 years, respectively. We then assume all else being equal, and the pCO_2 effect on ^{14}C reservoir age calculated as: $\text{B-Atm}_{t} = \text{B-Atm}_{\text{Holocene}} * (pCO_2)_{\text{Holocene}} / (pCO_2)_t$[18]. We also have calculated the ^{14}C age offset of each coral record from the two baseline curves, respectively, which is shown in Extended Data Fig. 5.

Effect of variable atmospheric $\Delta^{14}C$ propagated into the interior ocean. The impact of changing atmospheric $\Delta^{14}C$ on initial ^{14}C content of deep waters at the time of their formation makes it challenging to deconvolve ocean circulation changes from small B-Atm ^{14}C age excursions. It is interesting to explore whether the small variability in ventilation age of intermediate waters still holds during the last deglaciation when incorporating the influence of variable atmosphere $\Delta^{14}C$ propagated into the interior ocean. The projection age technique attempts to measure the time-lag between the entrainment of surface source waters down to greater depths and the time that this $\Delta^{14}C$ is recorded by benthic organisms (e.g., corals or foraminifera, see graphic illustration by Cook et al.[17]). The projection age thus has the potential to take into account the impact of variable atmosphere $\Delta^{14}C$ propagated into the interior ocean. We have calculated the projected age (Extended Data Fig. 7) of the intermediate waters to a hypothetical well-equilibrated surface source (i.e., Marine13 calibration curve[39]). In reality, changes in the projection age should reflect combined effects of source water aging and mixing in the deep ocean, as well as variability of the air-sea exchange disequilibrium in the surface ocean. In our study, a higher projection age could mean more isolated, 'older' source waters supplying the intermediate layers, or reduced surface air-sea carbon isotope exchange of the source waters before subducting into the intermediate layers, or a combination of these two effects. Overall, the calculated projection age (Extended Data Fig. 7) has a quite small variability and shows decline by only a few hundred years during the last deglaciation, in corroboration with the understanding based on 'B-Atm age' without considering the impact of variable atmospheric $\Delta^{14}C$ propagated into the interior ocean. In addition, the increasing projection age of the well-resolved Atlantic record from the early (~18 ka) to late HS1 (~15-16 ka) will reinforce our argument on increased mixing of isolated deep waters into the intermediate layers during this period as discussed in the main text. It should be noted that 'projection age' uses a simple assumption about initial ^{14}C signatures of source waters and still could not fully account for the complexity of various sources of deep waters with different ages[28] supplying the intermediate waters. Nevertheless, the coherency of the understanding from the evolution of 'B-Atm age' and 'projection age' lends strong support to our interpretation on ocean circulation change during HS1.

Data availability Sample location information, U-series ages, and radiocarbon data that support the findings of this study are available in Extended Table 1-3 and also Mendeley Data (http://dx.doi.org/10.17632/vxrmfch8h9.1). The atmosphere CO_2 concentration records, radiocarbon data, $^{231}Pa/^{230}Th$, authigenic uranium flux cited in this study were previously published in refs.[13, 21, 31, 39-41] and are available in the Source Data. The calculated B-Atm age trend without circulation change as well as projection age of the low-latitude coral samples are also available in the Source Data. Detailed information on published foraminifera age and radiocarbon data are available from recent comprehensive compilation[58] (https://www.ncdc.noaa.gov/paleo/study/21390).

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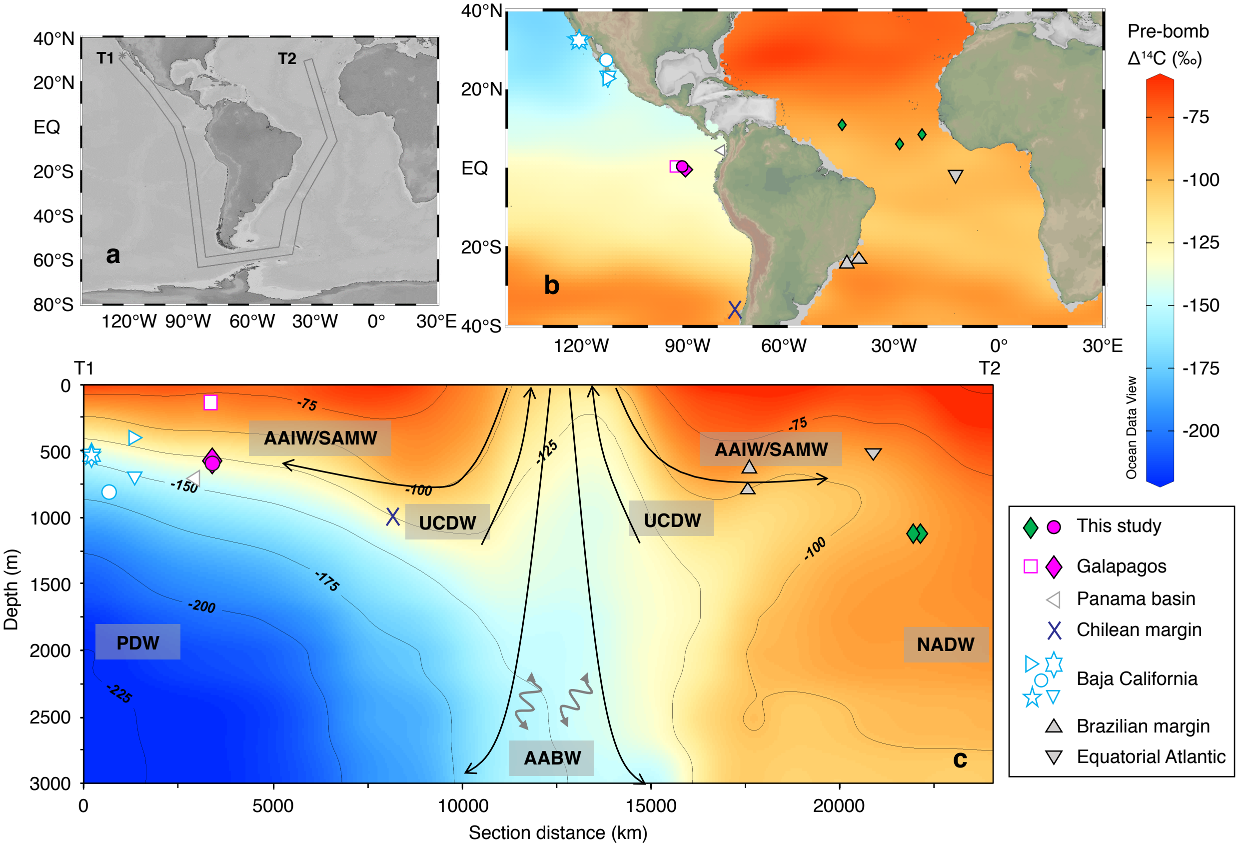
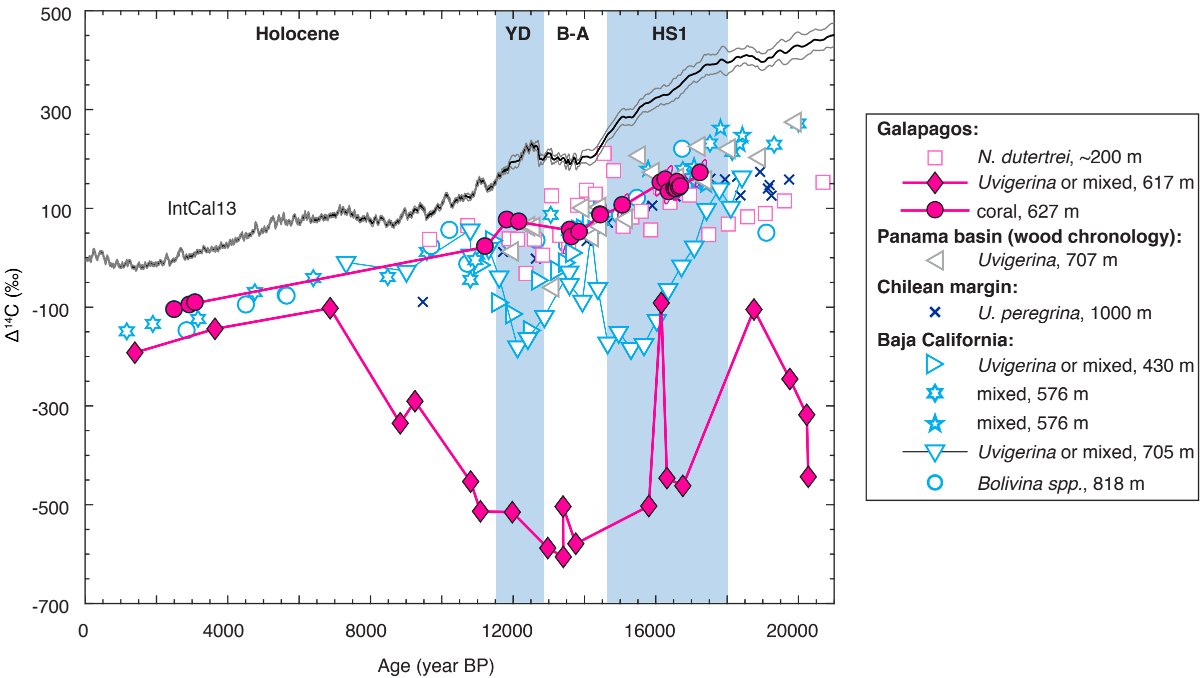
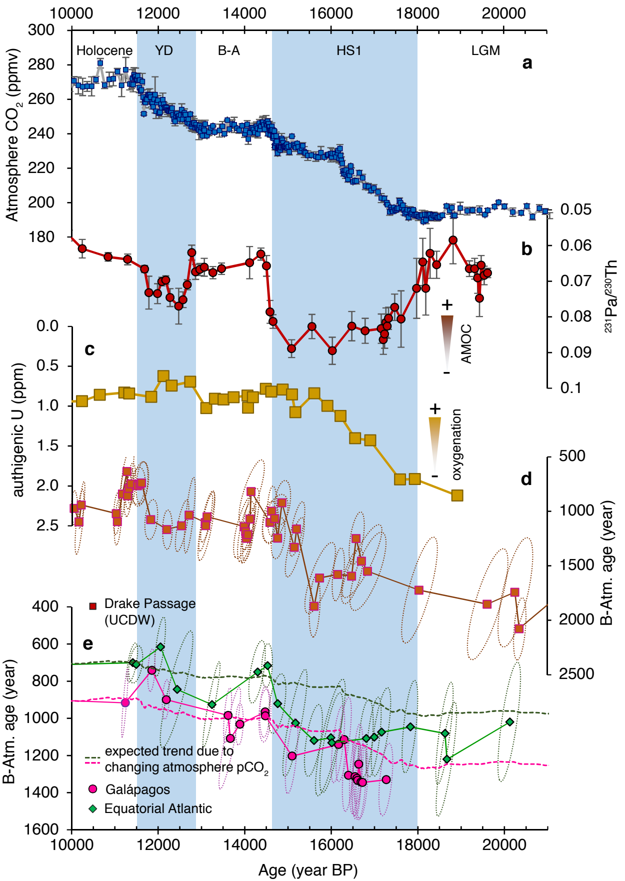
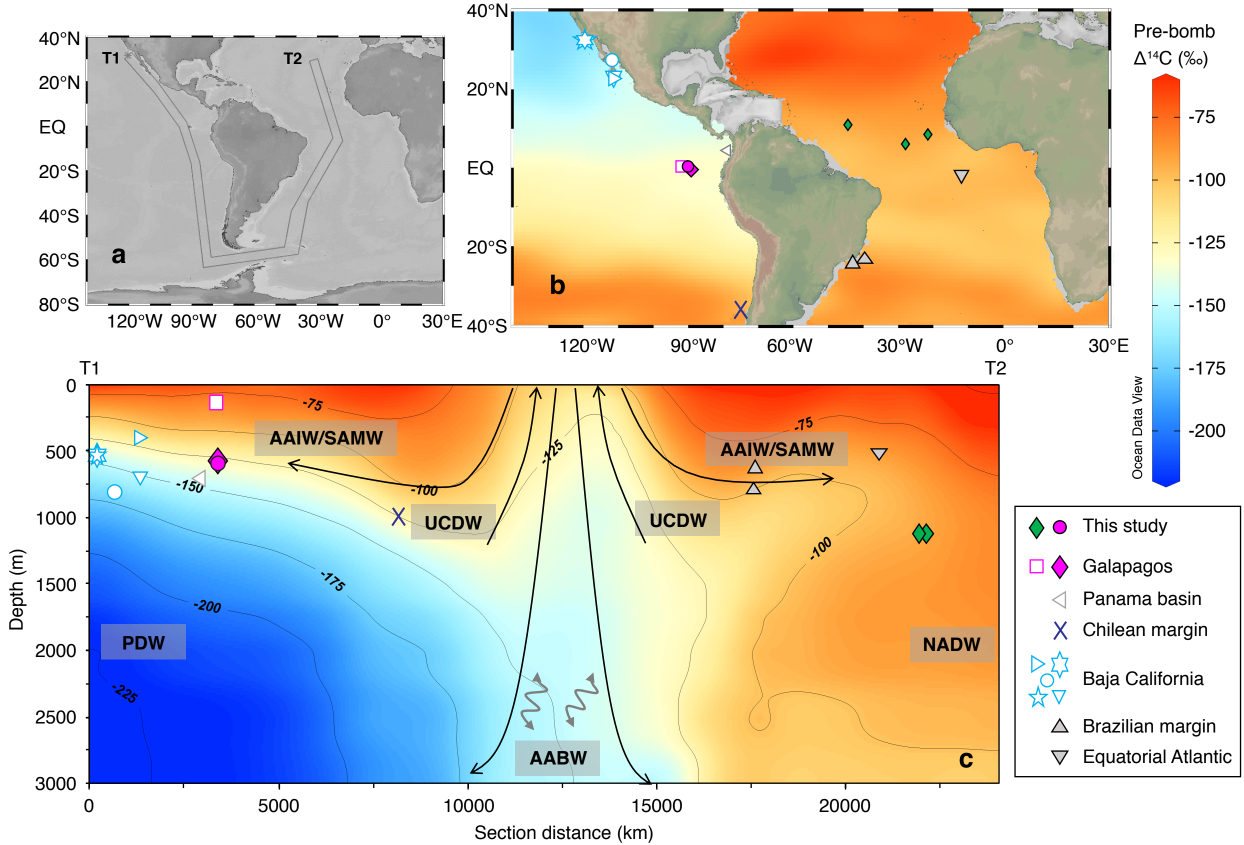


T1
T2
a
b
c
Pre-bomb
Δ14C (‰)
Ocean Data View
AAIW/SAMW
UCDW
AABW
PDW
NADW
Depth (m)
Section distance (km)
This study
Galapagos
Panama basin
Chilean margin
Baja California
Brazilian margin
Equatorial Atlantic

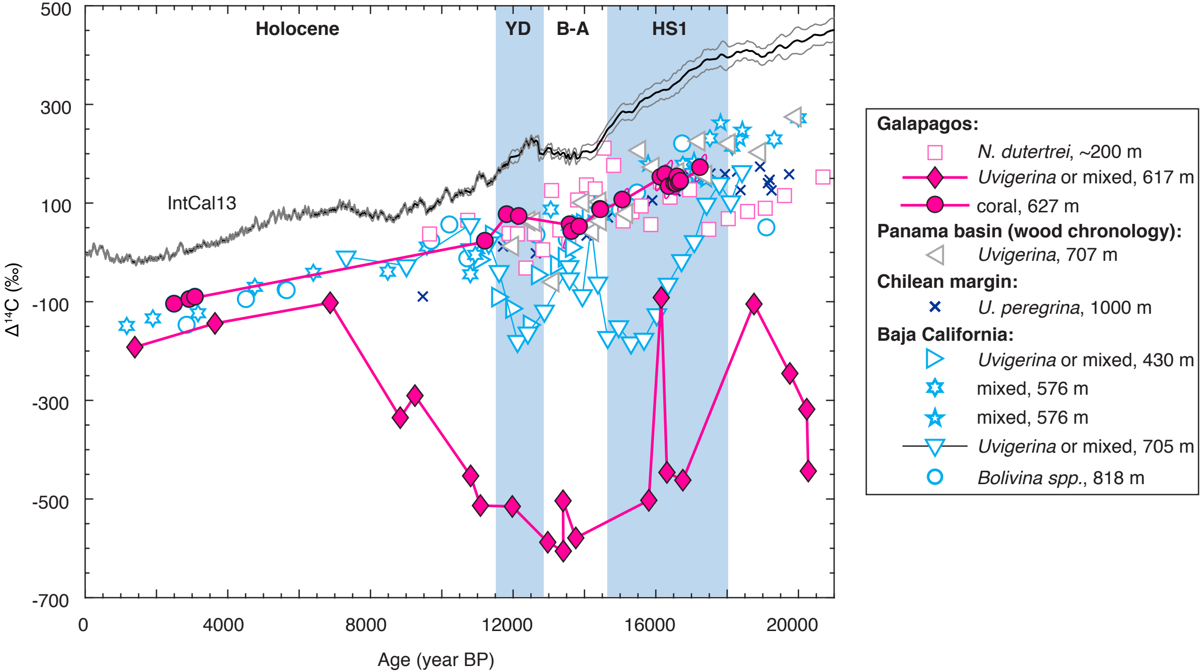


Holocene
YD
B-A
HS1
IntCal13
Δ14C (‰)
Age (year BP)
500
300
100
-100
-300
-500
-700
0
4000
8000
12000
16000
20000
Galapagos:
N. dutertrei, ~200 m
Uvigerina or mixed, 617 m
coral, 627 m
Panama basin (wood chronology):
Uvigerina, 707 m
Chilean margin:
U. peregrina, 1000 m
Baja California:
Uvigerina or mixed, 430 m
mixed, 576 m
mixed, 576 m
Uvigerina or mixed, 705 m
Bolivina spp., 818 m


10000
12000
14000
16000
18000
20000
Holocene
YD
B-A
HS1
LGM
a
b
c
d
e
Atmosphere CO2 (ppmv)
231Pa/230Th
AMOC
authigenic U (ppm)
oxygenation
B-Atm. age (year)
Drake Passage
(UCDW)
expected trend due to
changing atmosphere pCO2
Galápagos
Equatorial Atlantic
Age (year BP)



a
T1
T2
40°N
20°N
EQ
20°S
40°S
60°S
80°S
120°W 90°W 60°W 30°W 0° 30°E
b
40°N
20°N
EQ
20°S
40°S
120°W 90°W 60°W 30°W 0° 30°E
Pre-bomb
Δ14C (‰)
-75
-100
-125
-150
-175
-200
Ocean Data View
c
T1
T2
Depth (m)
Section distance (km)
AAIW/SAMW
UCDW
PDW
AABW
NADW
-75
-100
-125
-150
-175
-200
-225
This study
Galapagos
Panama basin
Chilean margin
Baja California
Brazilian margin
Equatorial Atlantic



Holocene
YD
B-A
HS1
IntCal13
500
300
100
-100
-300
-500
-700
0
4000
8000
12000
16000
20000
Age (year BP)
Δ¹⁴C (‰)
Galapagos:
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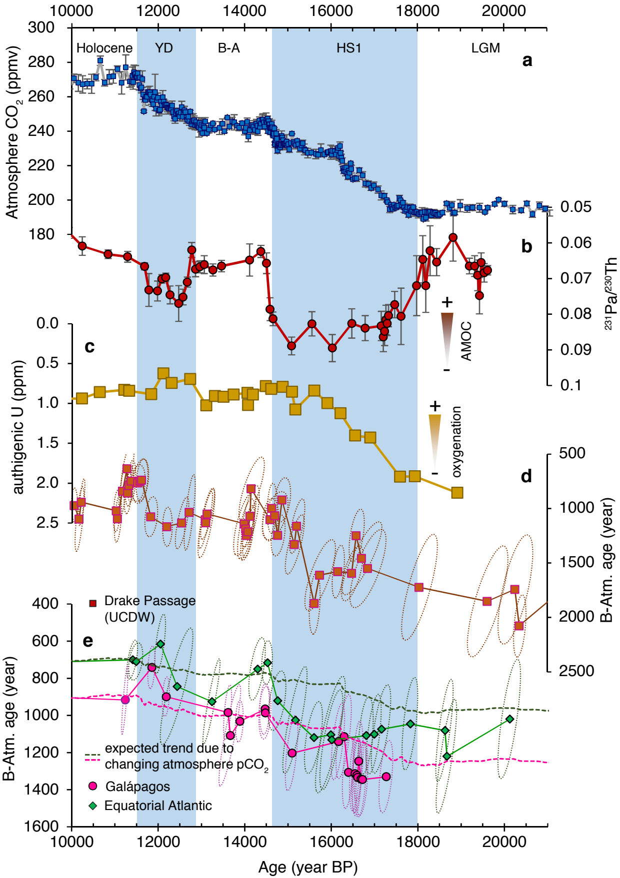

10000
12000
14000
16000
18000
20000
Holocene
YD
B-A
HS1
LGM
a
b
c
d
e
Atmosphere CO2 (ppmv)
300
280
260
240
220
200
180
231Pa/230Th
0.05
0.06
0.07
0.08
0.09
0.1
AMOC
+
-
authigenic U (ppm)
0.0
0.5
1.0
1.5
2.0
2.5
oxygenation
B-Atm. age (year)
500
1000
1500
2000
2500
Drake Passage
(UCDW)
400
600
800
1000
1200
1400
1600
expected trend due to
changing atmosphere pCO2
Galápagos
Equatorial Atlantic
Age (year BP)