Glacial heterogeneity in Southern Ocean carbon storage abated by fast South Indian deglacial carbon release

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Past changes in ocean ¹⁴C disequilibria have been suggested to reflect the Southern Ocean control on global exogenic carbon cycling. Yet, the volumetric extent of the glacial carbon pool and the deglacial mechanisms contributing to release remineralized carbon, particularly from regions with enhanced mixing today, remain insufficiently constrained. Here, we reconstruct the deglacial ventilation history of the South Indian upwelling hotspot near Kerguelen Island, using high-resolution ¹⁴C-dating of smaller-than-conventional foraminiferal samples and multi-proxy deep-ocean oxygen estimates. We find marked regional differences in Southern Ocean overturning with distinct South Indian fingerprints on (early de-)glacial atmospheric CO₂ change. The dissipation of this heterogeneity commenced 14.6 kyr ago, signaling the onset of modern-like, strong South Indian Ocean upwelling, likely promoted by rejuvenated Atlantic overturning. Our findings highlight the South Indian Ocean’s capacity to influence atmospheric CO₂ levels and amplify the impacts of inter-hemispheric climate variability on global carbon cycling within centuries and millennia.

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Past changes in the $^{14}\text{C}$ inventory of the atmosphere cannot solely be attributed to variations in cosmogenic production (Fig. 1). The fraction of atmospheric $^{14}\text{C}$ changes unaccounted for by production changes (referred here to as production-corrected $^{14}\Delta\text{C}_{\text{atm}}$) is believed to reflect large-scale reorganizations of the atmospheric and oceanic respired (i.e. atmosphere-unequilibrated) as well as pre-formed (i.e. atmosphere-equilibrated) carbon inventories with direct implications for atmospheric CO$_2$ ($\text{CO}_2_{\text{atm}}$) levels$^{1,2}$. As $^{14}\text{C}$ is a transient tracer (mean Libby life time $T = 8033\text{ yr}$), ocean-versus-atmosphere $^{14}\text{C}$ disequilibria reflect the rate and efficiency of ocean-atmosphere carbon exchange, and the strength of global-ocean overturning rates and by inference the accumulation of respired carbon in the ocean interior$^4$. Ocean $^{14}\text{C}$ disequilibria are often expressed as ventilation ages that are estimated for instance based on co-existing fossil planktic (surface-dwelling) and benthic (bottom-dwelling) foraminifera, which are believed to faithfully capture the $^{14}\text{C}$ activity of the water mass they grew in. However, despite a number of $^{14}\text{C}$ ventilation age reconstructions, important aspects of the past global carbon cycle remain yet unresolved, such as the location and extent of the glacial carbon-rich ($^{14}\text{C}$-depleted) reservoir that may explain glacial CO$_2_{\text{atm}}$ minima$^{5,2}$, the likely carbon transfer pathways during the last deglaciation$^{6-9}$, and the rates of oceanic carbon release (uptake) and associated CO$_2_{\text{atm}}$ increase (decrease)$^{4,10}$.

Intervals of deglacial CO$_2_{\text{atm}}$ rise coincide with cold spells in the northern-hemisphere, i.e. Heinrich stadial (HS) 1 and the Younger Dryas (YD), as well as a concomitant rise in Antarctic air temperature (the thermal bipolar seesaw; Fig. 1). Both phases are interrupted by stagnating CO$_2_{\text{atm}}$ levels, and warm climate conditions in the North Atlantic, i.e. the Bølling Allerød (BA) interstadial, yet cooling conditions in the Southern Ocean, i.e. the Antarctic Cold Reversal (ACR; Fig. 1). The main control of rising deglacial CO$_2_{\text{atm}}$ levels (and concomitant decreasing $^{14}\text{C}_{\text{atm}}$ levels) on millennial timescales is thought to be the upwelling of deep CO$_2$-rich water masses to the Southern Ocean surface$^{2,5,7,9}$. Southern Ocean upwelling is a highly localized process today that is favored by the interference of the Antarctic Circumpolar Current (ACC) with local bathymetry in regions often referred to as upwelling hotspots$^{11,12}$. One of these important regions is located in the South Indian Ocean, where the ACC impinges on the Kerguelen Plateau (Fig. 2, Supplementary Fig. 1), causing elevated vertical mixing rates, enhanced cross-frontal exchange and efficient export of subsurface waters to the north$^{11,12}$. Despite the significant leverage of the Kerguelen Island area and similar regions on carbon partitioning between the deep ocean and the atmosphere today (Fig. 2), little is known about their ventilation history and impact on deglacial CO$_2_{\text{atm}}$ variations.

Constraints on the past evolution of these upwelling regions can be gained by proxy-based ocean ventilation and oxygenation reconstructions. However, limited carbonate preservation, low abundances of foraminifera and/or uncertainties related to temporal changes in surface-ocean reservoir ages$^{13}$ can pose...
significant challenges to reconstructing deglacial marine carbon cycling in these regions. In addition, sedimentation in these areas can be highly dynamic with common occurrences of high-accumulation drift deposits\textsuperscript{14,15}, often preventing robust age models to be developed. In particular, stratigraphic alignments of sedimentary iron concentrations to Antarctic ice-core dust as recently employed for the Southwest Indian Ocean\textsuperscript{16} may be problematic as the lithogenic fraction may likely not be solely of aeolian origin\textsuperscript{17,18}. Many of these limitations were circumvented through paired uranium-series- and \textsuperscript{14}C-dated corals in the Drake Passage upwelling hotspot region\textsuperscript{2} complemented by deep-ocean \textsuperscript{14}C ventilation reconstructions from the South Atlantic\textsuperscript{5}. The data show a strengthening of upwelling and deep convection in this region from the last ice age until 14.6 kyr before present (BP), which likely contributed to the observed CO\textsubscript{2,atm} rise during HS1. However, given the challenges and potential shortcomings mentioned above\textsuperscript{16}, robust estimates of the deglacial evolution of ventilation changes in the deep (South) Indian Ocean are yet limited. While intermediate-ocean \textsuperscript{14}C-based ventilation reconstructions offshore the Arabian Peninsula hypothesize two distinct upwelling events in the Indian sector of the Southern Ocean\textsuperscript{10}, this hypothesis remains untested and ultimately translates into highly uncertain past global carbon budgets associated with the necessity of extrapolations to the deep Indian Ocean\textsuperscript{19,20}.

Here, we circumvent foraminiferal sample size requirements for most conventional accelerator mass spectrometer (AMS) systems (>1 mg CaCO\textsubscript{3}), and reconstruct the deglacial deep-ocean ventilation history of the South Indian Ocean upwelling hotspot of the Kerguelen Plateau (Fig. 2) via \textsuperscript{14}C analyses of small-sized, paired benthic (B) and planktic (P) foraminiferal samples (0.2–1 mg CaCO\textsubscript{3}) from sediment core MD12-3396CQ (47°43.88' S; 86°41.71' E; 3,615 m water depth; Fig. 2) with the Mil-Carbon-DAting-System (MICADAS) at the University of Bern\textsuperscript{20}, combined with multi-proxy bottom water oxygen estimates. We show based on multiple lines of evidence that, while the deep South Indian Ocean was a significant (remineralized) carbon sink during the last glacial, marked glacial interbasin differences in carbon storage existed in particular between the Atlantic and Indian sectors of the Southern Ocean, likely due to more weakly ventilated, yet geochemically distinct varieties of Antarctic Bottom Water (AABW). The dissipation of these regional differences was mediated by a reinvigoration of Southern Ocean mixing during the last glacial maximum (LGM) referenced to preindustrial\textsuperscript{9,10}. Star in both panels shows the location of the study core. Figure modified after ref. \textsuperscript{1}.  

**Fig. 2** Regions of intense interaction of the Antarctic Circumpolar Current with local bathymetry in Southern Ocean upwelling hotspots. **a** Zonal variations in the percentage of upwelling particles transport across the 1000-m water depth surface (averaged between 30–70°S) as obtained in simulations with the Geophysical Fluid Dynamics Laboratory’s Climate Model version 2.6 (CM2.6)\textsuperscript{30}, where particles were released between 1–3.5 km water depth along 30°S. Increased particle transport in the simulations highlights five major topographic upwelling hotspots in the Southern Ocean\textsuperscript{31}. **b** Spatial changes in particle transport in percent across the 1000 m-depth surface, with vectors showing the average speed and direction of ocean currents at mid-depth (1–3.5 km) based on the Global Ocean Data Assimilation System (GODAS) database (https://psl.noaa.gov/data/gridded/data.godas.html) representing the Antarctic Circumpolar Current between 40–60°S. Squares indicate reconstructed deep-water \textsuperscript{14}C ages in the Southern Ocean during the last glacial maximum (LGM) referenced to preindustrial\textsuperscript{9,10}.  

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**Shading:** Symbols: **Vectors:**

**Simulated particle transport (%) across 1000 m (CM2.6)**

**Deep water \textsuperscript{14}C-dated corals**

**Mean current speed**

(1–3.5 km)
Deglacial variations in upper ocean temperatures. Our study takes advantage of a comprehensive age model approach based on a stratigraphic alignment of multi-proxy (sub-)sea surface temperature ((sub-)SST) records and Antarctic air temperature as recorded in Antarctic ice cores. We reconstructed (sub-)SST variations at our study site based on three independent foraminiferal and lipid biomarker proxies (i.e. foraminiferal assemblages\(^{21}\), planktic foraminiferal Mg/Ca ratios\(^{22}\), and the TEX\(^{86}\) paleothermometer\(^{23}\), Methods). All temperature proxy records closely resemble Antarctic temperature variability (Fig. 3), and reconstructed late Holocene values agree well within uncertainties with the present-day annual SST range at the core site (7.2–8.4 °C, 0–50 m average)\(^{24}\) (Fig. 3). We assess (sub-)SST variability independently of the uncertainties inherent to each proxy based on the first principal component (PC1) of all three datasets (Fig. 3), which accounts for 77% of the data variance. Under the assumption of thermal equilibrium between sub-Antarctic and Antarctic temperatures\(^5\), we graphically align PC1 with Antarctic air temperature recorded by \(\delta D\) variations in the Antarctic EPICA Dome C (EDC) ice core\(^{25}\), Fig. 3, Supplementary Figs. 2–4). The obtained tiepoints provide estimates of local surface-ocean reservoir age (\(d^{14}B_{P-Atm}\)) variations (Fig. 4, Supplementary Fig. 5). Using these \(d^{14}B_{P-Atm}\) constraints, we correct and calibrate our 95 planktic foraminiferal \(^{14}C\) dates based on the atmospheric ShCal13 calibration\(^{26}\), and calculate a sediment deposition (i.e. a chronological) model for our study core (Methods, Supplementary Fig. 6).

Surface-ocean reservoir age changes. High-resolution \(^{14}C\) dates (Methods\(^{20}\) obtained from small monospecific planktic foraminiferal samples (Neogloboquadrina pachyderma) in gaseous (0.24–1.07 mg CaCO\(_3\); \(n = 57\)) and graphite form (0.63–1.15 mg CaCO\(_3\); \(n = 7\)) show a steady down-core \(^{14}C\) age increase without age reversals that exceed 2\(\sigma\)-uncertainties (Fig. 3; Supplementary Fig. 5). Gas \(^{14}C\) analyses of small samples are consistent within
100 ± 70 14C yr (n = 7) with those made on larger graphitized samples (0.59–13.75 mg CaCO3; n = 31), despite a sample mass difference up to a factor of ~8 (Fig. 3; Supplementary Fig. 5)20. We estimate δ14R,Atm variations through subtracting atmospheric 14C ages, derived from our 10 calendar (cal.) tiepoints, from our interpolated (high-resolution) planktic 14C age record (Fig. 4, Supplementary Fig. 5). We find that reconstructed δ14R,Atm ages deviate from preindustrial (i.e. prebomb) surface-ocean reservoir ages of 700 ± 150 yr at our study site (Fig. 4, Supplementary Fig. 5)27,28. Specifically, δ14R,Atm values during the last glacial and the early/late deglacial warming intervals were elevated by up to 800 and 200 yr, respectively (Fig. 4). In contrast, the Holocene and the brief deglacial period of the ACR show lower δ14R,Atm values by up to 400 yr (Fig. 4). Our proxy data-based constraints on δ14R,Atm are supported by transient simulations of marine reservoir ages in the study area (80–100°E, 45–50°S; Fig. 4a) that are forced by temporal changes in Δ14C atm (Methods7). Specifically, our glacial reconstructions resemble modelled δ14R,Atm under glacial boundary conditions, while δ14R,Atm values broadly agree with simulations under present-day climate boundary conditions beyond the ACR (Fig. 4a). Although uncertainties in our δ14R,Atm estimates including chronological, analytical and calibration errors (Methods) are nontrivial, changes in surface-ocean reservoir ages at our study site match with proxy-data-based δ14R,Atm increases22. HS1 Heinrich Stadial 1, ACR Antarctic Cold Reversal, BA Bølling Allerød, YD Younger Dryas.
Deglacial deep-ocean 14C disequilibria in the South Indian Ocean. Our study site is bathed by Lower Circumpolar Deep Water (LCDW), underlain by AABW and overlain by Indian Deep Water (IDW, Methods and Supplementary Fig. 1), and is thus ideally located to document the temporal evolution of vertical mixing due to its sensitivity to changes in northern- (emanating from the North Atlantic) and southern-sourced water masses (originating primarily from Addie Coast and the Ross Sea). This can be achieved by determining past changes in deep-ocean 14C disequilibria that are reflected in 14C age offsets between deep water (benthic foraminifera) and the surface ocean (planktic foraminifera) (d14RBP) as well as the contemporaneous atmosphere (d14RBAtnm).

We observe slightly higher d14RBP values at our core site during the last glacial maximum (LGM, i.e. from 23–18 kyr BP) (900 ± 250 yr, n = 6) when compared to the Holocene (last 11 kyr BP: 600 ± 200 yr, n = 14; Fig. 4b). Rapid decreases in d14RBP values occur at the end of the early- and late deglacial (Fig. 4b). Consistent with our d14RBP record, reconstructed d14RBAtnm values are significantly elevated during the LGM (2300 ± 200 yr, n = 6) compared to the Holocene (1100 ± 200 yr, n = 14) (Fig. 4c). They further show an abrupt d14RBP decrease at the onset of the early deglacial warming and rising CO2 atm levels at ~18.3 kyr BP, but an increase to glacial-like conditions shortly thereafter (Fig. 4c). Although this feature is constrained by only one paired 14C measurement, it is independently supported by bottom water oxygen reconstructions, as discussed below. The strongest deglacial change in d14RBAtnm occurs at ~14.6 kyr BP, when d14RBAtnm rapidly decreases by 1500 ± 300 14C yr within 600 ± 400 cal. yr in parallel with a CO2 atm increase32 of 12 ± 1 ppm at the onset of the ACR and BA (Figs. 4c, 5). The deep South Indian Ocean remained well-ventilated during the ACR, when d14RBAtnm ages are lower by up to 500 yr than prebomb values (Fig. 4c). Late deglacial (i.e. YD) warming in the southern high latitudes coincides with a rise of d14RBAtnm ages. At ~11.7 kyr BP, however, d14RBAtnm rapidly drops by 450 ± 250 14C yr within 850 ± 400 cal. yr, paralleling a marked centennial-scale CO2 atm rise30 of 13 ± 1 ppm (Fig. 5). Reconstructed mean late Holocene d14RBAtnm values at our core site (1200 ± 200 yr) agree well with prebomb values (Fig. 4c)25. Numerical simulations forced by changes in air-sea CO2 exchange through transient changes in Δ14C atm and CO2 atm (Methods)29 also broadly agree with our reconstructed deep-ocean ventilation ages during the late deglaciation and Holocene, while they significantly underestimate deep-ocean reservoir ages during the LGM and early deglaciation (Fig. 4c).

Our deep South Indian ventilation ages closely resemble those found further downstream at mid-depth of the Southwest Pacific30,33,34 (1.6–3 km; Fig. 5). In contrast, upstream in the South Atlantic at 3.8 km water depth3 (a site chosen because of a comparable hydrography and methodological approach used for our study core), glacial d14RBAtnm values are found to be much larger by 300–1500 14C yr than at our study site (Fig. 5), which is also reflected in differences in epibenthic δ13C and δ18O records34 (Supplementary Fig. 9). However, d14RBAtnm values in both regions converge and decrease simultaneously (within age uncertainties) at ~14.6 kyr BP and with identical magnitude (~1500 14C yr) (Fig. 5e, f), and share similar variability thereafter. During the subsequent ACR, reduced ventilation ages in both the deep South Indian and South Atlantic Oceans closely agree with intermediate water estimates from south of Tasmania (1.4–1.9 km)33 and with the mid-depth Southwest Pacific Ocean (1.6–2.3 km; Fig. 5e, f)30,33. A similar convergence can be observed at the end of the late deglacial warming interval, i.e. the end of the YD (Fig. 5c, d).

Deglacial bottom water oxygenation changes. The diagenetic precipitation of insoluble (authigenic) U compounds in marine sediments and in foraminiferal coatings is redox-driven, with oxygen-depleted conditions in pore waters favoring the enrichment of U (Methods). At our South Indian study site, changes in bulk sedimentary U levels during the last deglaciation closely parallel reconstructed variations in d14RBAtnm ages (Fig. 6), and are entirely consistent with changes in the enrichment of U compared to Mn in authigenic coatings of foraminifera (Fig. 6).

A more quantitative bottom water [O2] indicator at our study site is provided by the δ13C difference between the benthic foraminifera Globobulimina sp. and Cibicides sp., the Δ13C proxy36. This proxy is thought to reflect the oxygen-driven respiration of organic matter in marine subsurface sediments, and is thus a direct measure of bottom water [O2] (Methods). Applying the most recent Δ13C-[O2] calibration36, we find that bottom water [O2] during the LGM was lowered by 100 ± 40 µmol kg−1 from present-day concentrations (~220 µmol kg−1; Fig. 6f), which is consistent with higher d14RBAtnm values, increased sedimentary U enrichments and higher foraminiferal U/Mn ratios (Fig. 6). In contrast to our d14RBAtnm and bulk sedimentary U records, we do not observe a marked Δ13C-based bottom water [O2] change during the early deglaciation, suggesting that any bottom water [O2] change during this interval must have been confined to within the proxy uncertainty, i.e. 80 µmol kg−1, if it existed at all. Our foraminiferal Δ13C data additionally highlight a rapid bottom water [O2] increase of 50 ± 40 µmol kg−1 at the end of the early deglacial warming at 14.6 kyr BP in parallel with the rapid reductions in d14RBAtnm U levels and foraminiferal U/Mn ratios (Fig. 6).

Discussion

Enhanced deep South Indian Ocean carbon storage during the LGM. An isolated glacial deep-ocean 14C-depleted carbon reservoir that accommodated more respired carbon during the LGM has been proposed to explain the last glacial CO2 atm minimum3. Marine 14C proxy evidence supports the existence of such a reservoir in the Pacific Ocean36,37, in the deep South Atlantic5 and in the deep North Atlantic39. Larger-than-Holocene d14RBAtnm and d14RBP values (Fig. 4), lower Δ13C-derived glacial bottom water [O2] and enhanced glacial U accumulation at our study site (Fig. 6) extend those observations to the Indian sector of the Southern Ocean, which is consistent with recent findings from the Crozet and Kerguelen Plateau regions16. Assuming negligible saturation- and/or disequilibrium [O2] changes, our multi-proxy reconstruction highlights a substantially higher accumulation of respired carbon throughout the deep Indian Ocean basin during the LGM than during the Holocene. The last glacial ventilation age increase at our South Indian core site is about twice that of the global LGM ocean mean4,10 (Fig. 4), similar to observed trends in the Atlantic5 and Pacific33 sectors of the Southern Ocean (Fig. 5). We thus argue that large swaths of the Southern Ocean accommodated an above-average share of the global-ocean respired carbon at the LGM, largely contributing to the observed glacial reduction in CO2 atm through decreased upwelling and ocean-atmosphere CO2 exchange.

The observed ventilation age and Δ13C-derived oxygen changes of glacial South Indian deep waters at our study site are unlike any preindustrial water mass of the deep Atlantic and Indian Oceans, but instead resemble the oldest prebomb deep-water masses of the Pacific Ocean (Fig. 7). As North Atlantic Deep Water (NADW) flow to the glacial Indian Ocean was diminished40, the glacial (South) Indian increase in respired carbon storage may be related to reduced formation and/or overturning rate of southern-sourced water masses (i.e. AABW) and/or reduced air-sea CO2 equilibration in the Southern Ocean during the LGM. Adjustments of AABW were likely closely
linked to a reduction in shelf space around Antarctica owing to advancements of Antarctic ice sheet grounding lines, an expansion of Antarctic sea ice cover, and an associated shift of the mode and locus of AABW formation, from super-cooling underneath shelf ice and brine rejection in polynyas to open-ocean convection off the shelf break during the last glacial.

However, open-ocean convection during the LGM may have been localized and seasonal in nature (possibly facilitated by open-ocean polynyas) in all sectors of the Southern Ocean, and therefore less efficient at ventilating the ocean interior. Physical and biological changes in the South Indian Ocean are crucial in explaining the last glacial increase in respired carbon content, because simulated adjustments in air-sea gas exchange alone cannot explain our proxy data (Fig. 4). Therefore, a possible northward shift of the southern-hemisphere westerly wind belt and northward expansion of Antarctic sea ice may have changed the geometry of Southern Ocean density surfaces, in particular relative to the location of the Kerguelen Plateau, which combined with more poorly ventilated AABW may have curtailed the capacity of the Kerguelen Plateau region to act as a hotspot for the upwelling of CO$_2$-rich water masses to the surface and subsequent air-sea equilibration.

Fig. 5 Deglacial deep-ocean reservoir age variations in the Southern Ocean. a, c, e Benthic foraminiferal $^{14}$C age offsets in MD12-3396CQ (3.6 km water depth, WD) from the contemporaneous atmosphere, $^{14}$R$_{Atm}$ (blue), compared to deep-ocean ventilation ages in the deep sub-Antarctic Atlantic (MD07-3076CQ, 3.8 km WD; upstream, green), on the Chatham Rise (MD97-2121, 2.3 km WD; downstream, orange), in the New Zealand area (1.6-3.5 km WD; downstream, red), and south of Tasmania (corals, 1.4-1.9 km WD; downstream, purple), and b, d, f atmospheric CO$_2$ (CO$_2$ atm) changes. Lower panels zoom in on $^{14}$R$_{Atm}$ variations during specific intervals of rapid centennial CO$_2$ atm increase, i.e. the ~11.7 (c, d) and ~14.8 kyr events (e, f). Lines and envelopes show 1 kyr- (a) and 0.5 kyr- (c–f) running averages and 1σ-uncertainty-/66%-probability ranges. Vertical bars indicate intervals of rising CO$_2$ atm levels (dark bands highlight periods with centennial-scale CO$_2$ atm increases). HS1 Heinrich Stadial 1, ACR Antarctic Cold Reversal, BA Bølling Allerød, YD Younger Dryas.

NATURE COMMUNICATIONS | https://doi.org/10.1038/s41467-020-20034-1
ARTICLE

NATURE COMMUNICATIONS | (2020) 11:6192 | https://doi.org/10.1038/s41467-020-20034-1 | www.nature.com/naturecommunications
Glacial-ocean heterogeneities in Southern Ocean carbon storage. Comparison of our new data with existing palaeoceanographic reconstructions suggests common water mass characteristics in the Indian and Pacific sectors of the Southern Ocean during the LGM, yet significant offsets exist with the deep central South Atlantic (Fig. 5). We can rule out that this observed LGM offset is a methodological artifact. First, a consideration of foraminiferal blanks in the reconstruction of d^{14}R_{P-Atm} at our study site, although critically discussed\textsuperscript{35}, can only explain a small fraction of the observed glacial difference (Supplementary Fig. 7).

Second, glacial d^{14}R_{P-Atm} estimates in South Atlantic core MD07-3076CQ reach values larger than 2000 years during the LGM\textsuperscript{25}, which may be unrealistic according to a new compilation\textsuperscript{48}. Disregarding these extreme d^{14}R_{P-Atm} values (following similar sensitivity tests made by ref. \textsuperscript{8}) reduces but does not eradicate the observed LGM d^{14}R_{P-Atm} mismatch with our South Indian study site (Supplementary Fig. 8). Further supported by glacial offsets in benthic foraminiferal stable isotope records in both d^{14}R_{B-Atm} and Δδ^{13}C-derived [O\textsubscript{2}] values (Fig. 7) to be realistic, and argue...
Fig. 7 Relationship between seawater oxygen concentrations and conventional radiocarbon ages at present-day and in the past. Modern seawater [O₂] levels versus conventional 14C age in (a) the Atlantic Ocean (squares), (b) Indian Ocean (circles), and (c) Pacific Ocean (triangles) below 2 km water depth; modified after ref. 4. Symbol color represents the latitude of the seawater sample. Large symbols show reconstructed bottom water [O₂] (via the Δδ13C proxy) and ventilation ages (i.e. d18O_BW, representing paleo-conventional 14C ages) from the deep South Atlantic (green: MD07-3076CQ, 3.8-km water depth);6, the deep South Indian (blue, this study: MD12-3396CQ, 3.6-km water depth) and the deep Eastern Equatorial Pacific Ocean (black: sediment core TR163-23, 2.7 km water depth;27,28; please note that Holocene Δδ13C proxy data in this core overestimate present-day bottom water [O₂] in the study region by ~80 µmol kg⁻¹). Symbol labels indicate temporal bins over which the paleo-14C-[O₂] data were averaged (in kyr before present (BP), e.g. for 15 kyr BP: 15.99–15 kyr BP). The principal trend of increasing ventilation ages with decreasing seawater oxygen content can be ascribed to the accumulation of respired carbon, while deviations from this trend can be driven by the advection of well-ventilated water masses, e.g. from the Weddell Sea ([O₂] increase without 14C change), or through organic carbon respiration in upwelling regions ([O₂] decrease without 14C change)5. On multi-millennial timescales, the respiration rate may change (causing the 14C-[O₂] slope to steepen or flatten), and the ocean-atmosphere 14C and O₂ equilibration timescales change with varying atmospheric CO₂ levels (i.e. mean reservoir ages increase in a glacial 190 ppm CO₂ atmosphere without [O₂] change)50 and ocean temperature/salinity (i.e. [O₂] saturation increases without 14C change during glacials)51.

Transient early deglacial ventilation increase in the South Indian Ocean. At the onset of HS1, deep-ocean 14C ventilation (Fig. 5) and oxygenation in the South Indian (as indicated by the aU record in MD12-3396CQ, Fig. 6) and in the South Atlantic5 increased until 16.3 kyr BP, signaling a net loss of (reminalerized) carbon from the deep ocean during that time interval. A comparison to simulated deep-ocean reservoir changes in our study region implies that atmospheric 14C variability transmitted into the deep ocean cannot explain the observations (Fig. 4). Instead, this early deglacial ocean-carbon release was likely driven through an early deglacial reduction in Antarctic sea ice cover (promoting air-sea gas exchange)52; increased vertical mixing through AABW reinvigoration54 and -formation below ice shelves and within coastal/shelf polynyas (as shelf space becomes available), and/or a poleward shift/reinvigoration of the southern-hemisphere westerlies and associated Ekman pumping of subsurface waters5. These combined or in isolation might have altered the water column density structure near Kerguelen Plateau in such way that isopycnals increasingly interfered with the local bathymetry, leading to reinvigorated vertical mixing11,12,47 during the first half of HS1.

We hypothesize that increased South Indian vertical mixing during early HS1 would have fueled biological productivity through the supply of nutrient-rich water masses to the euphotic zone of our study region17. This is documented in nearby core MD02-2488 (46°28.8'S, 88°01.3'E; 3420 m water depth) by enhanced accumulation of benthic foraminiferal species indicative of highly seasonal and high surface-ocean productivity as these benthics feed on labile organic matter raining out of the euphotic zone (Fig. 6)55. A transient supply of labile phytodetrital organic carbon to the sediment may also help explain the insensitivity of the Δδ13C-[O₂] proxy as a result of ecological biases56 of C. kullenbergi and/or G. affinis during that time...
The Kerguelen Plateau or underlying bathymetry, causing elevated Kerguelen Plateau. These changes might have also impacted the role in modulating deglacial CO$_2$ through throughout HS1, attributing the South Indian Ocean a unique the Southern Ocean indicates that interbasin differences persisted of the BA warm period, linked to a rapid resumption of Atlantic overturning at the onset (Fig. 6). Our data signal a restratification of the South Indian water column during late HS1, which may be controlled by changes in the ventilation and formation of South Indian deep waters, for instance by increases in AABW salinity, along with a shift of the southern-hemisphere westerlies to south of the Kerguelen Plateau. A reduction in carbon release from the South Indian to the atmosphere during late HS1, likely due to an unfavorable superposition of the South Indian water column density structure with bathymetry around Kerguelen Island, may have halted the early deglacial CO$_2$ rise and promoted the plateauing of CO$_2$ during ~16.3 and ~14.8 kyr BP (Fig. 4). The fact that this late HS1 return to glacial-like conditions in $^{14}$C and O$_2$ ventilation is not observed in the South Atlantic and elsewhere in the Southern Ocean indicates that interbasin differences persisted throughout HS1, attributing the South Indian Ocean a unique role in modulating deglacial CO$_2$ variability.

Flush of the Southern Ocean carbon pool through AMOC reinvigoration. The marked glacial and early deglacial geochemical interbasin heterogeneity abruptly dissipated at the beginning of HS1, when the ventilation age and oxygen characteristics of deep waters in the different ocean basins rapidly approached prebomb values for the first time throughout the deglaciation (Fig. 7). We argue that the fast increase in $^{14}$C and O$_2$ ventilation in the South Indian Ocean at the end of HS1, at ~14.6 kyr BP, is linked to a rapid resumption of Atlantic overturning at the onset of the BA warm period, causing a rapid (decadal to centennial-scale) southward expansion and eastward deflection of NADW towards the Indian Ocean via the ACC. This may have caused a flushing of the deep-ocean carbon pool that is not only limited to the equatorial Atlantic but expanded into the South Atlantic and South Indian Ocean (this study), with remarkable, near-identical ventilation changes in the latter two regions and a much wider spatial impact on the global ocean than previously recognized (Fig. 5a). These flushing events were possibly amplified by a lagged response of sea ice to a rapid shift of the southern-hemisphere westerlies northward, allowing unabated, transient evasion of carbon from the Southern Ocean to the atmosphere and likely more efficient Ekman pumping around Kerguelen Plateau. These changes might have also impacted the geometry of Southern Ocean isopycnals relative to the location of the Kerguelen Plateau or underlying bathymetry, causing elevated mesoscale eddy activity, isopycnal exchange and/or upwelling along isopycnal surfaces in our study area. An associated transient upwelling event of CO$_2$ and nutrient-rich water masses to the surface ocean at the BA onset is supported by a second abrupt local abundance peak of benthic foraminiferal species that reflects upwelling-driven phytoplankton blooms (Fig. 6). Our data hence suggest that a large fraction of the concomitant 12 ± 1 ppm CO$_2$ increase was driven by a rapid loss of carbon from the South Indian Ocean associated with the reinvigoration of Atlantic overturning and wind-driven Ekman pumping. This transient upwelling event also heralded the establishment of the South Indian upwelling hotspot at ~14.6 kyr BP that was akin, if not stronger than its present-day counterpart.

Late deglacial South Indian ventilation decrease and early Holocene convergence. During the YD, we observe decreased $^{14}$C and O$_2$ ventilation at our deep South Indian study site, suggesting that the South Indian remained a moderate source of carbon to the atmosphere, and thus did not significantly contribute to the late deglacial CO$_2$ increase. This agrees with inferences made in other Southern Ocean regions, but disagrees with recent findings from the South Indian Ocean. We show that the disagreement results from insufficiently accounted surface-ocean reservoir age variability that can be improved by applying consistent surface-ocean reservoir ages for all sites (Supplementary Fig. 10). At the end of the YD, we observe a convergence of ventilation ages of different parts of the Southern Ocean, reminiscent of the 14.6 kyr BP-event (Fig. 5, Supplementary Fig. 9). We argue that this convergence may have caused a large fraction of the 13 ± 1 ppmCO$_2$ increase at that time through a rapid loss of carbon from the South Indian Ocean mediated by stronger Atlantic overturning and wind-driven Ekman pumping. This reinforces the role of the South Indian Ocean in centennial-scale CO$_2$ variability during the last deglaciation.

Based on our high-resolution multi-proxy analyses, we identify marked impacts of marine carbon cycling in the South Indian Ocean on glacial and deglacial CO$_2$ Variations. While our new high-resolution data support previous contentions of reduced vertical mixing and restricted air-sea gas exchange in all sectors of the Southern Ocean during the last ice age, they point to marked spatial differences, possibly due to lateral variations in the mode and rate of AABW formation and/or the efficiency of air-sea gas equilibration. We argue that marked increases in South Indian convection during the early HS1 as well as the ends of the YD- and HS1 stadials as shown by both $^{14}$C (d$^{14}$B~ocean~) and O$_2$ ventilation proxies (aU, $\Delta$B$^{13}$C and foraminiferal U/Mn) affected deglacial CO$_2$, variability significantly and rapidly. However, these changes alone cannot explain reconstructed $^{14}$C ventilation ages in the mid-depth North Indian Ocean off the Arabian peninsula during deglacial northern-hemisphere stadials.

Glacial and early deglacial spatial differences in deep-ocean ventilation and oxygenation were entirely eroded along with the southward expansion of northern-sourced water masses at the onset of the ACR, with potentially strong effects on CO$_2$ levels owing to the flushing of old carbon from the deep ocean including the Indian Ocean. We interpret this to mark the onset of the upwelling dynamics akin to the present-day hotspot region
in our study area. The Indian sector of the Southern Ocean, in particular the upwelling hotspot around Kerguelen Plateau, is thus highly sensitive to northern- and southern-hemisphere climate variability and effective in contributing to deglacial CO₂-rise, which needs to be accounted for in global carbon cycle budgets over centennial, millennial and glacial-interglacial timescales.

Methods

Area study. Core MD12-3396CQ was retrieved from the deep Australian-Antarctic Basin (AAB), south of the Southeast Indian Ridge and east of Kerguelen Plateau (Fig. 2), and is located slightly to the north of the Sub-Antarctic Front (47°23.88’S, 66°17.1’E, 3615 m water depth)41. The core site is currently bathed in LCDW, which represents the ACC layer influence from biogenic fluxes from above. First, the alleviation of the topographic control of the ACC passing the Amsterdam–Kerguelen Island Passage caused by decadal-scale warming (and subsequently local) material in the study area165 (Supplementary Fig. 1). Second, vigorous AABW flow in the deep cyclonic gyre in the western AAB leads to erosion and formation of sediment-laden nepheloid layers in the deepest reaches of the western AAB, and subsequent release of these sediments in calmer and shallower flow regimes, for instance at our study site166. Small-grained sediments at our core site (i.e. smaller than foraminifera) were likely subject to transport through currents and redistribution processes, but likely originated from a local source around Kerguelen Plateau, and possibly Crozet Plateau15.

Radio carbon measurements of planktic and benthic foraminifera. Bulk wet sediment samples were freeze-dried, disintegrated in de-ionized water, washed over a 150 µm-sized sieve to remove fine-grained particles and subsequently oven-dried at 45 °C. The planktic foraminifera N. pachyderma and mixed benthic foraminifera (excluding Pyrgo spp. and agglutinated specimens) > 150 µm were hand-picked from near planktic foraminifera abundance maxima (Supplementary Fig. 11), and were 14C-dated with both the MICADAS Accelerator Mass Spectrometer (AMS)–200 kV at the University of Bern (gas- and graphite samples) and with the AMS Petrotron-2.6 MV system at the ARTEMIS 14C laboratory of the University of Paris-Saclay (graphite samples only). Although we sampled our study core at high resolution, age reversals outside analytical uncertainties and outliers are entirely absent. None of our foraminiferal analyses were discarded.

For 14C analyses with the Bern-MICADAS, the benthic and planktic foraminiferal samples were size-matched and pretreated with a weak acid leach prior to complete acidification with orthophosphoric acid. The evolved sample CO2 was then purified with He and directly injected into the gas ion source. We have performed stable isotopic analyses on 14C and 8) between core-top and sediment cores from the study area (Supplementary Fig. 1). As C. wuellerstorfi is often considered a truly epibenthic-living foraminifera, the observed offset can be likely explained by a bias of C. kullenbergi δ13C towards more negative values because of a shallow infaunal habit48. In order to correct for this potential bias, we have performed duplicate 14C analyses on the planktic foraminifera C. kullenbergi. C. kullenbergi δ13C record to more positive values by 0.30 ± 0.30‰ prior to correction for both planktic foraminiferal and inorganic carbon contributions. We observe a significant offset of 0.30 ± 0.30‰ (n = 8) between core-top and bottom water δ13C values. We have performed stable isotopic analyses on 14C and 8) between core-top and sediment cores from the study area (Supplementary Fig. 1). As C. wuellerstorfi is often considered a truly epibenthic-living foraminifera, the observed offset can be likely explained by a bias of C. kullenbergi δ13C towards more negative values because of a shallow infaunal habit48. In order to correct for this potential bias, we have performed duplicate 14C analyses on the planktic foraminifera C. kullenbergi. C. kullenbergi δ13C record to more positive values by 0.30 ± 0.30‰ prior to correction for both planktic and 8C- and 8C-Mn ratios are also linked to bottom water δ13C changes similar to U/Ca ratios90. U/Ca and U/Mn analyses are carried out on oxidatively cleaned 10–60 specimens of the planktic foraminifer N. pachyderma (200–250 µm size) by inductively coupled plasma-mass spectrometry (ICP-MS)93. The reproducibility of replicate foraminiferal U/Ca and U/Mn analyses is within 7.0 mmol-1 and 0.7 mmol-1 (n = 13), respectively.

Bottom water oxygen reconstructions. We have used three approaches to determine bottom water oxygen changes at our core site over the last degl. The first two approaches provide qualitative oxygen reconstructions, and are based on the enrichment of redox-sensitive elements (such as U and Mn) in bulk sediments and in authigenic coatings of foraminifera. The second proxy provides quantitative [O2] estimates based on epibenthic to deep infu-
spectrometer (Agilent, 6130). Mg/Ca ratios were measured on 10–60 oxidatively cleaned specimens of N. pachyderma (200–250 µm fraction) via ICP-MS (ref. 5). They were converted into SST using the calibration Mg/Ca = 0.406 × e0.035T of ref. 22.

Our stratigraphic alignment of (sub-)SST variations at our core site to the EDC δD record is guided by the first principal component (PC1) of all three (sub-)SST datasets that explains 77% of the variance of all three records (Supplementary Figs. 2–4). Our approach makes the assumption of a fast (centennial-scale) thermal equilibration of circum-Antarctic surface waters and Antarctic air temperatures recorded at EDC through an atmospheric connection. We assign an ad-hoc cal. age of 20 kyr to our age markers at more than 20 kyr BP, but use an uncertainty of 1000 yr for tiepoints from the core section older than 20 kyr BP, owing to more subdued temperature variability for that period (Supplementary Figs. 2–4). For our tiepoint at 27.7 kyr, we consider an age uncertainty of 2000 yr, because the variability of PC1 and the reference Antarctic temperature record is low, and all three (sub-)SST records in our study core show strongest disagreement at that time (Supplementary Fig. 4).

We uncalibrated all age markers and converted them into 14C age-space based on the StCali3 calibration59. This conversion considers both the uncertainties of the cal. age tiepoints and the atmospheric calibration, and was performed with the radcal software package60 (Supplementary Fig. 5). Our high-resolution planktic 14C ages (and their 14C age uncertainties) are interpolated onto the depths of all (sub-) SST-based tiepoints. Using the now paired planktic foraminiferal and atmospheric 14C age constraints at our tiepoints, we compute the corresponding surface-ocean reservoir ages, i.e. the 14C age offsets between planktic foraminifera and the atmosphere at that time (i.e. d14C_Aus – P_Sequence). We report median d14C_Aus and the 66% (1σ) probability density range of each sample, as determined by the radcal.R script1. For the topmost sample (at 4.5 cm), we assume a prebomb surface-ocean reservoir age of d14C_Aus = 700 ± 150 yr (Supplementary Fig. 5); the uncertainty of 150 yr estimate incorporates possible centennial d14C_Aus variability in the study region59.

In order to subtract surface-ocean reservoir variations from our measured planktic 14C record, the d14C_Aus estimates were interpolated onto the depths where discrete planktic 14C measurements are available. Our d14C_Aus-corrected planktic 14C ages were then converted to cal. ages using the StCali3 calibration59 and the OxCal (version 4.3.2) program. The final model of the upper core section (<426 cm) with dense age constraints is based on a Bayesian deposition model, obtained by a P_Sequence model within OxCal (P_sequence (‘MD02-3396CQ’, 0.07, 0.2)) and an outlier assessment (Outlier Model (‘General’), τ(5), U(0,4), ‘+’), which considers all calibrated planktic d14C_Aus-corrected 14C ages and their tiepoints resulting from the stratigraphic alignment of (sub-)SST variations to Antarctic temperature (Supplementary Fig. 6). For the lower core section with sparser age constraints (>426 cm), the age-depth relationship was obtained through linear interpolation between our four lowest (sub-)SST-based tiepoints and their ad-hoc age uncertainties (Supplementary Fig. 6). In this study, we focus, however, on the upper 100 cm of the core, which records the last deglaciation and glacial periods with sedimentation rates ranging between 7–35 cm kyr−1 (Supplementary Fig. 4).

We estimate changes in deep-ocean ventilation based on 14C age offsets between benthic and planktic foraminifera (d14C_Ba-Bp) as well as between benthic foraminifera and the contemporaneous atmosphere (d14C_Aus–P-Atm). We use the radcal.R script1 to calculate d14C_Ba-Bp, which considers 10 uncertainties of our chronology (i.e. our sedimentation model), our measured 14C ages and the atmospheric calibration. We report the median and the 66% (1σ) probability density range of our deep-ocean ventilation age estimates as calculated by radcal.R.73

At our study site, we find a non-negligible 14C background of 14C-free foraminifera, with a fraction modern d14C = 0.0007 – 0.0009 for graphitized planktic foraminiferal samples, as well as d14C = 0.0024 – 0.0031 for foraminifera measured in gaseous form10. The foraminiferal backgrounds were determined for the core site with the Bern-MICADAS, but similar estimates from the ARTEMIS facilities are lacking. Given the uncertainties associated with a consistent and reliable foraminiferal blank correction of our Bern-MICADAS and Paris-Saclay 14C data60, we report all ventilation age reconstructions without consideration of foraminiferal 14C backgrounds. However, applying a foraminiferal background correction has a negligible impact on our d14C_Ba-Bp and d14C_Aus–P-Atm estimates during the last deglaciation, as it only slightly affects the absolute values of our results, mostly during the LGM, but not the observed trends (Supplementary Fig. 7).

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**Acknowledgements**

J.G. and S.L.J. (grants PP00P2_144811 and 200021_163003), A.S.S. (grant PBEZP2_145695), and A.M.-G. (grant P200P2_142424) acknowledge funding from the Swiss National Science Foundation. J.G. was also supported by a Global Research Fellowship from the German Research Foundation (DFG grant GO 2294/2-1), and a promotion grant from the Intermediate Staff Association of the University of Bern. A.M. and E.M. acknowledge financial support from the French Ministry of Research and Higher Education, the Swedish Research Council (grant VR-349-2012-6278), and the French National Institute of Sciences of the Universe at the French National Centre for Scientific Research. A.M.-G. further acknowledges funding from the Max Planck Society. M.B. is supported by the German Federal Ministry of Education and Research (BMBF), a Research for Sustainability initiative (FONA, www.fona.de) through the PalMod project (grant number: 01LP1918A). We thank the French Polar Institute Paul-Émile Victor, the captains and the crew of RV *Marion Dufresne* during the Indien-Sud cruises for their help retrieving sediment core MD12-3396CQ. We also thank the staff of the Laboratoire de Mesure du Carbone-14 of the ARTEMIS French National AMS facility, the Laboratory for Radiocarbon Analysis at the University of Bern (especially Michael Battaglia and Gary Salazar), Christopher Bronk Ramsey, Gulay Isguder, Fabien Dewilde, Derek Vance and Corey Archer for technical support. We are indebted to Sophie Hines, Ning Zhao, Jerry McManus, Limin Yu, Andy Hogg, Veronica Tamsitt, Spencer Jones, Danny Sigman, and Wally Broecker for insightful discussions.

**Author contributions**

J.G., E.M., and S.L.J. devised the study. E.M. and A.M. collected the core material. J.G., E.M., L.M.T., A.S.S., A.P.H., N.S., M.B., A.M.-G., and S.S. performed the analyses. J.G., E.M., A.S.S., A.P.H., and N.S. analyzed the proxy data. J.G. wrote the manuscript with contributions from all co-authors.

**Competing interests**

The authors declare no competing interests.

**Additional information**

Supplementary information is available for this paper at https://doi.org/10.1038/s41467-020-20034-1.

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**Peer review information** *Nature Communications* thanks Babette Hoogakker and Patrick Rafter for their contribution to the peer review of this work.

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