Title: Changes in North Atlantic deep-water oxygenation across the Middle Pleistocene Transition

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Abstract: The oxygen concentration of oceanic deep-water and atmospheric carbon dioxide ($p$CO$_2$) are intrinsically linked through organic carbon remineralization and storage as dissolved inorganic carbon in the deep sea. We present a high-resolution reconstruction of relative changes in oxygen concentration in the deep North Atlantic for the past 1.5 million years using a paleo-proxy: the carbon isotope gradient between epifaunal and infaunal benthic foraminifera species. We report a significant (>40 µmol kg$^{-1}$) reduction in glacial Atlantic deep-water oxygenation at ~960–900 ka that coincided with increased continental ice volume and a major change in ocean thermohaline circulation. Paleo-oxygen results support a scenario of decreasing deep-water oxygen concentrations, increased respired carbon storage, and a reduction in glacial $p$CO$_2$ across the Middle Pleistocene Transition.

One-Sentence Summary: An important reduction in glacial deep-sea oxygen concentrations accompanied ice volume and atmospheric $p$CO$_2$ changes ~900 ka.
Main Text: During the last 800 kyrs, glacial carbon dioxide concentrations (pCO₂) in Earth's atmosphere averaged 190 ppmv (1), about 90 ppm lower than the pre-Industrial value (1, 2). Beyond the ice core record, atmospheric pCO₂ records are fragmentary but suggest that glacial pCO₂ may have been lower by 20–40 ppm prior to ~1000–800 ka during the early Pleistocene (3–7). Such a drop in glacial atmospheric pCO₂ concentrations is one of the proposed causes of increased continental ice volume during glacial periods across the Middle Pleistocene Transition (MPT) (8, 9), which occurred between 1.25–0.64 Ma (10). During the MPT ice sheets grew larger and the duration of glacial cycles lengthened from dominantly 41-kyr oscillations before the MPT to quasi-100-kyr cycles afterwards (8, 10, 11). Lowering glacial pCO₂ across the MPT most likely involved increased carbon storage in the deep ocean via enhanced biological CO₂ uptake (12–14) and/or reduced CO₂ exchange between the atmosphere and surface ocean (12, 15, 16) especially in the Southern Ocean (17, 18). These processes would not only increase carbon storage in the deep ocean but would also reduce the deep-sea concentration of dissolved oxygen.

Here we applied a paleo-proxy to estimate past changes in oxygen concentration (hereafter expressed as [O₂]) in deep-water of the North Atlantic. The proxy utilizes an empirical calibration between the [O₂] and the carbon isotopic composition difference (Δδ¹³C) between surface and deep-dwelling benthic foraminifera, as first proposed by McCorkle et al. (19, 20) and re-calibrated by Hoogakker et al. (21) (supplementary materials). We report Δδ¹³C as a proxy for North Atlantic deep-water [O₂] at International Ocean Discovery Program (IODP) Site U1385 on the Iberian Margin [37°34.285’ N, 10°7.562’ W, 2578 meters below sea-level (mbsl)] (22, 23)
(Fig. 1; supplementary materials). We measured the epibenthic species *Cibicidoides wuellerstorfi* that provides a record of deep-water carbon isotopic composition ($\delta^{13}C$) (21, 24), whereas the infaunal species *Globobulimina affinis* records $\delta^{13}C$ near the oxic-anoxic boundary in sediment porewaters (21, 24, 25). The carbon isotope gradient ($\Delta\delta^{13}C_{\text{cib-aff}}$) expresses the difference between the two.

The main assumption of the $\Delta\delta^{13}C_{\text{cib-aff}}$-derived [O$_2$] method is that the carbon isotope gradient between the sediment water interface and the oxic-anoxic boundary is controlled by deep-water [O$_2$] and hence total dissolved inorganic carbon (DIC) released during aerobic respiration. Addition of excess DIC through anaerobic processes such as nitrate reduction or sulfate reduction serves to lower $\delta^{13}C$ at the oxic-anoxic boundary (20, 26), and causes an overestimation in [O$_2$] (26). Thus, derived paleo-[O$_2$] estimates represent maximum deep-water oxygen concentrations. The Hoogakker et al. (21) calibration includes release of DIC through anaerobic processes as demonstrated by the non-zero y-intercept in their empirical calibration (supplementary text section 2).

Low organic carbon concentrations (<1%) at Site U1385 leads to relatively low rates of organic carbon oxidation and a deep sulfate-methane transition at ~50 meters below the seafloor (mbsf) (27). The Iberian Margin near-surface sediment column demonstrates low rates of sulfate reduction in the upper tens of centimeters (28), suggesting that $\Delta\delta^{13}C$ at Site U1385 is governed predominantly by the amount of aerobic respiration controlled by deep-water [O$_2$]. Following
recommendations of Jacobel et al. (26), we compare the paleo-[O\textsubscript{2}] estimates derived from
\[\Delta\delta^{13}C_{\text{cib-aff}}\] with other redox/oxygenation proxies: U/Ca, U/Mn and C\textsubscript{26}OH/(C\textsubscript{26}OH + C\textsubscript{29}). We
also assess relative changes in paleo-productivity that may lead to variable organic carbon fluxes
using several proxies: \[\Delta\delta^{13}C_{\text{cib-uvig}}, Uvigerina\] spp. abundance and sedimentary Ba/Al ratios (see
supplementary text sections 3 and 4).

The cores from different holes at Site U1385 were spliced to produce a composite section that
extends from 0 to 166.5 meters composite depth (mcd), which is equivalent to Marine Isotope
Stage (MIS) 47 at ~1.45 Ma. Sedimentation rates average \~11 cm kyr\textsuperscript{-1} (23) which translates into
a mean sampling frequency of \~290 yrs with no significant difference across the MPT. Today the
water depth of 2578 mbsl places Site U1385 in well-ventilated North Atlantic Deep Water
(NADW) that is made up of varying proportions of Classical Labrador Sea Water (CLSW;
\~45\%) (see Table. S1 for [O\textsubscript{2}]), Iceland Scotland Overflow Water (ISOW; \~30\%) and Denmark
Strait Overflow Water (DSOW; \~5\%), with the remainder (\~20\%) derived from the modified
Antarctic Bottom Water (AABW) (29). Present-day [O\textsubscript{2}] at the site ranges between values of
\~237–245 \textmu mol kg\textsuperscript{-1} depending on [O\textsubscript{2}] profiles and databases considered (Figs.1, S1 and table
S1) and the \[\delta^{13}C\] value of DIC averages \~1‰ (Fig. 1). During glacial stages, the relative
proportion of northern-sourced deep-water decreased at Site U1385 as the fraction of southern-
sourced deep-water increased (30–33).
Hoogakker et al. (21) applied the $\Delta^{13}C_{cib-aff}$ method to estimate $[O_2]$ for the last 160 kyrs using piston core MD95-2042 (3146 mbsl), which is slightly deeper than nearby Site U1385 but with very similar modern-day oxygen concentrations (~245 µmol kg$^{-1}$). They concluded that deep-water $[O_2]$ was lower during the last two glacial stages relative to today, with the lowest values (~100 µmol kg$^{-1}$ lower) recorded during cold stadial conditions associated with Heinrich Events. They used the calibration of Hoogakker et al. (21) to convert $\Delta^{13}C_{cib-aff}$ to deep-water $[O_2]$ for values $<2.1\%$, which correspond to $[O_2] < 235$ µmol kg$^{-1}$ where a strong linear relationship exists: $\Delta^{13}C_{cib-aff} = 0.0064 \times [O_2] + 0.555$; ($R^2 = 0.95$) (Eqn. 1). Above 235 µmol kg$^{-1}$ the relationship breaks down, so we truncate estimated $[O_2]$ values at this concentration (Fig. 2E). For an alternative calibration, see supplementary materials (Fig. S2).

We assess the reliability of the $\Delta^{13}C_{cib-aff}$ proxy using a transect of core-top sediment recovered from the Iberian Margin during Cruise JC089 (22) (Fig. S1A). Core-top $\Delta^{13}C_{cib-aff}$ values demonstrate that the shape generally follows water-column $[O_2]$ with low values associated with Mediterranean Outflow Water (MOW) and higher values in NADW below (Fig. S1B; supplementary materials). Because only the MOW has $[O_2] < 235$ µmol kg$^{-1}$ we could only test the calibration on two core tops under the influence of the MOW. The difference between CTD measured $[O_2]$ and predicted $[O_2]$ using measured $\Delta^{13}C_{cib-aff}$ was 0 and 17 µmol kg$^{-1}$, which is within the estimated error of the calibration (±17 µmol kg$^{-1}$) (21) (see supplementary materials).
During the early Pleistocene from ~1435–960 ka (MIS 47–27) deep-water [O_2] was relatively high and only once fell below ~180 µmol kg\(^{-1}\) (~1120 ka) (Figs. 2E and S3). MIS 26 (~960 ka) marks the first time in the record when glacial [O_2] dropped to ~140 µmol kg\(^{-1}\). Lower values were attained over the next 3 glacials (MIS 24–20) and throughout subsequent terminations before dropping to the lowest [O_2] value (~95 µmol kg\(^{-1}\)) during MIS 18 (~750 ka; Fig. 2E). The interval from ~960–750 ka (MIS 26–18) is marked by a notable, persistent decrease in Δδ\(^{13}\)C\(_{\text{cib-aff}}\) (Fig. 2D), particularly throughout MIS 22 (~890 ka), when deep-water [O_2] exhibits a greater number of low-[O_2] events compared to those in the early Pleistocene. Over the MIS 27–21 (980–840 ka) interval, often referred to as the “900-ka event” \((10)\) or erroneously as the first 100 kyr cycle \((34)\), low-[O_2] values are supported by other recorders of paleo-[O_2]. High U/Ca and U/Mn (Figs. S4 and S5B) \((35, 36)\) in foraminifera coatings measured in the same Site U1385 samples as those used for Δδ\(^{13}\)C\(_{\text{cib-aff}}\) measurements (supplementary materials), support a drawdown in glacial deep-water [O_2]. Furthermore, qualitative indicators of low organic carbon flux during periods of low-[O_2] provide confidence in reconstructed depleted-[O_2] \((26)\) events at this time (Fig. S5C–E; supplementary materials).

Between ~620–470 ka (MIS 15 to early MIS 12) deep-water [O_2] decreases were weaker and less frequent than those during the preceding period (MIS 26–16), and similar to those observed during the early Pleistocene (Fig. 2E). MIS 14 was a weak glacial period which may explain why oxygen remained relatively high at this time. From ~470–10 ka (late MIS 12 to the early
Holocene), the magnitude of deep-water \([O_2]\) depletions intensify supported by the alcohol preservation proxy for oxygenation \((\text{C}_{26}\text{OH}/(\text{C}_{26}\text{OH} + \text{C}_{29}) (33); \text{Fig. S6})\) but remain higher than those centered on 900 ka.

The relationship between changes in deep-water \([O_2]\) and atmospheric \(p\text{CO}_2\) arises from the biological carbon pump through the production and consumption of organic matter \((18, 37–39)\). Whereas many marine processes are involved in \(p\text{CO}_2\) variations on glacial-interglacial timescales (see \((40)\) for a review), changes in the efficiency of the ocean’s soft-tissue pump directly impact dissolved deep-water oxygen through the regeneration of respired carbon. Weaker overturning circulation during glacial stages leads to a more efficient biological pump (i.e., an increase in the ratio of regenerated versus preformed nutrient content in the interior ocean \((18)\) by stemming the ventilation ‘leak’ of \(\text{CO}_2\) from the ocean to the atmosphere and increasing its oceanic residence time \((41)\). At the same time a stronger biological pump (i.e., an increased rate of organic carbon exported through greater productivity in the surface ocean \((42)\) will result in greater organic carbon regeneration in the ocean’s interior enhancing deep-sea respired carbon storage with a corresponding decrease in oxygen. Widespread reduction in deep-sea \([O_2]\) (without compensatory change at intermediate water depths \((43)\)), implies increased carbon storage and decreased atmospheric \(p\text{CO}_2\) \((18, 38, 44)\). Reconstruction of deep-water paleo-[\(O_2\)] therefore constitutes an important tool for estimating carbon transfers between the atmosphere and abyssal ocean and estimating glacial-interglacial changes in respired carbon storage and atmospheric \(p\text{CO}_2\) \((21, 26)\).
We compare estimated changes in \([O_2]\) from Site U1385 with other deep Atlantic paleoceanographic records to determine how widespread the observed oxygen depletions were. Comparison with the U/Ca record from the Deep Sea Drilling Project (DSDP) Site 607, which is bathed in a similar water mass as Site U1385 (Figs. 1 and 3D), reveals strong correlation between low-[O_2] events at Site U1385 and increases in U/Ca at Site 607 (16), suggesting intervals of reduced oxygenation were widespread in the deep North Atlantic. Tightly coupled changes in U/Ca and benthic $\delta^{13}C$ at Site 607 (Fig. S7) suggest a link to changes in deep-water circulation (16), with both proxies showing a change at MIS 26 (~970 ka).

Decreases in deep-water [O_2] values are expected to mirror increases in deep-water [PO_4^{3-}] according to stoichiometric ratios involved in organic matter respiration (38, 45). ODP Site 1267 in the Southeast Atlantic (Fig. 1) provides evidence of an increase of 0.5 μmol kg\(^{-1}\) in [PO_4^{3-}] at ~960 ka (Fig. 3E) (14) coeval with decreases in [O_2] and decreasing P/Al ratios at Site U1308 in the North Atlantic (Figs. S8 and S9B; supplementary text section 5). Both proxy-derived [PO_4^{3-}] at Site 1267 and [O_2] at Site U1385 show statistically significant changes across the MPT (Figs. S3 and S9A and Table S2 for statistics), suggesting a greater nutrient content and respired carbon pool in the deep Atlantic Ocean after 960 ka.

The change in nutrient content of the deep South Atlantic also corresponds to a major change in thermohaline circulation (THC) inferred from changes in neodymium isotopes (14, 46, 47).
Neodymium isotope ($\varepsilon$Nd) records from the North Atlantic DSDP Site 607/V30-97 (Figs. 1 and 2F; (47)) and ODP Sites 1267 (14), 1088 and 1090 (46) in the Southeast Atlantic (Figs. 1, 3F and 3G) document significant reduction in NADW contribution and/or an increased influence of AABW (14) between 950–900 ka (14, 46). These THC changes occurred at times of reduced upwelling and degassing of southern-sourced deep-waters under expanded sea ice cover in the Southern Ocean, thereby contributing to lowering of benthic foraminifer $\delta^{13}$C (16).

Changes in deep-water circulation can affect $[O_2]$ in several ways. Preformed $[O_2]$ may decrease at sites of deep-water formation if equilibration of oxygen with the atmosphere is reduced under sea ice, leading to undersaturation in preformed oxygen (48, 49). An increase in deep-water residence time results in decreased oxygen concentrations through organic matter oxidation with or without attendant changes in carbon flux from the surface (41, 50). Lastly, a change in the proportion of northern versus southern-derived deep-water masses (16, 51, 52) bathing Site U1385 can affect $[O_2]$; for example, NADW has $[O_2]$ values that are $\sim$50 µmol kg$^{-1}$ higher than those of AABW and 75 µmol kg$^{-1}$ higher than Circumpolar Deep Water (CDW) (Fig. 1; Table S1) (53).

Changes in source areas of NADW formation could substantially affect $[O_2]$ in the deep North Atlantic. Classical Labrador Sea Water (CLSW) is a well-ventilated water mass that distributes oxygen throughout the entire North Atlantic (54) including Site U1385 on the Iberian Margin that is bathed today by $\sim$45% CLSW (29). Oxygen saturation of CLSW is highly sensitive to
winter conditions in the Labrador Sea through bubble-mediated air-sea transfer associated with intensive winds, cooling and deep convection (55). The Labrador Sea has been described as a “trapdoor” through which the flux of oxygen ventilates much of the deep Atlantic basin (55).

During Heinrich stadials of the last glaciation the Labrador Sea was covered by extensive sea ice which would have reduced the ventilation of the deep Atlantic by CLSW (21, 56). Sea ice also expanded over the Nordic Seas (57–59) (source areas of DSOW and ISOW (53)) and the Irminger Sea where these water masses mix to form NADW (53, 60, 61).

The decrease in glacial deep-water [O₂] at ~960 ka coincides with an increase in continental ice volume and lowered sea-level, inferred from changes in δ¹⁸O of seawater at Site 1123 in the Southwest Pacific (51, 62, 63) (Figs. 2G, 3B and 4A). Available evidence suggests Northern Hemisphere glaciation intensified during MIS 22 ((51) and references therein), which resulted in changes to Atlantic deep-water circulation (52).

The timing of the initial decrease in deep-water [O₂] at ~960 ka during MIS 26 is associated with a short-lived excursion to very negative εNd values at Site 607 (Fig. 2F; (47)). This event is interpreted as resulting from extensive weathering and erosion of the North American craton between MIS 27–25 (~980–950 ka) and the possible transition from terrestrial- to marine-terminating ice sheets in the Northern Hemisphere (64). MIS 26 also represents the first time the polar front shifted to a zonal position south of Site 980/981 in the Northeast Atlantic (55.5°N, 14.7°W) as inferred from an increase in the percent of the polar planktonic foraminifera
Neogloboquadrina pachyderma sinistral (%NPS), a proxy for cold surface temperatures (Fig. S10; (65, 66)). These precursor changes heralded the major changes in ice volume, thermohaline circulation and $pCO_2$ associated with the “900-ka event” during MIS 24–22 in less highly resolved records than those of Sites U1385 and 980/981 (3, 4, 6, 10, 14, 46, 47, 62, 65).

The deep-water oxygen depletions may have been driven by increased freshwater inputs and ice rafting to the source areas of NADW formation. The lowest $[O_2]$ values during the last glacial cycle are associated with cold millennial events in the North Atlantic (Fig. S11B) (21).

Millennial $[O_2]$ depletion events occur throughout the entire record and are most common towards the latter part of glacial cycles and during terminations. These times are also associated with enhanced sea ice extent in the North Atlantic (67), cold stadial periods and increases in ice-rafted detritus (IRD) at both the North Atlantic Site U1308 (68) (Figs. 1, S11A and S12) and the Labrador Sea Site U1302/03 (Fig. S13; (69)). A close connection exists today between North Atlantic deep-water oxygenation and winter surface conditions in the Labrador, Irminger and Nordic Seas (55–58). We suggest enhanced surface stratification and reduced deep-water convection in NADW source areas may have caused episodic reduction of North Atlantic deep-water ventilation, leading to reduced deep-water oxygen and increased carbon storage (70).

Incursions of glacial southern-sourced deep-water into the North Atlantic associated with weaker NADW production (51, 52, 61, 71, 72) could also partly account for $[O_2]$ changes observed at Site U1385 over the past 1.5 Ma given the $[O_2]$ difference (~50 µmol kg$^{-1}$) between northern and southern-sourced deep-water today (Fig. 1; Table S1).
The changes in deep-water [O$_2$] at Site U1385 could equally be driven by surface processes in the Southern Ocean such as changes in productivity (13), surface stratification (15, 73), vertical mixing and sea ice extent. These changes would be transmitted to the deep-sea via an expanding southern-sourced deep-water mass such as Lower Circumpolar Deep Water (74) in which oxygen was considerably reduced during the last glacial period (26, 36, 44). Reduction in ventilation due to circulation changes, Southern Ocean stratification and sea ice expansion would have contributed to the inferred increase in deep-sea carbon storage over the MPT (14–16, 18, 73, 75–78).

Decreased glacial deep-water [O$_2$] and increased deep-sea carbon storage across the MPT have implications for atmospheric $p$CO$_2$. The $\Delta \delta^{13}$C$_{\text{cib-aff}}$ proxy of [O$_2$] at Site U1385 is consistent with changes in $p$CO$_2$ measured directly in ice cores and blue ice in Antarctica (1, 6, 7) and inferred indirectly using boron isotopes (3–5) (Fig. 4B). Prior to ~900 ka, minimum glacial $p$CO$_2$ values were ~24 ppm higher than they were afterwards (3, 5–7). If applied to the whole Atlantic Ocean influenced by an expanding southern-sourced deep-water mass Hoogakker et al. (21) estimated at least 15% of the reduction in atmospheric $p$CO$_2$ during the last glacial maximum (LGM) could be accounted for by the increases observed in their respired carbon pool. The reconstructed glacial [O$_2$] reported here for the MPT are at least 20 µmol kg$^{-1}$ lower than the LGM (21), implying greater carbon storage than the LGM. It remains to be seen from other
basins, such as the Pacific Ocean where deep-water \([O_2]\) is also reported to have been lower for the LGM (26, 44), whether this was a global phenomenon.

In summary, inferred changes in North Atlantic deep-water \([O_2]\) for the past 1.5 Ma reveal a significant (>40 \(\text{µmol kg}^{-1}\)) reduction in glacial deep-water \([O_2]\) at ~900 ka suggesting increased storage of respired carbon, which is consistent with a drawdown of glacial atmospheric \(pCO_2\) values (Fig. 4B) (1, 3, 4, 6, 7). The inferred change in \([O_2]\) is supported by trace metal records (16) and nutrient (14) proxy records in other Atlantic sites associated with a critical change in glacial THC at ~900 ka (46) (Fig. 3C–G). The close association between \([O_2]\) depletions and IRD events (68, 69, 79) suggests that increased stratification and sea ice cover in NADW source regions (56–59) reduced oxygen supply to much of the deep North Atlantic (55). In addition, northward expansion of southern-sourced deep-water into the North Atlantic and processes in the Southern Ocean (e.g., productivity, surface water stratification, vertical mixing, and sea ice extent) also contributed to reduced ventilation associated with a major change in deep-water circulation (46). Our results support a set of internally consistent changes in Atlantic deep-water beginning at ~960 ka across the MPT, that included a decrease in oxygen concentrations, increased nutrient concentrations (14) and storage of respired carbon that led to a reduction in glacial \(pCO_2\) (3–7, 47) and an associated increase in global ice volume (62).
References and Notes

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**Supplementary Materials**

Materials and Methods

Supplementary Text

Figs. S1 to S13

Tables S1 to S2

References (80–92)
Fig. 1. Atlantic hydrographic profiles of $[O_2]$ and $\delta^{13}C$ relative to site locations. (A) Deep-water $[O_2]$ profile at the depth (2578 mbsl) of IODP Site U1385 (yellow diamond; this study) in the Northeast Atlantic along with the geographic position of other sites.
referred to in text: North Atlantic DSDP Site 607 ((16, 47); pink triangle); IODP Site U1308 ((68); green circle); South Atlantic ODP Sites 1267 ((14); orange triangle), Sites 1090 and 1088 ((46); purple square and red circle respectively). Meridional cross-sections showing the hydrographic profiles of (B) $[O_2]$ and (C) $\delta^{13}C$ of DIC relative to the core locations. Maps created in Ocean Data View (Schlitzer, 2019; http://odv.awi.de).
Fig. 2. North Atlantic IODP Site U1385 deep-water records (this study) compared to Atlantic meridional overturning circulation, ice volume and $pCO_2$ estimates across the MPT. Site U1385 records extend the record of piston core MD95-2042 (21) (light grey). (A) *Cibicidoides wuellerstorfi* $\delta^{18}O$ (dark grey); MIS numbers denote interglacials (odd) and glacialis (even); black dashed boxes highlight a ~30-kyr hiatus and gap. (B and C) Benthic foraminiferal $\delta^{13}C$: *Cibicidoides wuellerstorfi* (yellow) and *Globobulimina affinis* (green). (D) $\Delta\delta^{13}C_{cib-aff}$ (orange); the dashed line at 2.1‰ (orange) marks the calibration threshold. (E) $[O_2]$ (black line with diamonds), truncated at 235 $\mu$mol kg$^{-1}$ above which the calibration is unreliable; MIS numbers denote important glacialis. (F) DSDP Site 607/V30-97 $\varepsilon_{Nd}$ proxy for North Atlantic Ocean circulation changes (brown line with orange squares; (47)). (G) ODP Site 1123 ice volume $\delta^{18}O_{seawater}$ (blue; (62)). (H) Atmospheric $pCO_2$ records: Antarctic 800-kyr ice core data (black line; (1)); Allan Hills Blue Ice ((6, 7); blue triangles and squares respectively); $\delta^{11}B$-based reconstructed $pCO_2$ (green line with diamonds, (5); early Pleistocene light green small circles and late Pleistocene light green circles (4); green squares, (3)). See methods for chronology.
Fig. 3. Comparison of Atlantic Ocean deep-water oxygenation, nutrient and circulation proxies and South Pacific ice volume changes through the MPT. (A) IODP Site U1385
Cibicidoides wuellerstorfi δ18O (dark grey, this study); MIS numbers denote interglacials (odd) and glacial (even). (B) ODP Site 1123 ice volume δ18O_{seawater} (blue; (62)). (C) [O2] estimated from Δδ13C_{cib-aff} (black diamonds, this study), MIS numbers denote important glacials. (D) DSDP Site 607 U/Ca redox proxy (purple diamonds; (16)). (E) ODP Site 1267 Cd/Ca derived [PO_{4}^3-] (green squares; (14)). (F and G) ODP Sites 1088 (mid-depth) and 1090 (deep) ε_{Nd} respectively ((46); interglacial (orange circles) and glacial (blue circles) maxima.

Fig. 4. Comparison of deep-water oxygen, pCO_2 and ice volume proxies. (A) IODP Site U1385 Δδ^{13}C_{cib-aff} including piston core MD95-2042 data (21) (Gaussian interpolated with a 500-
yr interval, orange line); the oxygen calibration is only valid for $\Delta \delta^{13}C_{cib-aff} < 2.1 \%$; ODP Site 1123 ice volume $\delta^{18}O_{seawater}$ ((62) Gaussian interpolated with a 500-yr interval, blue line); MIS numbers denote major interglacials and glacials. (B) IODP Site U1385 $\Delta \delta^{13}C_{cib-aff}$ (as in A); Atmospheric $pCO_2$ records: Antarctic 800 kyr ice core data ((1) Gaussian interpolated with a 500-yr interval, black line); Allan Hills Blue Ice ((6, 7) ; light blue filled triangles and squares respectively); $\delta^{11}B$-based reconstructed $pCO_2$ (green line with diamonds, (5); early Pleistocene light green line/small circles and late Pleistocene light green circles (4); green filled squares, (3)).