

Mode change of millennial CO₂ variability during the last glacial cycle associated with a bipolar marine carbon seesaw

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Important elements of natural climate variations during the last ice age are abrupt temperature increases over Greenland and related warming and cooling periods over Antarctica. Records from Antarctic ice cores have shown that the global carbon cycle also plays a role in these changes. The available data shows that atmospheric CO₂ follows closely temperatures reconstructed from Antarctic ice cores during these variations. Here, we present new high-resolution CO₂ data from Antarctic ice cores, which cover the period between 115,000 and 38,000 y before present. Our measurements show that also smaller Antarctic warming events have an imprint in CO₂ concentrations. Moreover, they indicate that during Marine Isotope Stage (MIS) 5, the peak of millennial CO₂ variations lags the onset of Dansgaard/Oeschger warmings by 250 ± 190 y. During MIS 3, this lag increases significantly to 870 ± 90 y. Considerations of the ocean circulation suggest that the millennial variability associated with the Atlantic Meridional Overturning Circulation (AMOC) undergoes a mode change from MIS 5 to MIS 4 and 3. Ocean carbon inventory estimates imply that during MIS 3 additional carbon is derived from an extended mass of carbon-enriched Antarctic Bottom Water. The absence of such a carbon-enriched water mass in the North Atlantic during MIS 5 can explain the smaller amount of carbon released to the atmosphere after the Antarctic temperature maximum and, hence, the shorter lag. Our new data provides further constraints for transient coupled carbon cycle-climate simulations during the entire last glacial cycle.

abrupt climate change | CO₂-temperature phasing | ice age variability | paleoclimate | greenhouse gas

The climate of the last glacial period is characterized by low global mean temperatures and a number of interhemispheric variations on time scales of several millennia. In the northern hemisphere, in particular on the Greenland ice sheet, large and very rapid temperature jumps of up to 15 °C within a few decades, followed by more steady decreases of temperature, have been identified from ice core analyses (1, 2). 25 of these so-called Dansgaard-Oeschger (DO) events have been found during the last glacial period. For each of these events, an associated Antarctic temperature variation has been also documented in Antarctic ice cores (Antarctic Isotope Maximum [AIM] events) (3). Antarctic temperatures increase siteadily when Greenland is in the cold phase and slowly decrease following the abrupt temperature increase in Greenland during a DO event. This behavior is explained by the concept of an oceanic thermal bipolar seesaw, where variations in the Atlantic Meridional Overturning Circulation (AMOC) and associated changes in heat transport across the equator to the North Atlantic modulate the southern and northern hemisphere temperatures during each of these events (4).

Previous reconstructions of atmospheric CO₂ concentrations during these millennial-scale variations show that CO₂ varies largely in parallel with the major AIM events (5, 6). Those data indicate stable CO₂ concentrations for up to 2,000 y after the peak of major AIM events during marine isotope stage (MIS)

3; however, data coverage and resolution are not sufficient to completely resolve CO₂ during the seesaw variability of the entire MIS 5–2 time window.

Here we present a new CO₂ record from the Talos Dome ice core (TALDICE, 159 °E; 73 °S) with 116 data points and an average resolution of 280 y and an extended record from the EPICA (European Project for Ice Coring in Antarctica) ice core retrieved in Dronning Maud Land (EDML, 0 °E; 75 °S) (7) with 92 new data points and 450 y average resolution going back to 115,000 y before present (115 kyr BP, where present refers to 1950). Together, the records permit a detailed insight into the coupled carbon cycle-climate dynamics over the time period from 115 to 38 kyr BP. For most parts of the TALDICE core, atmospheric CH₄ concentration data are available with a resolution of a few decades (8). The resolution of the available EDML CH₄ record (3, 9–11) is 170 y on average. The dense data coverage of CH₄ and CO₂ in both cores, the fact that all DO event onsets are associated with a rapid increase of atmospheric CH₄ concentrations (10), and the low CO₂ measurement uncertainty of better than 1.5 ppm by volume (ppmv) (*Materials and Methods*) allow us to investigate the interplay of variations in the bipolar seesaw and CO₂ in unprecedented detail. This new analysis offers new insights into mechanisms responsible for millennial carbon cycle variations during the last ice age and also provides further constraints for transient simulations using coupled carbon cycle-climate models.

Results

In Fig. 1, the TALDICE and EDML CO₂ records are shown, as well as the previously published records of the Byrd (6), Taylor Dome (5), and Vostok (12) ice cores. With the exception of the Vostok record, all records are synchronized on the EDML1 Scenario 4 gas age scale (13) by matching tie points in the CH₄ records (*SI Text*). There is a clear correlation of the CO₂ records where they overlap. However, the different cores show offsets of a few ppmv during certain parts of the records. These offsets are discussed in detail in the *SI Text*. Note that the offsets only affect the overall concentration level in the individual cores but not the temporal CO₂ changes; thus, also not the phase relationship between CO₂ and the onset of DO events investigated in this study. Accordingly, they do not affect the conclusions drawn in our study. Our two new records represent the temporal evolution of CO₂ in even more detail than previous records, because the main features are covered by two independent ice cores, the

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uncertain but model experiments suggest that soil carbon changes contribute only little to the atmospheric CO₂ concentration during DO-like events (18).

Dust input in the Southern Ocean [approximated by the non-sea salt (nss) Ca²⁺ flux in the EDML ice core] (Fig. 1) (19) also provides iron to the surface water which is considered a fertilizer for marine biology and, hence, can influence atmospheric CO₂. If changes in dust input caused the detected shift in the phase lag of CDM relative to the DO onset, the peaks of the dust input should also shift relative to AIM events from MIS 3 to 5. Here, we compare the phasing between nssCa²⁺ as proxy for Fe input into the Southern Ocean, on the one hand, and water isotopes in the ice as Southern Ocean temperature indicator, on the other hand. Due to the large error in the gas age/ice age difference, a direct comparison of CO₂ and nssCa²⁺ is not conclusive. Neither a clear shift of dust input relative to AIM events between MIS 3 and 5, nor a significant dust decrease after the peak of AIM events, is recorded. Furthermore, latest carbon cycle model experiments show that dust flux variations associated with a full AIM event (changes of factor 10) explain only a 4–8 ppmv CO₂ change (25). Accordingly, the dust fertilization effect has most likely contributed to the CO₂ increase before the onset of DO events but is an unlikely explanation for the shift in the phasing between CDM and DO events.

Changes in ocean circulation and the distribution of major water masses can influence atmospheric CO₂ concentrations (22, 26). One proxy for ocean hydrographic changes is the Nd isotope ratio in ocean sediment cores (expressed as ε_{Nd}), which traces deep water origin at the drill site (Fig. 3C) (27). An Indian Ocean record (28) shows stable ε_{Nd} values at Holocene values (about –8.8) during the whole MIS 5, indicating continuous presence of North Atlantic Deep Water (NADW) at that site. At the transition into MIS 4, there is a pronounced increase in ε_{Nd} to glacial values (about –7.1) within a few thousand years indicating a decrease of NADW reaching this site. In a higher resolved record of the southern Atlantic (29), a similar change is found at the MIS 5–4 transition from about –8.3 (late MIS 5) to –7.3 (MIS 4). At that site, the late MIS 5 level seems to be slightly higher than the Holocene level (–8.9). Furthermore, a slowdown of the AMOC at the MIS 5–4 transition, with relatively constant NADW export during MIS 5, has been reported from a sediment core in the deep northeastern Atlantic (Fig. 3F) (30). Taken together, these records indicate that at the MIS 5–4 transition deep ocean circulation in the Atlantic changed significantly from a modern-like situation, where NADW flows southward at depth to the southern Atlantic and then partly into the Indian Ocean, to a glacial situation with a slower AMOC and a shallower Glacial North Atlantic Intermediate Water (GNAIW) cell lying on top of an increased Antarctic Bottom Water (AABW) mass extending into the North Atlantic (Fig. 4) (26). The salient difference from MIS 5 to MIS 3 conditions therefore appears to be the presence of a weakly ventilated, carbon-rich water mass in the deep Atlantic Ocean. This is supported by a synthesis of benthic δ¹³C records (31), which shows a decrease of δ¹³C in the Atlantic basin and the Southern Ocean at the MIS 5–4 transition (Fig. 3E). This indicates an increase in respired carbon in these bottom waters and could be partly responsible for the atmospheric CO₂ decrease at the MIS 5–4 transition.

During the AIM events 8, 12, 14, and likely also 17, the southern Atlantic ε_{Nd} record—and somewhat also weaker in the Indian ocean record—suggests associated variations of NADW reaching these sites (Fig. 3C) (28, 29). In the case of the southern Atlantic record, synchronous δ¹³C variations have been found, indicating the possibility of circulation-driven changes in the carbon content of deep waters during these variations (29). Moreover, increased upwelling in the Southern Ocean of deep ocean waters has been inferred from records of opal accumulation associated with AIM events 8, 12, 14, 17, 19, 20, and possibly 21

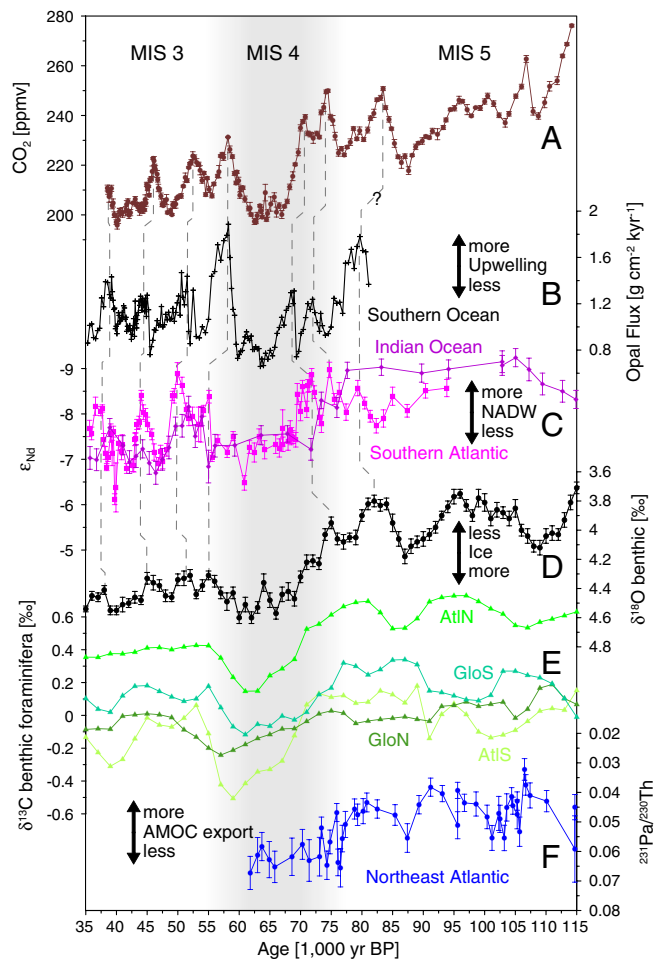


Fig. 3. Compilation of atmospheric CO₂ and ocean sediment records from 115 to 35 kyr BP. (A) CO₂ composite record from TALDICE and EDML ice core. (B) Opal Flux record from a sediment core at the polar front region in the Southern Ocean (32). Changes in these data are interpreted as changes in marine productivity in the polar front region. (C) Nd-isotope record from two sediment cores from the Indian (3° N, 76° E, 3,800 m below sea level [b.s.l.]) and Southern Atlantic ocean (40° S, 8° E, 4,981 m b.s.l.) indicating which portion of the local water masses originated from NADW (28, 29). (D) Benthic foraminifera δ¹⁸O composite as a combined proxy of total ice volume/sea level and deep ocean temperature (41). (E) δ¹³C data of benthic foraminifera from sediment cores below 2,500 m depth in the ocean. The four lines represent averages of the corresponding data from (31) over the following four different sectors: Atlantic North (AtIN) contains the data from the Atlantic basin between 90° N and 40° S and Atlantic South (AtIS) those between 40° S to 90° S. Global North (GloN) and Global South (GloS) include all data from (31) of the same latitudinal sectors as before (GloS: 40° S–90° S; GloN: 90° N–40° S) except those from the Atlantic. δ¹³C values of benthic foraminifera are partly linked to the respired carbon content of the local deep ocean waters, hence, age of the water mass. (F) ²³¹Pa/²³⁰Th record from a sediment core in the northeastern Atlantic (30) (39° N, 13° W, 3,547 m b.s.l.) interpreted as proxy of AMOC export strength. The gray-shaded area indicates likely relations between CDM and variations in the sediment records. All records are plotted on their own age scales.

(Fig. 3B) (32). These millennial-scale variations in the sediment proxies and the differences in ocean hydrography between MIS 5 and 3 could explain the different characteristics of CDM between MIS 3 and 5 in the following way: Before the DO onset of a MIS 3 type event, the Atlantic ocean is in a glacial state with a carbon-rich and expanded AABW and a shallower GNAIW cell (Fig. 4). At the DO onset, the AMOC accelerates rapidly bringing heat to the north but also causes—at least during major events—a resumption of NADW formation. This leads to increased ventilation of

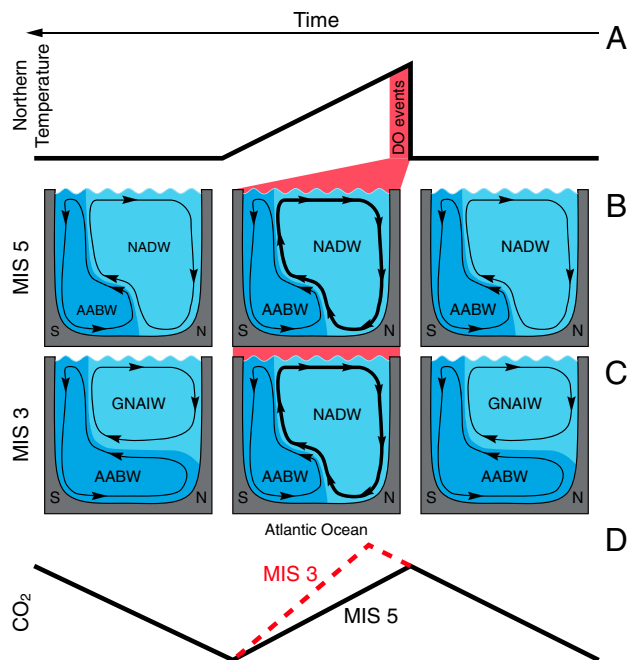


Fig. 4. Illustration of suggested water mass changes during DO events. (A) Schematic evolution of Greenland temperature during a DO event. (B) Schematic time slices of North Atlantic deep water mass (NADW, light blue) and Antarctic bottom water mass distributions (AABW, dark blue) as well as circulation (arrows) states in the Atlantic Ocean during MIS 5 type DO events. (C) Same as in (B) but with presence of Glacial North Atlantic Intermediate Water mass (GNAIW, light blue) as suggested for MIS 3 type DO events. (D) Idealized evolution of atmospheric CO₂ during a MIS 5 (black line) and MIS 3 type DO event (black plus red dashed line).

the deep Atlantic Ocean associated with an expansion of the NADW cell at depth at the expense of AABW and an upwelling of carbon-rich waters in the south. The ventilation of carbon-rich waters adds CO₂ to the atmosphere for another 500 to 1,000 y while NADW formation continues until the CO₂ release is counterbalanced by decreasing southern temperatures and a re-expansion of the AABW mass and the associated carbon pool in the deep ocean. The crucial difference of the MIS 5 events compared to the MIS 3 events is the modern-like ocean state on which DO events are superimposed (Fig. 4). During MIS 5, AMOC also accelerates at the DO onset. However, we conclude that this acceleration releases significantly less CO₂ to the atmosphere than during MIS 3, since the additional carbon-enriched AABW reservoir was not as far extended into the North Atlantic as during MIS 3.

For the MIS 3 events, the TALDICE records show that atmospheric CO₂ increases after the DO event for another 8.4 ± 2.6 ppmv equivalent to 18 ± 5.5 GtC (Fig. 2, *Insets*). According to our hypothesis, this carbon should be provided by exchanging the deep waters in the North Atlantic from carbon enriched AABW to NADW. Is there enough carbon stored in AABW in the North Atlantic during MIS 3 stadial conditions? Multi-box modeling of the glacial carbon cycle (33) found a dissolved inorganic carbon (DIC) content of glacial AABW of nearly $2,600 \mu\text{mol/kg}$ whereas glacial North Atlantic waters contain about $2100 \mu\text{mol/kg}$. Thus, leaving everything else unchanged, replacing this AABW by NADW changes the DIC content in the deep North Atlantic by about $500 \mu\text{mol/kg}$. This replacement starts at the onset of the DO event. Using an exchanged water volume of $5.5 \cdot 10^{16} \text{ m}^3$ in the North Atlantic from 0° to 60°N, 60° longitude, and from 3,500 to 5,000 m water depth, 330 GtC could be released to the atmosphere by this exchange. However, the largest amount of that CO₂ is taken up by the surface ocean and is mixed again into the global ocean. The proportion of CO₂ that could remain in the atmosphere after such a release is given by

the ratio of the undisturbed atmospheric carbon pool and the ocean carbon pool corrected for the buffer factor (approximately 11). This results in a net input from the ocean into the atmosphere of 36 GtC, i.e., about twice as much as needed to explain the CDM lag.

Clearly, this represents a rough estimate that is only intended to show that in principle enough carbon is available in the extended AABW to explain the CDM lag. The temporal constraints recorded in the different proxies are not strong enough to argue for or against this mechanism, because a common time scale for the various paleorecords in Fig. 3 is missing. Furthermore, the estimates of the size of the carbon reservoirs—in particular the DIC content of AABW and NADW during glacial times—is uncertain and model dependent, although even a two times smaller DIC change could still account for the observed CO₂ change. The assumption that all ventilated waters reach the surface and release CO₂ provides an upper limit. Finally, our estimate only accounts for the overall amplitude of the CO₂ change, but it cannot constrain the temporal dynamics of these changes. Accordingly, we cannot make a quantitative statement on the size of the phase lag between the CO₂ maximum and the onset of the DO events based on this calculation.

Support for our hypothesis could come from an analysis of carbon isotopes at atmospheric CO₂ (¹³C) during CDM. So far, only one study presents data over CDM 12 from two different ice cores (34). The record from the Berkner Island ice core indicates a decrease in $\delta^{13}\text{C}_{\text{CO}_2}$ in the atmosphere at the peak of CDM 12 in line with a release of ¹³C depleted carbon from the deep ocean; however, the data uncertainty is still large. Better constraints for the CDM patterns can be provided by improved reconstructions of $\delta^{13}\text{C}_{\text{CO}_2}$ signals during several CDM and a common age scale of the different climate archives (35).

Coupled carbon cycle-climate model simulations may provide further quantitative insight. The classical approach to model abrupt climate changes associated with DO events was to trigger a change in the AMOC by perturbing the surface freshwater balance in the Atlantic. On the one hand, such experiments show atmospheric CO₂ variations, which are consistent in amplitude with the CDM found in the ice cores (17,18,24, 36–38). On the other hand, the different model studies are inconsistent with respect to the major carbon reservoir changes responsible for the atmospheric CO₂ variations. Under preindustrial conditions, most models respond with a release of CO₂ from the terrestrial biosphere and an uptake of CO₂ by the ocean to a reduction of the AMOC (17,18,24), however, also an opposite response is found in one study (38). Under glacial conditions, the models show a larger contribution of changes in the ocean inventory compared to those in other reservoirs (24, 37, 38). One model (24) even shows decreasing atmospheric CO₂ concentrations during the cold phase in Greenland when the AMOC is reduced, which is inconsistent with the available CO₂ records. From the available model studies, only one exhibited a lag of the CDM with respect to the AMOC resumption (38). In those simulations, the source of the carbon after the AMOC resumption seems to be the ocean, however, whether it originates from the mechanism as described here is not clear. Taken together, models still do not provide a consistent picture regarding the causes of CO₂ variations associated with millennial scale climate variations. Hence, a quantitative investigation of our hypothesis with current coupled carbon cycle-climate models is not yet feasible. Nevertheless, further model studies of the coupled carbon cycle-climate system during the last ice age will be indispensable to understand the subtle, and yet significant, mode changes discovered here.

Materials and Methods

The basic measurement technique is analogous to ref. 39. Here, modifications made in this study are discussed. Due to gas fractionation which occurs in the Bubble-Clathrate Transformation Zone (BCTZ) and which persists also just below the BCTZ (7), six to 10 replicate measurements at the TALDICE core

between 35 and 42 kyr BP have been performed to reduce the measurement uncertainties (usually four replicates). Since the fractionation is strongest just below the BCTZ (at 35 kyr BP in the TALDICE core on our time scale) and decreases with distance from the BCTZ due to diffusion in the ice, we could not reduce the uncertainty sufficiently in the TALDICE record between 35 kyr and 38 kyr with this approach (Fig. 1, pink area). This also applies to the EDML record between 30 kyr and 45 kyr BP (7). Nevertheless, the uncertainty between the time span from 38 to 42 kyr BP in the TALDICE record could be reduced to 1.5 ppmv (one sigma), which is comparable to 1.3 ppmv in the EDML record (Fig. 1). From 42 kyr BP back to the end of the TALDICE record, the average uncertainty is just 1.0 ppmv, owing to the good ice quality and the high air content (about 10% higher than the EDML ice core) of the TALDICE core. Furthermore, the TALDICE core has a higher accumulation rate than central east Antarctic ice cores, providing a narrower age distribution of the air bubbles included and thus making it a very suitable core for studying the atmospheric variability during MIS 3.

1. North Greenland Ice Core Project Members (2004) High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* 231:147–151.
2. Huber C, et al. (2006) Isotope calibrated Greenland temperature record over Marine Isotope Stage 3 and its relation to CH₄. *Earth Planet Sci Lett* 243:504–519.
3. EPICA Community Members (2006) One-to-one coupling of glacial climate variability in Greenland and Antarctica. *Nature* 444:195–198.
4. Stocker TF, Johnson J (2003) A minimum thermodynamic model for the bipolar seesaw. *Paleoceanography* 18 art. no. 1087.
5. Indermühle A, Monnin E, Stauffer B, Stocker TF, Wahlen M (2000) Atmospheric CO₂ concentration from 60 to 20 kyr BP from the Taylor Dome ice core, Antarctica. *Geophys Res Lett* 27:735–738.
6. Ahn J, Brook EJ (2008) Atmospheric CO₂ and climate on millennial time scales during the last glacial period. *Science* 322:83–85.
7. Lüthi D, et al. (2010) CO₂ and O₂/N₂ variations in and just below the bubble-clathrate transformation zone of Antarctic ice cores. *Earth Planet Sci Lett* 297:226–233.
8. Schüpbach S, Federer U, Bigler M, Fischer H, Stocker TF (2011) A refined TALDICE-1a age scale from 55 to 112 ka before present for the Talos Dome ice core based on high-resolution methane measurements. *Climate of the Past* 7:1–16.
9. Buiron D, et al. (2011) TALDICE-1 age scale of the Talos Dome deep ice core, East Antarctica. *Climate of the Past* 7:1–16.
10. Schilt A, et al. (2010) Atmospheric nitrous oxide during the last 140,000 years. *Earth Planet Sci Lett* 300:33–43.
11. Capron E, et al. (2010) Synchronizing EDML and NorthGRIP ice core using $\delta^{18}\text{O}$ of atmospheric oxygen ($\delta^{18}\text{O}_{\text{atm}}$) and CH₄ measurements over MIS5 (80–123 kyr). *Quaternary Science Reviews* 29:222–234.
12. Petit JR, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399:429–436.
13. Loulergue L, et al. (2007) New constraints on the gas age-ice age difference along the EPICA ice cores, 0–50 kyr. *Climate of the Past* 3:527–540.
14. Bereiter B, Schwander J, Lüthi D, Stocker TF (2009) Change in CO₂ concentration and O₂/N₂ ratio in ice cores due to molecular diffusion. *Geophys Res Lett* 36:1–5, AGU identifier: L05703 doi: 10.1029/2008GL036737.
15. Scholze M, Knorr W, Heimann M (2003) Modeling terrestrial vegetation dynamics and carbon cycling for an abrupt climatic change event. *Holocene* 13:327–333.
16. Wang YJ, et al. (2001) A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *Science* 294:2345–2348.
17. Obata A (2007) Climate-carbon cycle model response to freshwater discharge into the North Atlantic. *J Clim* 20:5962–5976.
18. Bozbiyik A, Steinacher M, Joos F, Stocker TF, Menviel L (2011) Fingerprints of changes in the terrestrial carbon cycle in response to large reorganizations in ocean circulation. *Climate of the Past* 7:319–338.
19. Fischer H, et al. (2007) Reconstruction of millennial changes in dust emission, transport and regional sea ice coverage using the deep EPICA ice cores from the Atlantic and Indian Ocean sector of Antarctica. *Earth Planet Sci Lett* 260:340–354.
20. Schneider R (2011) Quantifying past changes of the global carbon cycle based on $\delta^{13}\text{C}$ measurements in Antarctic ice cores. PhD thesis (Univ of Bern).
21. Monnin E, et al. (2001) Atmospheric CO₂ concentrations over the last glacial termination. *Science* 291:112–114.
22. Fischer H, et al. (2010) The role of Southern Ocean processes in orbital and millennial CO₂ variations—A synthesis. *Quat Sci Rev* 29:193–205.
23. Sanchez Goni MF, et al. (2008) Contrasting impacts of Dansgaard-Oeschger events over a western European latitudinal transect modulated by orbital parameters. *Quat Sci Rev* 27:1136–1151.
24. Menviel L, Timmermann A, Mouchet A, Timm O (2008) Meridional reorganizations of marine and terrestrial productivity during Heinrich events. *Paleoceanography* 23:1–19, AGU identifier: PA1203 doi: 10.1029/2007PA001445.
25. Porekh P, Joos F, Müller SA (2008) A modeling assessment of the interplay between aeolian iron fluxes and iron-binding ligands in controlling carbon dioxide fluctuations during Antarctic warm events. *Paleoceanography* 23:1–14, AGU identifier: PA4202 doi:10.1029/2007PA001531.
26. Sigman DM, Hain MP, Haug GH (2010) The polar ocean and glacial cycles in atmospheric CO₂ concentration. *Nature* 466:47–55.
27. Rempfer J, Stocker TF, Joos F, Dutay J, Siddall M (2011) Modeling Nd-isotopes with a coarse resolution ocean circulation model: Sensitivities to model parameters and source/sink distributions. *Geochim Cosmochim Acta* 75:5927–5950.
28. Piotrowski AM, et al. (2009) Indian Ocean circulation and productivity during the last glacial cycle. *Earth Planet Sci Lett* 285:179–189.
29. Piotrowski AM, Goldstein SL, Hemming SR, Fairbanks RG (2005) Temporal relationships of carbon cycling and ocean circulation at glacial boundaries. *Science* 307:1933–1938.
30. Guihou A, et al. (2010) Late slowdown of the Atlantic Meridional Overturning Circulation during the last glacial inception: New constraints from sedimentary (²³¹Pa/²³⁰Th). *Earth Planet Sci Lett* 289:520–529.
31. Oliver KIC, et al. (2010) A synthesis of marine sediment core $\delta^{13}\text{C}$ data over the last 150,000 years. *Clim Past* 6:645–673.
32. Anderson RF, et al. (2009) Wind-driven upwelling in the Southern Ocean and the deglacial rise in atmospheric CO₂. *Science* 323:1443–1448.
33. Köhler P, Fischer H, Munhoven G, Zeebe RE (2005) Quantitative interpretation of atmospheric carbon records over the last glacial termination. *Global Biogeochem Cycles* 19:1–24, AGU identifier: GB4020 doi: 10.1029/2004GB002345.
34. Schaefer H, et al. (2011) On the suitability of partially clathrated ice for analysis of concentration and $\delta^{13}\text{C}$ of palaeo-atmospheric CO₂. *Earth Planet Sci Lett* 307:334–340.
35. Leduc G, Thouveny N, Bourlès DL, Blanchet CL, Carcaillet JT (2006) Authigenic ¹⁰Be/⁹Be signature of the Laschamp excursion: A tool for global synchronization of paleoclimatic archives. *Earth Planet Sci Lett* 245:19–28.
36. Marchal O, Stocker TF, Joos F (1998) Impact of oceanic reorganizations on the ocean carbon cycle and atmospheric carbon dioxide content. *Paleoceanography* 13:225–244.
37. Schmittner A, Galbraith ED (2008) Glacial greenhouse-gas fluctuations controlled by ocean circulation changes. *Nature* 456:373–376.
38. Bouttes N, Roche DM, Paillard D (2011) Systematic study of the fresh water fluxes impact on the carbon cycle. *Clim Past* 7:1363–1392.
39. Lüthi D, et al. (2008) High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature* 453:379–382.
40. Stenni B, et al. (2010) The deuterium excess records of EPICA Dome C and Dronning Maud Land ice cores (East Antarctica). *Quat Sci Rev* 29:146–159.
41. Lisiecki LE, Raymo ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20:1–17, AGU identifier: PA1003 doi:10.1029/2004PA001071.

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