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Temporal evolution of mechanisms controlling ocean carbon uptake during the last glacial cycle

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ABSTRACT

Many mechanisms have been proposed to explain the ∼85–90 ppm decrease in atmospheric carbon dioxide (CO_2) during the last glacial cycle, between 127,000 and 18,000 yrs ago. When taken together, these mechanisms can, in some models, account for the full glacial–interglacial CO_2 drawdown. Most proxy-based evaluations focus on the peak of the Last Glacial Maximum, 24,000–18,000 yrs ago, and little has been done to determine the sequential timing of processes affecting CO_2 during the last glacial cycle. Here we use a new compilation of sea-surface temperature records together with time-sequenced records of carbon and Nd isotopes, and other proxies to determine when the most commonly proposed mechanisms could have been important for CO_2 drawdown. We find that the initial major drawdown of 35 ppm 115,000 yrs ago was most likely a result of Antarctic sea ice expansion. Importantly, changes in deep ocean circulation and mixing did not play a major role until at least 30,000 yrs after the first CO_2 drawdown. The second phase of CO_2 drawdown occurred ∼70,000 yrs ago and was also coincident with the first significant influences of enhanced ocean productivity due to dust. Finally, minimum concentrations of atmospheric CO_2 during the Last Glacial Maximum resulted from the combination of physical and biological factors, including the barrier effect of expanded Southern Ocean sea ice, slower ventilation of the deep sea, and ocean biological feedbacks.

1. Introduction

Many hypotheses have been put forth to explain the 80–100 ppm changes in atmospheric carbon dioxide (CO_2) concentrations that occurred during glacial–interglacial cycles over the past 800,000 yrs, including physical and biological changes affecting the partitioning of carbon between the ocean and the atmosphere (Sigman et al., 2010) and changes in volcanic emissions of CO_2 (Huybers and Langmuir, 2009).

Many hypotheses put forth to explain glacial–interglacial CO_2 change focus on the total amplitude of CO_2 change that can be attributed to individual mechanisms, and then consider the total amount of CO_2 that could be sequestered at equilibrium by summing up various mechanisms until they account for the total change observed (Kohfeld and Ridgwell, 2009). However, the mechanisms of CO_2 removal need not have occurred simultaneously (Gildor and Tziperman, 2001; Peacock et al., 2006; Sigman et al., 2010). While CO_2 release to the atmosphere during the last deglaciation occurred rather abruptly over the ∼10,000 yrs following the Last Glacial Maximum (hereafter LGM, ∼18–24 ka), CO_2 removal from the atmosphere occurred through multiple steps between the last Interglacial period ∼127,000 yrs ago and the LGM. We focus on three intervals of CO_2 drawdown: the first occurred 115–100 ka, the second from 72–65 ka, and the final step, after a brief return to slightly higher CO_2 levels, occurred 40–18 ka (Fig. 1). The steps are roughly equivalent to the transitions between Marine Isotope Stages (MIS) 5e to 5d, MIS 5a to MIS 4, and the end of MIS 3 to MIS 2, respectively. How these steps of atmospheric CO_2 reductions are linked with specific physical and biological mechanisms remains an open question.

Both process-based modeling (Kohfeld and Ridgwell, 2009; Sigman et al., 2010; Hain et al., 2014) and paleo-environmental data reconstructions (Kohfeld et al., 2005) suggest that marine biology feedbacks are important but cannot account for the full 80–100 ppm magnitude of change, and cannot explain CO_2 drawdown during the early part of the glacial cycle. These studies implicate physical processes as important drivers of ocean carbon uptake. Some proposed physical mechanisms include a) reduced air-sea gas exchange in response to sea ice expansion (Stephens and Keeling, 2000; Ferrari et al., 2014), b) a decrease

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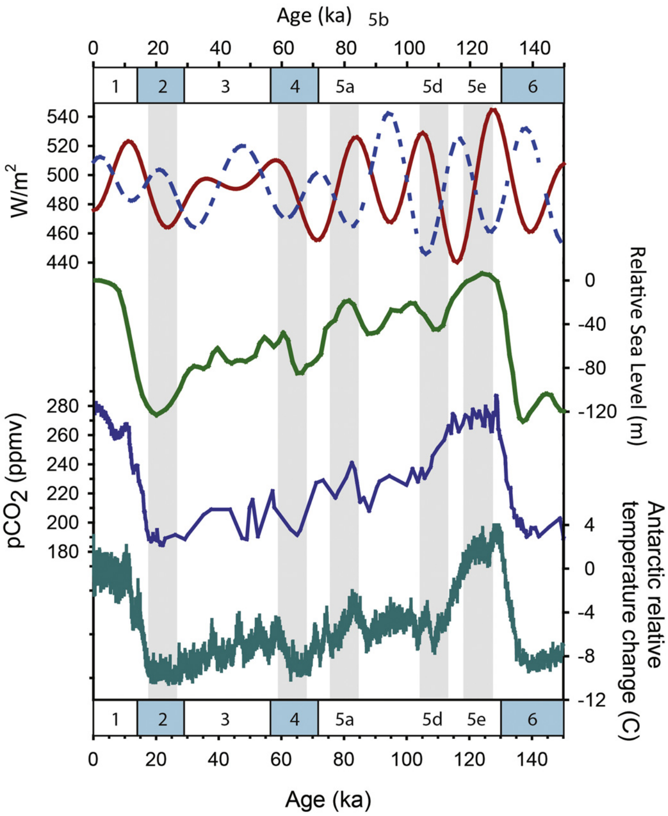

Fig. 1. (**A**) June insolation at 60°N (red line) and December insolation at 60°S (blue dashed line) (Berger and Loutre, 1991), (**B**) relative sea level (green line) reconstructed from benthic foraminiferal oxygen isotope records (Waelbroeck et al., 2002), (**C**) atmospheric carbon dioxide concentration (blue) (Jouzel et al., 2007) and (**D**) ice core temperatures reconstructed from Antarctica (EPICA Dome C, grey) (Jouzel et al., 2007). Marine Isotope Stages (MIS) 5e, 5d, 5a, 4, 3, 2, and 1 are indicated at the top and bottom of the figure; grey shading indicates the 10,000-yr time periods over which characteristics are averaged in Fig. 4.

in the rate of supply of nutrients and CO_2 to polar surface waters, initially called 'polar stratification' (François et al., 1997; Sigman et al., 2004), and c) slower ventilation of the deep ocean, leading to increased isolation of deep waters from the atmosphere (Toggweiler, 1999; Watson and Naveiro Garabato, 2006; Adkins, 2013; De Boer and Hogg, 2014).

Here we use a new compilation of sea-surface temperatures (SSTs) from 136 deep-sea core records for the past 130,000 yrs (Fig. 2; Supplementary Tables A.1–A.2) to constrain the physical mechanisms responsible for oceanic uptake of CO_2 during the most recent glacial cycle. We use SST because this variable provides a critical link between the atmosphere and ocean, influencing processes such as buoyancy forcing (Watson et al., 2015) and sea ice formation (Gordon, 1981). Furthermore, SST is the focus of many paleoceanographic reconstructions, and therefore has near-global coverage, high temporal resolution, and broad agreement between multiple proxy types (see Supplementary Information). We use this compilation, together with observational constraints from sea ice and ocean circulation proxies, to evaluate which mechanisms were acting during glaciation, and to develop a plausible sequence of events that enhanced CO_2 uptake by the world's oceans during the last glacial cycle. Changes in ocean circulation are constrained using a global compilation of the carbon isotopic composition of benthic foraminifera (Oliver et al., 2010) and supported by more limited reconstructions using the Nd isotope proxy (Piotrowski et al., 2005; Böhm et al., 2014; Jonkers et al., 2015; Wilson et al., 2015). To track changes in Southern Ocean sea ice, we use the ice core proxy of sea salt sodium (ssNa) fluxes (Wolff et al., 2010) and diatom-based proxies from marine sediment cores (Gersonde and Zielinski, 2000; Crosta et al., 2004).

2. Materials and methods

We compiled SST reconstructions and high latitude North Atlantic faunal assemblages that extend from 130,000 yrs ago to today (Fig. 2). The compilation includes SST data reconstructed using a range of techniques, including alkenones (52 sites), Mg/Ca ratios (16 sites), and faunal assemblage reconstructions (planktonic foraminifera, diatoms, and radiolaria, 78 sites; Fig. A.1). These estimates come from a total of 136 deep-sea cores distributed between 72°N and 57°S (Tables A.1 and A.2). We examine conditions for several time slices of comparable length during the last glacial cycle: Marine Isotope Stages 5e (118–127 ka), 5d (105–114 ka), 5a (75–85 ka), 4 (59–68 ka), 3 (45–55 ka), 2 (18–28 ka), and 1 (0–9 ka) (Table A.3).

The cores used to reconstruct changes in SST are found between 56°S and 57°N and thus do not provide adequate coverage of surface ocean changes in the highest polar latitudes. We supplement SST reconstructions using data on relative percentages of *N. pachyderma* (s.) from North Atlantic sites between 60 and 76°N (10 sites; Tables A.1–A.2 and Fig. A.2). Many previous studies have shown that the relative percentage of *N. pachyderma* (s.) in the planktonic foraminiferal assemblage provides a reasonable, first order estimate of mean annual surface temperatures between 3 and 12 °C in the North Atlantic Ocean ($R^2 = 0.83$, Kohfeld et al., 1996).

We have used the latest published age model associated with each deep-sea core with minor modification. The published age

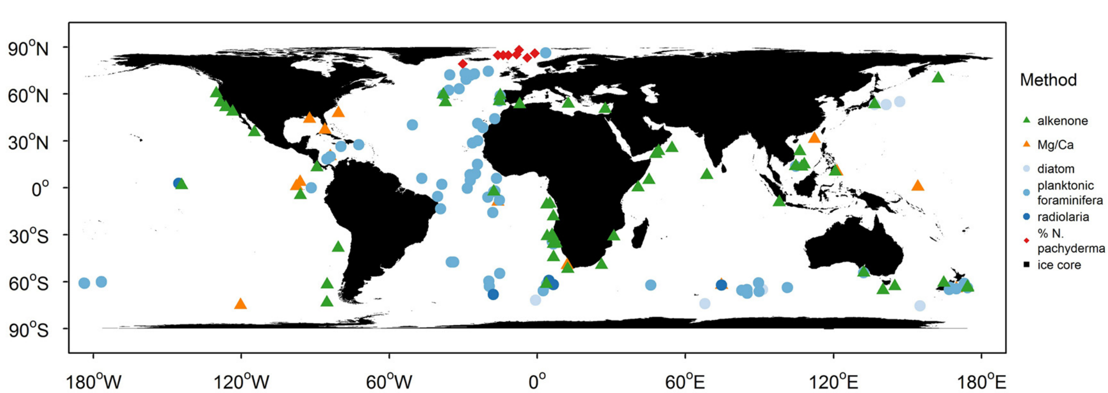

Fig. 2. Core locations for sea surface temperature estimates made from alkenone (green triangles), Mg/Ca ratios from planktonic foraminiferal calcite (orange triangles), and faunal assemblages (circles) of planktonic foraminifera (blue), diatoms (light blue) and radiolaria (dark blue). Red diamonds show locations of % *N. pachyderma* (s.) sites, and black square denotes location of the EPICA Dome C ice core.