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# The simulated climate of the Last Glacial Maximum and the insights into the global carbon cycle

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## Abstract

Global climate models (GCMs) provide an important tool for simulating the earth's climate. Here we present a GCM simulation of the climate of the Last Glacial Maximum (LGM), which was obtained by setting atmospheric greenhouse gas concentrations

- <sup>5</sup> and the earth's orbital parameters to the values which prevailed at 21 000 years before present (BP). During the LGM, we simulate a significant cooling of the ocean and a dramatic expansion of the sea-ice extent. This behaviour agrees with reconstructions from paleoclimate archives. In the ocean, the LGM simulation produces a significant redistribution of dissolved oxygen and carbon. The oxygen levels rise and the volume
- <sup>10</sup> of anoxic water declines by more than 50 %, which is consistent with paleoclimate reconstructions of denitrification. The simulated LGM climate also stores more carbon in the deep ocean (below 2000 m), but with a reduced atmospheric  $CO_2$  level the total carbon stored in the ocean declines by 600 Pg C. The LGM ocean circulation preconditions the ocean to store carbon in the deep; however, the ocean circulation and
- <sup>15</sup> sea-ice changes are insufficient alone to increase the total carbon stored in the ocean and modifications to the ocean biogeochemical cycles are required. With modifications to organic and inorganic carbon export and organic carbon remineralization one can increase ocean carbon storage (240 Pg C) to a level that is sufficient to explain the reduction in atmospheric and land carbon during the LGM (520 ± 400 Pg C). With the
- <sup>20</sup> modified biogeochemical cycling in the ocean, the simulated aragonite lysocline depth and dissolved oxygen become more consistent with paleo-reconstructions.

#### 1 Introduction

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Over recent glacial cycles, global temperature and atmospheric  $CO_2$  are highly correlated with an 80–100 ppm lower atmospheric  $CO_2$  concentration during colder glacial cycles (Sigman et al., 2010). The atmospheric  $CO_2$  reduction during the glacial periods

provides a global climate forcing that may help explain the cooling during glacial peri-



ods. However, the cause of the reduction in CO<sub>2</sub> during glacial periods remains a fundamental unresolved question for earth system science (Archer et al., 2000; Brovkin et al., 2007). Given the much larger carbon storage in the ocean than either the land or the atmosphere, it is widely accepted that the ocean must dominate glacial/interglacial changes in atmospheric CO<sub>2</sub> (Peacock et al., 2006).

The most promising explanations for the atmospheric  $CO_2$  decrease involve ocean biogeochemistry and its interaction with ocean circulation (Sigman et al., 2010). Potential causes include increased iron supply to the ocean (Martin, 1990), increased sea-ice cover (Stephens and Keeling, 2000), changes in the availability of nutrients (Procedure and Handerson, 1000), circulation changes (Taggweiler and Buscell, 2006).

- <sup>10</sup> (Broecker and Henderson, 1998), circulation changes (Toggweiler and Russell, 2006), increased ocean alkalinity (Archer et al., 2000) and some combination of all the above. These explanations for the reduction in atmospheric  $CO_2$  during glacial periods identify the importance of the ocean circulation and its effect on the global efficiency of the biological pump.
- The response of the earth system to the conditions of the Last Glacial Maximum (LGM) provide a useful assessment of our ability to simulate the earth's climate (Falkowski et al., 2000). We use a global climate model (GCM) to simulate the conditions during the LGM. Our study assesses the ability of the GCM to simulate the LGM climate and then explores how the simulated ocean state during the LGM affects
- the carbon content and distribution in the ocean. Following Tagliabue et al. (2009), we employ global 3-D climate and ocean models to enable a spatial assessment of the simulated behaviour. Using our simulated LGM ocean state, we provide a new perspective on the mechanisms responsible for the 80–100 ppm drawdown in atmospheric CO<sub>2</sub> during glacial cycles. The simulations also provide an additional perspective on
- the distribution of biogeochemical (BGC) tracers in the ocean, which can be used to assess the simulated LGM climate, and the changes in oxygen concentrations and ocean acidification.



#### 2 Model and experiments

The model simulations were performed with the MK3L modelling system (Phipps et al., 2011, 2012), which includes atmosphere–land–sea-ice and ocean components with a simple biogeochemical module embedded in the ocean model (Matear and Lenton,

<sup>5</sup> 2014). The horizontal resolution of the atmosphere–land–sea-ice module is 5.6° × 3.2° east–west and north–south respectively, with 18 vertical levels. The ocean module has a horizontal resolution of 2.8° × 1.6° with 21 vertical levels. The modelling system is flexible and simulations of atmosphere, land, sea-ice and ocean components can be done either separately or in coupled mode. For this study, we used both coupled and ocean-only simulations. A description of the physical modelling system is presented in Phipps et al. (2011).

The Holocene climate (HC) was obtained by integrating the coupled model for 5000 years using an atmospheric  $CO_2$  value of 280 ppm and present-day orbital parameters. The LGM climate was obtained by setting the orbital parameters to the conditions

- <sup>15</sup> 21 ka BP and the equivalent atmospheric CO<sub>2</sub> to 167 ppm. The coupled system was then integrated to quasi-equilibrium (5000 years). The equivalent CO<sub>2</sub> was reduced to 167 ppm following the Paleo Modelling Inter-comparison Project 2 (PMIP2) experiment design to account for the reduction in CO<sub>2</sub>/CH<sub>4</sub>/N<sub>2</sub>O concentration from the HC values of 280 ppm/760 ppb/270 ppb to the LGM values of 185 ppm/350 ppb/200 ppb
- (see http://pmip2.lsce.ipsl.fr/). In the LGM simulation, the atmospheric CO<sub>2</sub> was prescribed and the climate state was obtained without considering the feedback from the ocean.

The PMIP2 experimental design for a LGM simulation prescribes changes in greenhouse gases and orbital parameters, which we applied. Changes in ice sheet extent, topography and coastline, however, were not applied in our simulations, hence our simulation could be described as being a "simplified version" of PMIP2 experimental design. While we do not take into account the topographic effects of increased land ice, the model is nonetheless capable of simulating a shift in the ice-albedo feedbacks



through changes in sea-ice cover and the position of the snowline through its dynamic sea and land snow schemes.

Once the HC and LGM climates were reached, 50 year averages of the simulated sea surface temperature (SST), sea surface salinity (SSS), wind stress, sea-ice concentra-

tion and incident solar radiation from the respective simulated climates were used to force the ocean-only simulations. For the ocean-only simulations, the heat and freshwater fluxes into the ocean were determined by relaxing the SST and SSS fields to the coupled model monthly fields with a 20 day timescale.

The ocean-only simulations were employed to explore ocean BGC changes in the LGM climate. Table 1 summarizes the different BGC experiments, which will be discussed in more detail in the next section.

With the ocean-only model, 10000 year simulations were performed to ensure the ocean carbon cycle was in steady state and these fields were then used in our analysis. To assess whether the BGC behaviour in the coupled model differed from the ocean-

- only experiments, we ran the coupled model with ocean BGC for another 1000 years, using the BGC fields from the steady-state ocean-only simulations. This assessment was done for both the HC and LGM climates. For key diagnostics like meridional over-turning, ocean carbon content and global export production the behaviour of the ocean-only simulation differed by less than 1 % from the coupled simulation. Hence, we only
- present BGC results from the ocean-only simulations. Given the computational speed of the ocean-only simulations, they provided an ideal platform to test the sensitivity of BGC fields to the parameterizations used in the BGC module. A detailed description of the BGC module is provided in Appendix, and the next section will present our results.

#### 3 Results and discussion

In presenting our simulations, we focus on the difference between the HC and LGM. We first show the simulated physical changes in the ocean and sea-ice, and then compare them to LGM reconstructions. We then present how the ocean biogeochemical fields



differ between the HC and LGM climates with a fixed ocean BGC parameterization. Finally, we explore how modifying the BGC parameterization alters the ocean carbon storage and key ocean BGC fields like dissolved oxygen and aragonite saturation state.

# 3.1 LGM climate evaluation

### 5 3.1.1 Temperature

The LGM-simulated surface ocean is much colder than the HC (3.2°C, see Table 2) with the greatest cooling in the annual averaged SST occurring in the North Pacific, North Atlantic and the Pacific sector of the Southern Ocean (Fig. 1). In comparison to paleo-reconstructions, our simulated LGM cooling was more than 1°C colder than the Waelbroeck et al. (2009) estimates (Table 2), and more comparable to Ballantyne 10 (2005) reconstructed SST cooling over the tropical ocean ( $30^{\circ}$  S– $30^{\circ}$  N) of  $2.7 \pm 0.5^{\circ}$ C. Our simulation did not produce a large difference in the averaged value of SST cooling between the Atlantic, Pacific and Indian Oceans (less than 0.3°C difference). In contrast, the Waelbroeck et al. (2009) paleo-reconstruction shows more than a 1°C greater cooling in the Atlantic Ocean than the global average (Table 2), while the Bal-15 lantyne (2005) reconstructions had regional tropical coolings of 3°C in the Atlantic, 2.5°C in the Indian, and 1°C in the central Pacific Oceans. Perhaps some of the discrepancies between the simulated LGM cooling and paleo-reconstructions reflect the large spatial variability in SST cooling and the limited samples available for the paleo-

<sup>20</sup> reconstructions.

As synthesised by Otto-Bliesner et al. (2009), the paleo-reconstructions of SST difference between the HC and LGM based on foraminiferal assemblages display large spatial variability in the tropical oceans. In the eastern tropical Atlantic (east of 25° W), the LGM cooling is in excess of 3°C with several cores suggesting cooling of greater than 5°C (Kucera et al., 2005). In contrast, LGM cooling in the western tropical Atlantic

than 5 °C (Kucera et al., 2005). In contrast, LGM cooling in the western tropical Atlantic Warm Pool is only 1 to 2 °C (Otto-Bliesner et al., 2009). In the tropical Pacific, most cores show LGM cooling of less than 3 °C, however several cores suggest cooling of



greater than 6 °C (Kucera et al., 2005). In contrast, the LGM cooling in the Western Pacific Warm Pool is less than 1 to 2 °C (Chen et al., 2005) and several cores in the central tropical Pacific show little or no cooling in the LGM (Otto-Bliesner et al., 2009).

The simulated change in SST between the HC and LGM show similar spatial variability to paleo-reconstructions, with greatest cooling in the equatorial oceans and high latitudes, and least cooling in the subtropics (Fig. 1d). Like the paleo-reconstructions, the LGM simulation has less cooling in the tropical West Pacific than the East Pacific (Fig. 1d). While our simulated LGM cooling was greater than the paleo-reconstructions, it was generally within the existing spread of paleo-reconstructions and the spatial variability they displayed.

## 3.1.2 Sea-ice extent

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Associated with the colder SST, our LGM simulation has much greater sea-ice cover than the HC. The simulated maximum seasonal sea-ice cover in the LGM included a large portion of the North Pacific, North Atlantic and Southern Ocean as far north as 45° S (Fig. 2).

The simulated maximum Southern Hemisphere sea-ice cover in the LGM ( $35 \times 10^6 \text{ km}^2$ ) is approximately double the simulated HC value ( $15 \times 10^6 \text{ km}^2$ ). A reconstruction of the maximum sea-ice extent (sea-ice concentration greater than 15%) shows sea-ice extending to 47° S in the Atlantic and Indian sector and 57° S in the Pacific sec-

<sup>20</sup> tor – a northward displacement of 7 to 10° in latitude (Gersonde et al., 2005), which is comparable to a our LGM simulation (Fig. 2). From the LGM reconstructions, the maximum sea-ice extent in the Southern Hemisphere expands to about  $39 \times 10^{6}$  km<sup>2</sup> (Gersonde et al., 2005), which represents an approximate doubling of ice cover from the modern value ( $19 \times 10^{6}$  km<sup>2</sup>; Comiso, 2003). Importantly, for both HC and LGM the simulated maximum Southern Hemisphere sea-ice cover, 15 and  $35 \times 10^{6}$  km<sup>2</sup> respec-

tively, were similar to the observed and reconstructed values.

In the Northern Hemisphere, the simulated maximum sea-ice extent increased from the HC value of  $12 \times 10^6$  km<sup>2</sup> to a LGM value of  $29 \times 10^6$  km<sup>2</sup>, a more than doubling the



sea-ice extent. Sarnthein et al. (2003) LGM reconstructions give a North Atlantic winter maximum sea extent that extends south of Iceland then across to the southern tip of Labrador, which is similar to the simulated LGM maximum sea-ice extent in the North Atlantic (Fig. 2). Both the paleo-reconstructions (Sarnthein et al., 2003) and our LGM
 simulation show a dramatic retreat of the minimum sea-ice extent to north of Greenland during the summer (Fig. 2).

In the North Pacific, the simulated maximum LGM sea-ice extent went as far south as 45° N, with much greater sea-ice cover in the western Pacific than the eastern Pacific. Consistent with our LGM simulation, paleo-reconstructions suggest that LGM winter sea-ice extent was greater in the western Pacific (Nürnberg and Tiedemann, 2004), with no sea-ice cover at 50° N and 167° E in the Central North Pacific (Jaccard et al., 2005).

During the summer season, the simulated LGM minimum seasonal sea-ice extent rapidly contracts to cover 6.5 and  $11.4 \times 10^{6}$  km<sup>2</sup> in the Southern and Northern Hemi-<sup>15</sup> spheres, respectively, which is less than 1/3 of the maximum sea-ice extent (Fig. 2). In contrast, the HC simulation had sea ice cover of 1.7 and  $5.3 \times 10^{6}$  km<sup>2</sup> in the Southern and Northern Hemispheres, respectively. The simulated minimum sea-ice extent was much greater in the LGM than the HC with areas in the North Pacific, North Atlantic and Southern Ocean where sea-ice was present throughout the year.

#### 20 3.1.3 Meridional overturning

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Associated with the changes in the surface ocean during the LGM were changes in the meridional overturning circulation. In the LGM simulation, the rate of Antarctic Bottom Water (AABW) formation increased to 15 from 7 Sv in the HC (Fig. 3, minimum in high latitude Southern Hemisphere Global overturning). The LGM simulation also showed
 <sup>25</sup> greater subduction of Southern Ocean intermediate water than the HC simulation. The simulated LGM North Atlantic Deep Water (NADW) overturning (15 Sv) was similar to the HC simulation (17 Sv), but the cell was shallower in the LGM simulation (maximum



in the North Atlantic overturning). The LGM simulated meridional overturning showed that the deep water was dominated by water from the Southern Ocean.

The simulated features of meridional overturning in the LGM are supported by the paleo-reconstructions. Paleonutrient tracers indicate that the boundary between NADW

- and AABW was substantially shallower during the LGM than today (Otto-Bliesner et al., 2007). In our simulations, the NADW cell rises from a maximum depth of 3000 m in the HC to a depth of 1500 m in the LGM. During the LGM, paleonutrient tracers (Yu et al., 1996; McManus et al., 2004) also suggest the North Atlantic Ocean was more stratified, with a shoaling of the NADW cell, and the AABW penetrating much farther into the
- North Atlantic than at present. The greater contribution of AABW to the deep water of the North Atlantic is also consistent with Atlantic sediment cores that indicate that in the LGM the Atlantic deep waters were much colder and saltier than modern day (Adkins, 2002). The paleo-reconstructions imply that the bottom waters of the LGM were considerably more saline than today and that the deep ocean was more stratified, with the
- stratification being due to salinity rather than temperature as the bottom temperatures were close to freezing throughout much of the world ocean (Adkins, 2013). The simulated salinity distribution in the LGM was consistent with paleo-reconstructions (Fig. 4). As postulated by Watson and Garabato (2006), the very cold and saline bottom water was the consequence of more production of brines from greater sea-ice formation in the LGM simulation.

Originally, the shoaling of the NADW and greater contribution of AABW to the North Atlantic was interpreted as a significant reduction in NADW formation (McManus et al., 2004; Yu et al., 1996), but more recent paleonutrient tracers suggest that during the LGM, NADW was similar or slightly reduced compared to HC (Otto-Bliesner et al.,

<sup>25</sup> 2007). This slight reduction and shoaling of the NADW in the LGM was consistent with our simulations.



## 3.2 Ocean biogeochemical fields

The colder surface water coupled with greater subduction of AABW in the LGM simulation filled the deep LGM ocean with water of a much greater oxygen concentration than the HC (Fig. 5). The LGM-simulated oxygen concentration in the ocean was more than  $100 \text{ mmol m}^{-3}$  greater than the HC simulation with the simulated LGM global oxygen

content about 55 % greater than the HC simulation (Table 3).

For dissolved inorganic carbon (DIC), the difference between the LGM and HC simulations was more complicated than oxygen (Fig. 6). In the deep water (below 2000 m), the DIC concentrations in the LGM simulation were greater than the HC simulation, consistent with the subduction of more colder AABW during the LGM, which, like oxy-

- <sup>10</sup> consistent with the subduction of more colder AABW during the LGM, which, like oxygen, elevated DIC concentrations in the deep ocean. However, in the upper ocean during the LGM, the simulated DIC concentrations were much less than the HC simulation, which reflected the lower atmospheric  $CO_2$  concentration in the LGM than in the HC: 185 vs. 280 ppm. The reduction in atmospheric  $CO_2$  between the HC and LGM
- <sup>15</sup> meant the surface water equilibrated with a much lower atmospheric CO<sub>2</sub> level in the LGM and subsequently reduced DIC concentrations. The reduced DIC concentration exceeded 200 mmol m<sup>-3</sup> in the surface water of the subtropical and tropical ocean. The reduced DIC concentrations extended furthest into the interior in the downwelling regions of the subtropical gyres. There was also a similar decline in the high latitude
- Northern Hemisphere in the LGM simulation associated with reduced atmospheric CO<sub>2</sub> and freshening of surface water.

In the LGM simulation, alkalinity concentrations (like salinity) were much greater in the deep ocean than the HC simulation (Fig. 7). Compared to the HC, LGM-simulated alkalinity decreased in the water above 1500 m and increased in the water below

<sup>25</sup> (Fig. 7). The zonal averaged change in the simulated alkalinity was very similar to the salinity change between the LGM and HC (Fig. 4).

The large increase in the DIC concentrations in the deep ocean in the LGM simulation reflects changes in the subduction of DIC-rich polar water and changes in the



remineralization of organic matter in the ocean interior. From the apparent oxygen utilization (AOU) of the interior waters and the prescribed ratio of O/P/C of particulate organic matter (POM) used in the model, we computed the amount of phosphate and DIC produced from the remineralization of sinking POM. In the LGM simulation, the

- <sup>5</sup> remineralized phosphate was much less than the HC simulation, with the total amount of remineralized phosphate being only 40% of the HC simulation. While the amount of remineralized phosphate declined in the LGM simulation, the phosphate concentrations in the deep water (below 1500 m) were greater than the HC simulation (Fig. 8). The increase in phosphate in the LGM simulation reflects the increased subduction of
- <sup>10</sup> phosphate-rich AABW which transfers phosphate into the abyss and reduces the phosphate concentrations in the upper ocean (Fig. 8). Paleo cadmium and  $\delta^{13}$ C data from the tropical and North Atlantic show increased nutrient levels in the abyss, but reduced levels above 2000 m during the LGM (Boyle, 1992; Marchitto and Broecker, 2006; Tagliabue et al., 2009). This redistribution of nutrients in the LGM was consistent with our
- <sup>15</sup> simulated LGM phosphate distribution. The LGM simulation relies on increased AABW subduction to transfer nutrients from the upper ocean into the abyss and elevate DIC, phosphate, alkalinity and oxygen concentrations below 2000 m.

Associated with the transfer of phosphate from the upper ocean into the abyss in the LGM simulation there was a global reduction in the export of POC from the euphotic

- <sup>20</sup> zone (Table 3). The LGM simulation (Fig. 9) was consistent with paleonutrient data which suggests export production in the Antarctic Zone of the Southern Ocean was lower during the LGM (Jaccard et al., 2005). Similarly, the simulated decline in North Pacific export production (Fig. 9) was consistent with paleonutrient data from the Sub-arctic Northwest Pacific, which indicates less export production during glacial periods
- (Jaccard et al., 2005). The simulated reduction of export production in the Subarctic Northwest Pacific and North Atlantic, in regions not covered by sea ice, is caused by reduced phosphate availability (Fig. 10). The equatorial upwelling region of the east Pacific and east Atlantic were two other regions in the LGM simulation where the sur-



face phosphate concentration declined and there was a reduction in export production in the LGM.

The increased AABW subduction in the LGM simulation supported a greater vertical gradient in DIC, alkalinity and phosphate, while the reduction in POC and particulate
 <sup>5</sup> inorganic carbon (PIC) export in the LGM simulation partially reduced these vertical gradients. Comparing the simulated DIC and phosphate concentrations in the ocean interior in the HC to the LGM showed that remineralization of POC in the LGM declined by 60 %. The decline in remineralized DIC and phosphate (60 %) exceeded the decline in POC export from the euphotic zone (44 %), which implied the LGM circulation had increased resupply of remineralized phosphate and carbon back to the upper ocean. In general, outside of the Southern Ocean the surface water became phosphate limited.

## 3.3 Carbon budget

As discussed in the introduction, the strength of deep water formation, the suppression of air-sea gas exchange due to sea-ice expansion in the Southern Ocean and the in-

<sup>15</sup> crease in POC export, have all been suggested as playing possible roles in reducing atmospheric CO<sub>2</sub> during glacial periods (Otto-Bliesner et al., 2007). Our LGM simulation had the first two features, and in the following discussion we also explore how increasing POC export could impact the LGM ocean carbon content.

In our coupled simulations, we fixed atmospheric CO<sub>2</sub> concentrations in the LGM and

- <sup>20</sup> HC, and then simulated the carbon content of the ocean. The ocean carbon content in the LGM simulation was 604 Pg C less than the HC simulation (Table 3). The changes in the simulated LGM ocean carbon reflect changes in the ocean state and the reduction in the atmospheric  $CO_2$  level in the LGM. To separate the impact of changes in the ocean state from changes in atmospheric  $CO_2$  we used a LGM ocean-only sim-
- <sup>25</sup> ulation where the atmospheric  $CO_2$  was set at 280 ppm (OLGM-2). By increasing the atmospheric  $CO_2$  in the LGM climate, the carbon content in the ocean was 1130 PgC greater than the HC simulation (OLGM-2, Table 3), which was 1730 PgC greater than the standard LGM simulation (OLGM-1). Thus, the ocean state during the LGM was



preconditioned to increase the carbon content of the ocean, but this preconditioning was not enough to counter the effect of reduced atmosphere  $CO_2$  during the LGM, which caused a loss of carbon from the upper ocean (Fig. 6).

While the LGM ocean was conditioned to store more carbon in the deep ocean,
changes in BGC cycles appear necessary to get the carbon into the ocean. To explore potential biogeochemical processes that could increase the carbon storage in the LGM ocean we considered three plausible modifications to the standard BGC formulation:
(1) increased POC export, (2) increased depth of POC remineralization, (3) reduced PIC export. The resulting impacts on the carbon storage in the ocean are summarized in Table 3 and discussed in the next few paragraphs.

To increase POC export, the scaling factor for POC export was increased by  $10 \times$  (see Eq. 34 in Appendix). The motivation for increased POC export in the LGM is the increased iron supply to the ocean, which relieves iron limitation in regions like the Southern Ocean and increases POC export (Martin, 1990; Tagliabue et al., 2009).

- <sup>15</sup> Although the POC export scale factor was increased by 10×, the POC export increase was much smaller (1.3×) because in the LGM simulation most of the oceans outside of the Southern Ocean were phosphate limited (Fig. 10a) and weakly responded to the increased scaling factor for POC export; the increase in export production mostly occurred on the northern boundary of the Southern Ocean (Fig. 9d). By increasing the
- <sup>20</sup> POC export scaling factor, POC export increased by 1.5 to  $5.9 \text{ Pg C y}^{-1}$ , which was still less than the HC value of 8.1 Pg C y<sup>-1</sup>, and the carbon content of the ocean increased by 260 Pg C (Table 3).

The POC remineralization depth was increased by changing the power law exponent from -0.9 to -0.7 (see Eq. 010, Appendix). The motivation for such a change is that a colder ocean would reduce the rate of bacterial remineralization of POC (Rivkin, 2001). The change in remineralization increased the fraction of POC sinking through 1000 m from 12.5 to 20%. Increasing the depth of POC remineralization reduced POC export from the photic zone by  $1.2 \text{ Pg C y}^{-1}$  (OLGM-5, Table 3). The deeper reminer-

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alization depth for POC increased the carbon content of the ocean by 170 PgC. The

sensitivity of the carbon content of the ocean in the LGM to changes in the depth of POC remineralization was much less than suggested by the simulation of Kwon et al. (2009). The reduced sensitivity of carbon storage in our LGM simulation to the depth of POC remineralization reflects the conclusion that the steady-state response is much smaller than the transient response (Schneider et al., 2008).

Finally, the PIC export was set to zero, which was motivated by the strong relationship between calcification and temperature (Lough and Barnes, 2000), i.e. in a cooler ocean less PIC would be produced. With no PIC export, the solubility of  $CO_2$  in the surface water increased and enabled the ocean to store an additional 260 Pg C.

Independently, none of the BGC modifications comes close to increasing the LGM storage of carbon in the ocean above the HC simulation (OLGM-3, 4 and 5, Table 3). By employing all three BGC modifications discussed above in one LGM simulation (OLGM-6), the ocean stored 830 Pg C more than the standard LGM simulation (OLGM-1; Table 4). With all three BGC modifications, the increase in the ocean carbon storage

- in the LGM simulation was 240 PgC more than the HC simulation and now comes within the bounds of the combined reduction in atmospheric carbon and terrestrial carbon reported by Ciais et al. (2011) of 520±400 PgC (Table 4). Combined, the three BGC modifications are sufficient to explain the change in land and atmospheric carbon between the HC and LGM. The consequences of these BGC modifications will next
- <sup>20</sup> be assessed by looking at how they change the stability of calcium carbonate, export production and dissolved oxygen. However, before exploring these changes we first consider other biogeochemical processes that could account for the additional carbon storage in the ocean.

The carbon stored in the ocean could be increased by increasing alkalinity input from the land due to terrestrial weathering. The  $1120 \pm 400$  Pg C would require an alkalinity increase of  $112 \pm 40$  Pmol Eq (10 % global increase). During the LGM, a drier climate, extensive ice sheets and exposed shelf areas would change terrestrial weathering but such changes are unlikely to have increased alkalinity input from the HC (Brovkin et al., 2007). Therefore, increased alkalinity input into the ocean seems an unlikely explana-



tion by itself for increasing the carbon stored in the ocean. Further, the additional alkalinity would increase the stability of calcium carbonate in the ocean, which we show in the next section was already too deep in the standard LGM simulation (OLGM-1, Fig. 12a).

<sup>5</sup> Greater cooling of the ocean in the mid and low latitudes is another potential mechanism to store more carbon in the LGM ocean. Given that our SST cooling was near the paleo-reconstruction maximum cooling this mechanism seems unlikely for storing more carbon in the ocean in our LGM simulation.

Changes in the large-scale ocean circulation could provide another way to increase carbon storage. In our LGM simulation, the increased subduction of AABW dominated the increased carbon content of the deep ocean as shown by the reduction in the regenerated phosphate concentrations in the deep ocean. Reducing AABW subduction would reduce ocean carbon storage by reducing the amount of high-DIC water subducted into the deep ocean. Therefore, increased subduction of AABW could increase the calubility nump and come carbon storage but as available at a

the solubility pump and ocean carbon storage but, as explored by Tagliabue et al. (2009), this did not greatly increase the carbon stored in the ocean. Further, our LGM simulation already has a much greater increase in AABW than other LGM simulations e.g. Tagliabue et al., 2009).

Another way of removing carbon from the land-ocean-atmosphere in the LGM is the net burial of organic carbon in ocean sediments. The required net organic carbon burial using our LGM simulation and Ciais et al. (2011) reported atmosphere and land carbon changes (1120  $\pm$  400 PgC) is not large given the long timescale involved in the drawdown of atmospheric CO<sub>2</sub> from the interglacial to glacial (100 000 years). The burial of organic carbon would cause a loss of phosphate from the ocean. Assuming

<sup>25</sup> a C/P ratio of 106 this would equate to a loss of  $10.5 \pm 4 \text{ Gmol P y}^{-1}$ . Present estimates of the burial rate of phosphorus in the ocean sediments is about 320 Gmol P y<sup>-1</sup> (Tamburini and Foellmi, 2009), which is an order of magnitude greater than the required net phosphate loss (10.5 Gmol P y<sup>-1</sup>). The small additional loss of phosphate and carbon needed to close the LGM carbon budget requires either a small increase in organic



matter burial or a small decrease in the continental supply of organic matter to the ocean that is remineralized in the ocean. Either is a feasible mechanism to account for the missing carbon in the interglacial–glacial drawdown in atmospheric CO<sub>2</sub>. Importantly, burying carbon in ocean sediments would not change the simulated lysocline depth of the LGM ocean simulation produced by standard BGC formulation (OLGM-1; Fig. 12).

The LGM ocean is preconditiond to store carbon, but just the physical changes alone are insufficient to get the required carbon into the ocean. Individually, the BGC modifications also do not get sufficient carbon into the ocean. With all three BGC modifications (increased POC export, reduce POC remineralization and no PIC export), sufficient carbon is stored in the ocean and in the next section we will use the modifications to calcium carbon stability, export production and oxygen levels in the ocean to assess these changes.

## 3.3.1 Calcium carbonate stability

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- In comparing the stability of calcium carbonate in the ocean between the LGM and HC states two key factors need to be considered the large reduction in atmospheric CO<sub>2</sub> and the large cooling of the ocean in the LGM. The former factor increases the stability of calcium carbonate while the latter factor, through its impact on carbon speciation, would reduce the stability of calcium carbonate.
- To represent the surface ocean carbon chemistry we use the aragonite saturation state ( $\Omega$ ), which is a useful indicator of calcification rates (Langdon et al., 2000). At the surface, the simulated annual averaged  $\Omega$  in the LGM was slightly less than the HC between 40° S and 40° N, but much greater poleward of these two latitudes (Fig. 11). The BGC modifications only slightly alter the surface  $\Omega$  value and for the simulations with alter the SUCCENTRY ( $\Omega$ ) there is a slightly result of the simulations of the simulation of the simulations of the simulation of the simulations of the surface  $\Omega$  value and for the simulations with alter the SUCCENTRY ( $\Omega$ ) there is a slightly result of the simulation of th
- with all the BGC modifications (OLGM-6) there is a slight increase in surface aragonite saturation state (Fig. 11). Interestingly, the simulated  $\Omega = 3.25$  isoline, the value at present used to define the location of viable coral reef conditions (Hoegh-Guldberg et al., 2007), was nearly unchanged between the LGM and HC simulations. Recent



sonar and coring in the southern portion of the Great Barrier Reef (Yokoyama et al., 2011; Abbey et al., 2011) detected the presence of drowned coral reefs in the LGM that were as far south as the present-day Great Barrier Reef. Such observations are consistent with our simulation that the southern extent of coral reefs along the east coast of Australia was not dramatically different between the LGM and HC, which is consistent with the modified BGC simulation (OLGM-6).

The changes in the carbon chemistry also extend into the ocean interior. The simulated depth of the aragonite lysocline in the LGM simulation (OLGM-1) was much deeper than the HC simulation (Fig. 12). Outside of the tropical regions, the OLGM-1 simulation's entire water column was super-saturated for aragonite, which is unrealistic.

- Simulation's entire water column was super-saturated for aragonite, which is unrealistic. Hence, processes that increase alkalinity input into the ocean and increase the carbon burial in the sediments do not provide a means of shoaling the aragonite lysocline. Individually, the three BGC modifications still have very deep lysoclines: the Southern Ocean and North Pacific Ocean lysoclines are still near the ocean bottom and more
- than 1000 m deeper than the HC simulation (Fig. 13). However, in the experiment with the three modified BGC processes employed to increase the carbon storage in the ocean (OLGM-6), the depth of the lysocline was much shallower than the OLGM-1 simulation (Fig. 12d). For this simulation (OLGM-6), the lysocline depth in the Pacific Ocean was deeper than the HC simulations but it was shallower in the North Atlantic
- and along the northern boundary of the Atlantic and Indian sectors of the Southern Ocean (Fig. 13). There is good evidence that the mean position of the lysocline did not change much between the LGM and the HC, with the average lysocline depth being less than 1 km deeper in the LGM than the HC in the North Pacific and Southern Ocean (Catubig et al., 1998), which is most consistent with the LGM simulation with the three
- <sup>25</sup> BGC modifications (OLGM-6). The paleo-reconstructions also suggest the lysocline shoaled in the Atlantic and deepened in the Pacific (Anderson and Archer, 2002), consistent with the OLGM-6 simulation. Further, recent observations from seamounts just south of Tasmania suggest deep cold water coral communities during the LGM occured slightly deeper than at present (Thiagarajan, 2013), consistent with a deepening of the



lysocline during the LGM. To have a LGM simulation agree with the paleo observations for south of Tasmania, all three BGC modifications are required (Fig. 13).

# 3.3.2 Export of particulate organic carbon

In the standard LGM simulation (OLGM-1), the export of POC from the upper ocean
 generally declines everywhere relative to the HC simulation except on the southern boundary of the eastern equatorial upwelling region of the Pacific (Fig. 14b). The setting of the export of PIC to zero had no impact on POC export, while deepening the POC remineralization increased the transfer of carbon and nutrients into the ocean interior and the POC export was generally reduced everywhere relative to the HC
 (Fig. 14d). With the POC scaling factor increased, the total POC export was still less than the HC (Table 3), but now the Southern Ocean and small areas of the North Atlantic and Northwest Pacific have increased POC export compared to the HC. The simulation with all three BGC modifications shows a very similar pattern of change as the simulation with just increased POC export (Fig. 14a and c). In this simulation, POC

export increased in the Southern Ocean and declined in the subtropics and tropics (Fig. 14a). Such changes to POC export are consistent with paleo-reconstructions of POC differences between the LGM and HC (Sigman et al., 2010). The increase in POC export in the Southern Ocean reduces the phosphate concentrations in the subducted Antarctic Intermediate Waters, which has been shown to be critical to setting the magnitude of the export production outside of the Southern Ocean (Sarmiento et al., 2004).

#### 3.3.3 Dissolved oxygen

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The greater subduction of AABW in the LGM dramatically increased the oxygen levels in the ocean and coupled with the decreased remineralization of POC in the ocean there was a large increase in oxygen concentrations in the ocean interior during the standard LGM simulation relative to the HC simulation (55% increase). In the standard LGM simulation, oxygen levels increased in both the intermediate and deep water



(Fig. 5) and the total volume of anoxic water was less than the HC simulation. With the BGC modifications to POC export and remineralization, the oxygen levels in the intermediate water decrease in the Southern Ocean and increase in the North and Equatorial Pacific (Fig. 16a, b and d). In the deep ocean, the large increase in the simulated LGM oxygen levels only start to be eroded in the simulation where POC export scaling was increased to reflect increased iron supply to the surface ocean (Fig. 17). However, it is only in the simulation with increased POC export and deeper remineral-

- ization that the oxygen levels in the deep Southern Ocean start to decline compared to the HC simulation (Fig. 17a).
  The consequence of reduced anoxic water in the LGM should be reduced dentri-
- <sup>10</sup> The consequence of reduced anoxic water in the LGM should be reduced dentrification in these regions. Recent analysis of sedimentary  $\delta^{15}$ N records suggest that global aggregate rates of N fixation and water column denitrification rates over the past 200 000 years were less active during glacial periods and more active during interglacial periods (Galbraith et al., 2004). This is consistent with our simulated decreases
- in the extent of water column anoxia during the LGM, which occurred in all LGM simulations. As Nameroff et al. (2004) and Meissner et al. (2005) suggested, changes in ventilation and water characteristics are more important factors to the variability of water column denitrification than regional changes in export production, because the decrease in anoxia is not due to reduced POC export but rather due to increased sub-
- <sup>20</sup> duction of oxygen-rich water. Such a conclusion is consistent with our LGM simulations where the volume of anoxic water showed little difference among the LGM simulations with varying BGC formulations. For example, the simulation with all three BGC modifications (OLGM-6) had only 5 % less volume of anoxic water than the standard LGM simulation (OLGM-1). While the impact of the three BGC modifications on the volume
- of anoxic water was small, these modifications did have a significant impact on oxygen levels in the intermediate and deep water. Paleo-reconstructions suggest oxygen levels in the upper 1500 m were greater in the LGM than HC (Jaccard and Galbraith, 2011). That all LGM simulations show this behaviour again demonstrates that oxygen levels in the upper ocean are primarily set by the ocean dynamics consistent with the change



in the volume of anoxic water. In contrast, the deep ocean trace metal records generally suggest a decline in LGM oxygen levels relative to the HC (Jaccard and Galbraith, 2011). Such a decline in LGM oxygen requires oxygen consumption in the deep ocean to increase by an amount that exceeds the temperature-driven increase of oxygen sol-<sup>5</sup> ubility. Such a behaviour is only achieved when all three BGC modifications are used in the LGM simulation.

## 4 Conclusions

Simulating the earth's climate is a challenging task given the complexity of the climate system, but an ideal way to test our knowledge of the climate system. We have used a 3-D GCM to simulate the climate state during the LGM by changing the orbital parameters and the concentration of greenhouse gases in the atmosphere. Compared with the HC simulation, our simulated LGM showed more than a doubling of the winter sea-ice extent in both hemispheres; a slight reduction and 1000 m shoaling of the NADW overturning; a near doubling in the strength of the AABW overturning; a reduc-

- tion in temperature and an increase in salinity of the deep water; and an increase in the vertical stratification. A recent review by Adkins (2013) highlighted that the robust changes between the HC and LGM include: a shoaling of the NADW in the Atlantic, increased contribution of AABW to the deep water and a salt-stratified deep ocean (rather than heat-stratified). These changes are reproduced by our simulations, which revealed the same confidence in using the simulations to investigate what caused the SO
- $_{20}$  provides some confidence in using the simulations to investigate what caused the 80– 100 ppm reduction in atmospheric CO<sub>2</sub> between the interglacial and glacial periods.

The response of the BGC tracers to the simulated LGM ocean state was a large increase in salinity, oxygen, phosphate, DIC and alkalinity in the deep ocean and an increase in the vertical gradient of these tracers with depth. The increase in the deep water concentrations and in the vertical gradient was due to greater subduction of AABW and not due to increased remineralization of sinking POC. The simulation identified the importance of changes in the subduction of AABW to the BGC tracers. How-



ever, with just these simulated physical changes to the LGM simulation, the storage of carbon in the ocean declined by 600 PgC from the HC. This reduction in ocean carbon in the LGM simulation reflects the reduced atmospheric CO<sub>2</sub> concentration in the LGM. A subsequent simulation using the simulated LGM climate with an atmospheric CO<sub>2</sub> concentration at the the LGM simulation state the simulated LGM climate with an atmospheric CO<sub>2</sub> concentration at the the LGM simulation state the simulated LGM climate with an atmospheric CO<sub>2</sub> concentration at the the LGM simulation state the simulation state of the simulated LGM climate with an atmospheric CO<sub>2</sub> concentration at the the simulation state of the simulating state of the simulation state of the simulation s

- <sup>5</sup> CO<sub>2</sub> concentration set to the HC value (280 ppm) shows that the LGM ocean stored 1130 Pg C more than the HC simulation, demonstrating that the simulated LGM ocean was preconditiond to store carbon in the deep ocean. Importantly, this LGM simulation had an aragonite lysocline that was too deep and deep water oxygen concentrations that were too high.
- With the simulated LGM climate we explored how to increase the carbon content of the ocean between the LGM and HC and make the simulation more consistent with the aragonite lysocline and deep water oxygen concentrations. The existing paleo data suggest a loss of  $520 \pm 400$  Pg C from the atmosphere and land during the LGM (Ciais et al., 2011). To increase the carbon stored in the LGM ocean we considered three
- <sup>15</sup> plausible modifications to the biogeochemical processes in the ocean: (1) increased POC export due to increased atmospheric iron deposition, (2) increased depth of POC remineralization due to cooler interior water, (3) no PIC export due to colder surface water. Each process increased the ocean carbon storage but separately they could not account for the total decline in atmosphere and land carbon. However, when com-
- <sup>20</sup> bined in one simulation the ocean carbon storage was 240 Pg C greater than the simulated HC ocean and now within the bounds of the required increased carbon storage (520±400 Pg C Ciais et al., 2011). The simulated differences between the HC and LGM with the three BGC modifications was also more consistent with paleo-reconstructions than the simulation without the BGC modifications. These include the presence of less
- <sup>25</sup> anoxic water, a less than 1000 m deepening of the lysocline in the North Pacific and Southern Ocean, the reduction in POC export, the increase in nutrient-limited conditions in the LGM, and deep water oxygen levels that did decline in the Southern Ocean. Further, the surface  $\Omega$  values were nearly equivalent to the HC values.



Our GCM simulations produced a plausible LGM state, which provides some confidence in our ability to model the complex climate system. With the inclusion of the BGC in the simulation we generated new information to assess the simulated climate and pursue the link between climate and the global carbon cycle. While the simulations have shown the need to modify ocean BGC processes to increase the carbon storage in the ocean we now need to better constrain these modifications. A key future effort will be to better quantify the BGC process that could increase the carbon stored in the

ocean in a manner that is consistent with 3-D paleo-reconstructions.

#### Appendix: Ocean biogeochemical model equations

- The ocean BGC module used in this study is the same one as used by Matear and Lenton (2014) and is based on an earlier model by Matear and Hirst (2003). For a complete description of the model refer to the Appendix in Matear and Lenton (2014). Here we provide the basic description of the model to enable the reader to follow the BGC modifications performed in this study.
- The ocean BGC module links the biological export of particulate organic carbon from the euphotic zone (POC) to phosphate, light, temperature and mixed layer parameterizations. The BGC model fixes the ratio of POC and particulate inorganic carbon (PIC) export from the euphotic zone (50 m) and prescribes the depth-dependent remineralization equations for POC and PIC. The key equations for these processes are summarized below.

In the euphotic zone, which is set to be the surface layer of the model (upper 50 m), biological formation of particulate organic and inorganic matter occurs. For particulate organic matter, the production of particulate organic phosphorus (POP) was defined by the following equations:

<sup>25</sup>  $V_{\text{max}} = 0.6(1.066)^T$  $F(I) = \left[1 - e^{R(I)}\right]$ 



(A1)

(A2)

$$R(I) = \frac{I(x,t)\alpha PAR}{V_{max}}$$

$$J(P) = S_{npp} V_{max} \min\left(\frac{P}{P + P_k}, F(I)\right) \Delta z$$

 $V_{\text{max}}$  is the maximum growth rate in day<sup>-1</sup>, which is a function of the surface layer temperature (*T*, [°C]). *F*(*I*) is the productivity vs. irradiance equation used to describe <sup>5</sup> phytoplankton growth, which is given as a unitless value and provides a measure of light-limited growth. *R*(*I*) is a unitless function of the light availability for growth, which comes from the incident short wave radiation (*I* in W m<sup>-2</sup>) scaled by photosynthetic active radiation PAR (unitless factor), and the initial slope (*a*) of the productivity vs. radiance curve for phytoplankton growth [day<sup>-1</sup> (W m<sup>2</sup>)<sup>-1</sup>]. *J*(*P*) gives the production of POP in mmol m<sup>-2</sup> P day<sup>-1</sup>, which is function of a tuneable parameter (*S*<sub>npp</sub>) in mmol m<sup>-3</sup> P, the thickness of the surface layer ( $\Delta z$ ) in meters and the growth limitation function. The growth limitation function compares nutrient- and light-limited growth and

uses the minimum value. The nutrient-limited term is based on the *P* of the surface layer and half-saturation uptake ( $P_k$ ) value for phosphate utilization. A constant value was used to link the POP production to the organic carbon, oxygen

<sup>15</sup> A constant value was used to link the POP production to the organic carbon, oxy and alkalinity sink as follows:

J(C) = 106J(P)(A5) J(Q) = -136J(P)(A6)

$$J(A) = -16J(P)$$
 (A7)

<sup>20</sup> The PIC export (K) in the model was linked to POC by using a fixed rain ratio of 8 % to give the following sinks of carbon and alkalinity:

 $K(C) = 0.08 \times 106 \ J(P)$  (A8)  $K(A) = 0.08 \times 2 \times 106 \ J(P)$  (A9)



(A3)

(A4)

The POC and PIC produced in the euphotic zone was instantaneously remineralized in the ocean interior over where it was produced according to the following equation:

$$r \text{POC}(z) = \text{POC} \times \left(\frac{z}{100 \,\text{m}}\right)^{-0.9}$$

$$r \text{PIC}(z) = \text{PIC} \times \exp\left[z/3500 \,\text{m}\right]$$
(A10)
(A11)

s where POC and PIC denote the exports from the euphotic zone and z is depth in meters. No remineralization occurs above 100 m.

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**Table 1.** Summary of modelling experiments performed. An O before a model name denotes that it was an ocean-only simulation.

Experiment	Model	Greenhouse Gas Forcing (CO <sub>2</sub> ppm)	Orbital Parameters	Comment
HC	Coupled	280	0 kBP	Standard BGC
LGM	Coupled	167	21 kBP	Standard BGC
Experiment	Model	Atmospheric CO <sub>2</sub> (CO <sub>2</sub> ppm)	Climate State	Comment
OHC-1	Ocean	280	HC	Standard BGC
OHC-2	Ocean	185	HC	Standard BGC
OLGM-1	Ocean	185	LGM	Standard BGC
OLGM-2	Ocean	280	LGM	Standard BGC
OLGM-3	Ocean	185	LGM	No PIC export
OLGM-4	Ocean	185	LGM	10× POC export scaling increase
OLGM-5	Ocean	185	LGM	Increased depth of POC remineralization*
OLGM-6	Ocean	185	LGM	BGC modifications of OLGM-3, 4 and 5

\* Power law exponent for POC remineralization changed from -0.9 to -0.7.



Discussion Pa	<b>CPD</b> 11, 1093–1142, 2015					
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**Table 2.** SST temperature calculated from LGM–HC from the reconstructions for the global and Atlantic basin (Waelbroeck et al., 2009) and simulated by our GCM (LGM–HC experiment).

Data Cat				
Data Set				
	15° S–15° N	30° S–30° N	60° S–60° N	90° S–90° N
Observed: Global	$-1.7 \pm 1.0$	$-1.5 \pm 1.2$	$-1.9 \pm 1.7$	$-1.9 \pm 1.8$
Observed: Atlantic	$-2.9 \pm 1.3$	$-2.3 \pm 1.5$	$-2.6 \pm 2.0$	$-2.4 \pm 2.2$
Simulated: Global	-3.3	-3.2	-3.9	-3.2
Simulated: Atlantic	-3.6	-3.4	-3.9	-3.3

Model	Atmospheric CO <sub>2</sub> (ppm)	POC (Pg C y <sup>-1</sup> )	PIC (Pg C y <sup>-1</sup> )	Carbon (PgC)	$\Delta \text{ Carbon}^1$ (Pg C)	C <sub>organic</sub> (Pg C)	Oxygen (Pmol)	P <sub>organic</sub> (Pmol)	Alkalinity (Pmol Eq)	Salinity	Temperature (°C)
OHC-1	280.0	8.01	0.64	33974.0	0.0	1650.0	232.0	1.30	3003.0	34.505	3.992
OHC-2	185.0	8.01	0.64	32206.0	-1560	1650.0	232.0	1.30	3001.0	34.505	3.992
OLGM-1	185.0	4.48	0.36	33 370.0	-604.0	667.0	360.0	0.52	3054.0	35.099	1.649
OLGM-2	280.0	4.48	0.36	33 1 4 5.0	1127.0	667.0	360.0	0.52	3055.0	35.099	1.649
OLGM-3	185.0	4.48	0.00	33 407.0	-342.0	667.0	360.0	0.52	3055.0	35.099	1.649
OLGM-4	185.0	5.92	0.47	33 333.0	-416.0	772.0	348.0	0.61	3055.0	35.099	1.649
OLGM-5	185.0	3.25	0.26	33295.0	-454.0	721.0	354.0	0.57	3054.0	35.099	1.649
OLGM-6	185.0	4.82	0.00	34213.0	239.0	1004.0	323.0	0.79	3055.0	35.099	1.649

<sup>1</sup> Change in the ocean inventory of carbon relative to the OHC-1 simulation with atmospheric CO<sub>2</sub> at 280 ppm.



**Table 4.** (a) Simulated change in carbon content of the ocean relative to the HC (OHC-1). (1) assumes all three biological modifications that were postulated (see Table 1, experiments OLGM-3, 4 and 5) occurred to provide an upper bound estimate of ocean carbon storage. (b) Recent estimate of LGM–HC change in carbon (Ciais et al., 2011). (3) is the estimated additional ocean storage of carbon calculated as (2) - (1).

(a) Process	∆Carbon (PgC)
HC with atmospheric $CO_2$ of 185 ppm	-1560
LGM	-600
LGM with atmospheric CO <sub>2</sub> of 280 ppm	1130
(1) LGM with modified BGC (OLGM-6)	240
(b) LGM–HC change (Ciais et al., 2011)	
Atmosphere change	$-194 \pm 2$
Land change	$-330 \pm 400$
(2) Ocean change	$520 \pm 400$
(3) Additional ocean change	$280 \pm 400$





**Figure 1.** Annual sea surface temperature (SST) for **(a)** the LGM simulation (OLGM-1), **(b)** the observations from Levitus (2001), **(c)** the HC simulation (OHC-1) and **(d)** the simulated LGM– HC difference (OLGM-1 – OHC-1). In panels **(a–c)** the black line denotes the 18 °C isotherm.





**Figure 2.** Fraction of sea-ice cover for the annual monthly minimum (left) and annual monthly maximum (right) from the LGM (top) and the HC (bottom) simulations.













**Figure 4.** Zonal average salinity for **(a)** the LGM simulation (OLGM-1), **(b)** the observations from Levitus (2001), **(c)** the HC simulation (OHC-1) and **(d)** the simulated LGM–HC difference (OLGM-1 – OHC-1).





**Figure 5.** Zonal average dissolved oxygen  $(mmol m^{-3})$  for **(a)** the LGM simulation (OLGM-1), **(b)** the observations from Garcia et al. (2006a), **(c)** the HC simulation (OHC-1) and **(d)** the simulated LGM–HC difference (OLGM-1 – OHC-1).





**Figure 6.** Zonal average dissolved inorganic carbon  $(mmol m^{-3})$  for **(a)** the LGM simulation (OLGM-1), **(b)** the observations from Key et al. (2004), **(c)** the HC simulation (OHC-1) and **(d)** the simulated LGM–HC difference (OLGM-1 – OHC-1).





**Figure 7.** Zonal average alkalinity ( $\mu$ mol Eq m<sup>-3</sup>) for (a) the LGM simulation (OLGM-1), (b) the observations from Key et al. (2004), (c) the HC simulation (OHC-1) and (d) the simulated LGM-HC difference (OLGM-1 – OHC-1).



**Figure 8.** Zonal average phosphate concentration  $(mmol m^{-3})$  for **(a)** the LGM simulation (OLGM-1), **(b)** the observations from Garcia et al. (2006b), **(c)** the HC simulation (OHC-1) and **(d)** the simulated LGM–HC difference (OLGM-1 – OHC-1).







**Figure 9.** Annual export of particulate organic carbon from the upper 50 m (g C m<sup>-2</sup> y<sup>-1</sup>) for **(a)** the LGM simulation (OLGM-1), **(b)** OLGM-6 simulation, **(c)** the HC simulation (OHC-1) and **(d)** the simulated LGM–HC difference (OLGM-1 – OHC-1).





**Figure 10.** Annual mean phosphate concentration at 25 m (mmol  $m^{-3}$ ) for **(a)** the LGM simulation (OLGM-1), **(b)** the observations from Garcia et al. (2006a), **(c)** the HC simulation (OHC-1) and **(d)** the simulated LGM–HC difference (OLGM-1 – OHC-1).





**Figure 11.** The annual average surface ocean aragonite saturation state for **(a)** the LGM simulation (OLGM-1), **(b)** calculated from the observations of Key et al. (2004), **(c)** the HC simulation (OHC-1) and **(d)** from the OLGM-6 simulation that used the modified BGC parameterizations.















**Figure 14.** Change in annual export of particulate organic carbon from the upper 50 m  $(g C m^{-2} y^{-1})$  between the LGM and HC from the simulations with modified BGC formulations for **(a)** OLGM-6 – OHC-1, **(b)** OLGM-3 – OHC-1, **(c)** OLGM-4 – OHC-1 and **(d)** OLGM-5 – OHC-1.













**Figure 16.** Change in oxygen concentration  $(mmol m^{-3})$  at 500 m between the LGM and HC for (a) OLGM-6 – OHC-1, (b) OLGM-3 – OHC-1, (c) OLGM-4 – OHC-1 and (d) OLGM-5 – OHC-1.



**Figure 17.** Change in oxygen concentration  $(mmol m^{-3})$  at 3000 m between the LGM and HC for (a) OLGM-6 – OHC-1, (b) OLGM-3 – OHC-1, (c) OLGM-4 – OHC-1 and (d) OLGM-5 – OHC-1.

