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Glacial CO₂ cycle as a succession of key physical and biogeochemical processes

V. Brovkin^{1,*}, A. Ganopolski², D. Archer³, and G. Munhoven⁴

¹Max Planck Institute for Meteorology, Bundesstr. 53, 20146, Hamburg, Germany

²Potsdam Institute for Climate Impact Research, Telegraphenberg A31,
14473 Potsdam, Germany

³Department of the Geophysical Sciences, University of Chicago, 5734 S Ellis Ave,
Chicago, IL 60637, USA

⁴Institute of Astrophysics and Geophysics, University of Liège, 4000 Liège, Belgium
*also at: Potsdam Institute for Climate Impact Research, Potsdam, Germany

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Correspondence to: V. Brovkin (victor.brovkin@zmaw.de)

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Abstract

During glacial-interglacial cycles, atmospheric CO₂ concentration varied by about 100 ppmv in amplitude. While testing mechanisms that had led to the low glacial CO₂ level could be done in equilibrium model experiments, an ultimate goal is to explain CO₂ changes in transient simulations through the complete glacial-interglacial cycle. A computationally efficient Earth System model of intermediate complexity CLIMBER-2 is used to simulate global biogeochemistry over the last glacial cycle (126 kyr). The physical core of the model (atmosphere, ocean, land and ice sheets) is driven by orbital changes and reconstructed radiative forcing from greenhouse gases, ice, and aeolian dust. The carbon cycle model is able to reproduce the main features of the CO₂ changes: a 50 ppmv CO₂ drop during glacial inception, a minimum concentration at the last glacial maximum by 80 ppmv lower than the Holocene value, and an abrupt 60 ppmv CO₂ rise during the deglaciation. The model deep ocean $\delta^{13}\text{C}$ also resembles reconstructions from deep-sea cores. The main drivers of atmospheric CO₂ evolve with time: changes in sea surface temperatures and in the volume of bottom water of southern origin controls atmospheric CO₂ during the glacial inception and deglaciation, while changes in carbonate chemistry and marine biology are dominant during the first and second parts of the glacial cycle, respectively. These feedback mechanisms could also significantly impact the ultimate climate response to the anthropogenic perturbation.

1 Introduction

Ice core records of atmospheric CO₂ concentration through the last 800,000 years show the carbon cycle amplifying the climate forcing from variations in Earth's orbit (Augustin et al., 2004). This positive climate-carbon cycle feedback could weaken or even possibly reverse present-day fossil fuel CO₂ uptake by the natural carbon cycle. Despite much effort over the last decades, process-based explanation of the carbon cycle feedbacks responsible for the glacial-interglacial CO₂ cycles remains elusive.

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Mechanisms proposed to explain atmospheric glacial CO₂ cycles include changes in sea surface temperatures (SSTs), nutrients, circulation, and carbonate chemistry (Archer et al., 2000; Sigman and Boyle, 2000; Toggweiler et al., 2006; Kohfeld and Ridgwell, 2009). The main focus has been to explain the low atmospheric CO₂ concentration of about 190 ppmv at the last glacial maximum (LGM) about 21 000 years ago. Most of these studies were built on the assumption that the carbon cycle was in equilibrium at this time (e.g., Hain et al., 2010; Kurahashi-Nakamura et al., 2010). However, the carbon cycle is most likely never in steady state during a glacial cycle because of the long time scales of many of the biogeochemical processes. Our goal is to explain the CO₂ concentration not only for the LGM snapshot, but through the whole glacial cycle focusing on a possible sequence of mechanisms that drive CO₂ changes (Kohler and Fischer, 2006; Peacock et al., 2006; Watson et al., 2000). A challenge is to explore these mechanisms using the Earth system models that include 2- or 3-D models of oceanic circulation, a key driver of oceanic biogeochemistry.

2 Methods

2.1 The model

The CLIMBER-2 model includes six components of the Earth system: atmosphere, ocean, sea ice, land surface, terrestrial vegetation and ice sheets. The first five components are represented by coarse-resolution modules of intermediate complexity (Brovkin et al., 2002; Petoukhov et al., 2000). The ice-sheet component is represented by the three-dimensional polythermal ice-sheet model SICOPOLIS (Greve, 1997). The ice sheet model is applied only to the Northern Hemisphere (NH). The carbon cycle model includes land carbon, oceanic biogeochemistry, a model for marine biota, and a sediment model (Archer, 1996; Brovkin et al., 2007). Weathering rates scale to runoff from the land surface (Munhoven, 2002), the coral reef growth depends on the sea level rise (Kleypas, 1997), the scale of nutrients utilization in sub-Antarctic ocean

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is proportional to the dust deposition derived from Antarctic ice core (Augustin et al., 2004), and volcanic outgassing was assumed to have remained constant through the whole glacial cycle. The carbon cycle components were run off-line, i.e. simulated changes in atmospheric CO₂ were not fed back into the radiative calculation, which evolved as a function of the history of equivalent CO₂ derived from ice core data.

2.2 Experimental setup of the physical model

The experimental set-up followed a recent simulation of the last glacial cycle without carbon fluxes (Ganopolski et al., 2010). We prescribed temporal variations in the orbital parameters (Berger, 1978) and the equivalent CO₂ concentration which accounts for changes in three major greenhouse gases (GHGs – CO₂, N₂O and CH₄) derived from the Antarctic ice cores (Barnola et al., 1987; Augustin et al., 2004; Monnin et al., 2004), and aeolian dust deposition (Mahowald et al., 1999) (Fig. 1). As initial conditions for the physical climate system, we used equilibrium climate and ice sheets simulated for the present-day orbital configuration and GHG concentrations. The model was run from 126 kyr BP for 136 000 years, i.e. for 10 000 years into the future. This was done to estimate the equilibration time of the global carbon cycle after deglaciation. For the future, GHG concentrations were kept constant at the preindustrial level, i.e. the human influence on climate was not considered. Under the influence of the orbital forcing only, CLIMBER-2 simulates a slow growth of the ice sheets in the Northern Hemisphere which is explained by favorable orbital configuration but small eccentricity and, therefore, weak orbital forcing. Simulated changes in climate and ice volume over the past 126 000 years agree favorably with paleoclimate reconstructions (Ganopolski et al., 2010).

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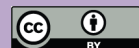
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2.3 Experimental setup of biogeochemistry

2.3.1 Ocean carbon cycle: the Southern Hemisphere outgassing

The physical ocean component of the model is essentially the same as used in the glacial equilibrium simulation (Brovkin et al., 2007). The only important difference is that the background vertical diffusivity in the ice free Southern Ocean south of 50° S was enhanced by an order of magnitude, i.e. to ca. $10^{-3} \text{ m}^2 \text{ s}^{-1}$ (under the sea ice, the standard values of $10^{-4} \text{ m}^2 \text{ s}^{-1}$ was retained). This was done to obtain CO_2 outgassing in the Southern Ocean (SO) consistent with recent empirical estimates of about 0.4 Gt yr^{-1} (Gruber et al., 2009). The standard version of CLIMBER-2, similar to many other even much more sophisticated ocean models (Fletcher et al., 2007) considerably underestimates the magnitude of CO_2 outgassing in the SO. We found that enhanced outgassing in the SO considerably increases atmospheric $p\text{CO}_2$ sensitivity to changes in sea ice cover in the SO. For example, in the standard model version (without enhanced vertical diffusivity in the SO) under present-day climate conditions, complete insulation of the SO (south of 50° S) from the atmosphere leads to changes in atmospheric $p\text{CO}_2$ by only 4 ppm (Archer et al., 2003) while in the version with enhanced outgassing, the SO insulation leads to a 20 ppm decrease in atmospheric CO_2 . The model version with the enhanced southern outgassing simulates glacial/interglacial CO_2 variations of ca. 10 ppm higher than the standard model version.

2.3.2 Inventories of ^{12}C and ^{13}C

Global inventories of ^{12}C and ^{13}C in the model experiments are conservative. Atmospheric CO_2 concentration was calculated from the total carbon inventory equation:

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$$\begin{aligned}
 & \left({}^iC^{\text{atm}}(t) - {}^iC_0^{\text{atm}} \right) + \left({}^iC^{\text{oc}}(t) - {}^iC_0^{\text{oc}} \right) + \left({}^iC^{\text{land}}(t) - {}^iC_0^{\text{land}} \right) \quad (1) \\
 & = \int_0^t \left({}^iF^{\text{volc}}(t) - {}^iF^{\text{atm-cons}}(t) \right) + \left({}^iF^{\text{weath}}(t) \right) - \left({}^iF^{\text{burial}}(t) \right) dt
 \end{aligned}$$

where the index i denotes the carbon isotope (^{12}C or ^{13}C), iC stands for carbon isotope compartments (GtC) and iF is for carbon isotope fluxes (GtC yr^{-1}), indices atm, oc, land are for atmospheric, ocean, and land compartments, t is the model year starting from 0 at 126 000 yr BP, and the index 0 for carbon compartments is for the initial storages of carbon at the beginning of the simulations. ${}^iF^{\text{volc}}$ is for volcanic outgassing of CO_2 fixed to a value $0.066 \text{ GtC yr}^{-1}$ with ^{13}C content of 6‰ (relative to the Pee Dee Belemnite (PDB) standard). ${}^iF^{\text{atm-cons}}$ is the atmospheric consumption of CO_2 due to terrestrial weathering of both carbonate and silicate rocks, ${}^iF^{\text{weath}}$ is the bicarbonate flux from the land to the ocean arising from continental weathering. The consumption of atmospheric CO_2 and the production of riverine HCO_3^- by continental weathering processes are calculated as a function of geographically distributed runoff (interactively simulated by the land surface module) and lithology (Amiotte-Suchet et al., 2003). ${}^iF^{\text{burial}}$ is an interactively simulated CaCO_3 burial on shelves (coral reefs) (Kleypas, 1997) and in the deep sea (Archer, 1991). Ocean carbon storage ${}^iC^{\text{oc}}$ is for dissolved inorganic and organic carbon components in the ocean including marine biota biomass. Land carbon compartment ${}^iC^{\text{land}}$ is for vegetation biomass and soil carbon storages. In the land carbon cycle, we assumed that the decomposition of soil organic matter is slowed if annual mean temperatures fall below -10°C .

In the Eq. (1), the land surface area accounted in the land carbon storage ${}^iC^{\text{land}}$, weathering ${}^iF^{\text{weath}}$, and atmospheric CO_2 consumption ${}^iF^{\text{atm-cons}}$ was linearly scaled between pre-industrial and LGM land surface areas proportionally to the sea level. The ocean volume was also adjusted in accordance with the sea level. The total ocean alkalinity is calculated as a balance of alkalinity fluxes due to terrestrial weathering and burial of CaCO_3 on shelves and in the deep ocean.

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Atmospheric CO₂ calculated from Eq. (1) (using a conversion factor of 0.47 ppm/GtC) was analysed in a diagnostic mode and not used as a radiative forcing for the atmospheric model component.

2.3.3 Initial conditions of the carbon cycle

5 Initial conditions of the land and oceanic biogeochemistry were set up following a procedure used recently for initializing interglacial carbon cycle (Kleinen et al., 2010). Firstly, the model was run with equilibrium conditions of 126 000 BP with CO₂ kept at the level of 280 ppmv. Ocean alkalinity was increased to get a carbonate sedimentation flux of 16 Tmol/a in the deep ocean and 2 Tmol/a on the shelves in order to simulate the
10 maximum in CaCO₃ preservation in the deep sea before the onset of the interglacial. Atmospheric δ¹³CO₂ was initialized to -6.7‰. The model was run with prescribed CO₂ for 5000 years. In the transient simulations through the last glacial cycle, the carbonate accumulation on the shelves and in the deep sea was interactive with a shift towards carbonate accumulation in shallow waters during interglacials. This setup of initial
15 conditions ensures that the ocean biogeochemistry is in equilibrium with the climate at the onset of interglacial while it is in transition from the glacial to interglacial state hereafter.

3 Results and discussion

3.1 Transient experiments

20 The physical part of the CLIMBER-2 model has been used to simulate the last glacial cycle, forced by changes in orbital forcing and greenhouse gases concentrations (Ganopolski et al., 2010). The model reproduces the temporal and spatial dynamics of the major Northern Hemisphere (NH) ice sheets, including rapid glacial inception started at 120 to 118 kyr BP (thousand years before present) (Fig. 2a).

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Here, we focus on the response of various parts of the carbon cycle to the changes in the physical system. Simulation P includes physical mechanisms effects (changes in SST, circulation, and sea level), PC adds changes in carbonate chemistry (carbonate compensation, shift of carbonate sedimentation from shallow waters to deep sea and vice versa, and weathering), PCB complements with changes in nutrient utilization by marine ecosystems (iron fertilization due to aeolian dust deposition), and PCBL adds land carbon dynamics. All simulations start from the same non-equilibrium conditions in oceanic carbonate chemistry at 126 kyr BP and continue to 10 000 years after the present day without considering anthropogenic forcings. The atmospheric CO₂ response of the full PCBL simulation is shown in black in Fig. 2c, and the responses are broken down into component mechanisms in Fig. 2d.

During the last interglacial, CO₂ dynamics in the full experiment, PCBL, increases from an initial value of 280 ppmv at 125 kyr BP to 285 ppmv at 118 kyr BP (Fig. 2c). This is driven by the response of the terrestrial biosphere to climate trends during the late interglacial, as drying of the northern subtropical regions and the southward retreat of boreal forests decrease land carbon storage by about 200 GtC (difference between PCB and PCBL). In the P simulation, increased CO₂ solubility in seawater due to lowered SSTs leads to an atmospheric CO₂ decrease during the late interglacial, while carbonate chemistry processes including accumulation of shallow-water carbonates force CO₂ in the opposite direction leading to CO₂ stagnation till 117 kyr BP in the PC and PCB simulations (Fig. 2d).

The glacial inception results in an abrupt CO₂ drop by ca. 50 ppmv to the level of 230–235 ppmv (Fig. 2c) by 110 kyr BP which is in a good agreement with the Vostok ice core data (Barnola et al., 1987). The main mechanism here is physical (simulation P), as reduction in SSTs and substitution of North Atlantic deep water by colder and saltier waters of Antarctic origin lead to ca. 30 ppmv drop (Brovkin et al., 2007). The essence of the later mechanism, called the “standing volume effect” (Skinner, 2009), is that the dissolved inorganic carbon (DIC) content of southern-source water is higher than that of North Atlantic deep water, leading to increased DIC storage in the deep ocean. The

model simulates a decrease in $\delta^{13}\text{C}$ in deep tropical Atlantic by 0.2‰ during glacial inception, qualitatively in agreement with the benthic $\delta^{13}\text{C}$ data (Lisiecki et al., 2008; Oliver et al., 2010) (Fig. 2e).

During the rest of the glacial period, physical mechanisms only have a substantial impact on atmospheric CO_2 during Heinrich and Dansgaard-Oeschger type events, when Atlantic meridional overturning circulation (AMOC) is greatly perturbed (Fig. 2b). During these events, the model produces spikes of CO_2 of about 10 ppmv (Fig. 2d) due to outgassing of oceanic CO_2 , mostly via the Southern ocean (Anderson et al., 2009). At the same time, the deep ocean is enriched in negative $\delta^{13}\text{C}$ due to increased water mass ages (Fig. 2e and f). Millennial-scale variability in atmospheric CO_2 during abrupt changes in AMOC is visible in the high resolution Byrd ice core data (Ahn and Brook, 2008). The specific timings of the abrupt climate events are different from observed as they depend on freshwater discharge into North Atlantic, which are random output of instability of the Laurentide ice sheet within the model.

By 110 kyr BP, carbonate compensation amplifies the physics-driven changes in atmospheric CO_2 by an additional 10 ppmv (Fig. 2d, the difference between P and PC). Between 110 and 100 kyr BP, atmospheric CO_2 rises by 20 ppmv as a consequence of partial deglaciation due to increasing summer insolation in boreal region (Fig. 2a). This CO_2 response to the precessional cycle is not evident in the ice core data (Barnola et al., 1987), although the coarse temporal resolution of the only available Vostok record might be not enough for resolving the cycle. Between 90 and 70 kyr BP, the model generates atmospheric CO_2 at the level of 230–240 ppmv, in line with the Byrd record (Ahn and Brook, 2008). After 70 kyr BP, the effect of fertilization of marine biological production in sub-Antarctic Atlantic ocean by aeolian dust leads to a CO_2 drop to a level of 200–210 ppmv (Fig. 2d). The timing of the dust effect on the carbon cycle is in line with box model simulations (Hain et al., 2010; Martinez-Garcia et al., 2009; Ridgwell and Watson, 2002; Watson et al., 2000) although the magnitude of the impact of iron fertilization differs among models (e.g., Parekh et al., 2008). Enhanced marine productivity leads to an additional CO_2 drop with minima of $p\text{CO}_2$ at 63 and 20 kyr BP in

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simulations PCB and PCBL (Fig. 2d). These minima are also pronounced in the deep ocean $\delta^{13}\text{C}$ concentration, both in the model and in the data (Fig. 2e and f). During the remaining period of 60–20 kyr BP, modelled atmospheric CO_2 stays in the range of 200–220 ppmv with exceptional spike at 54 kyr BP when CO_2 quickly rises to 230 ppmv during a Heinrich-type event and relaxes slowly to the level of 220 ppmv thereafter.

The model atmospheric CO_2 concentration during the Last Glacial Maximum was about 200–210 ppmv, ca. 10 ppmv higher than in the EPICA Dome C (EDC) record (Monnin et al., 2004). In the equilibrium model simulation (Brovkin et al., 2007), the CO_2 level was higher (about 220 ppmv). In equilibrium experiments, the total ocean alkalinity is prescribed by initial conditions, whereas in transient simulation, it is calculated interactively as a cumulative balance between weathering and sedimentation fluxes, which vary in time.

Deglaciation in the model starts by ca. 20 kyr BP in line with the sea level data (Fig. 2a). The increased freshwater flux to the North Atlantic shuts down the AMOC (Fig. 2b). In response, atmospheric CO_2 rises by about 10 ppmv (Fig. 2c). Between 18 kyr BP and 11 kyr BP, CO_2 rises from about 200 to 235 ppmv, which is less than the steep increase in the EDC data by about 50 ppmv. A substantial part of the CO_2 increase (ca. 20 ppmv) is due to reduced marine productivity in the sub-Antarctic region (Fig. 2d, PCB). After the AMOC recovers, the ocean circulation returns to an interglacial condition with enhanced North Atlantic Deep Water formation, interrupted by a short shutdown at 11 kyr BP.

In the PCBL simulation, accumulation of CaCO_3 in the deep tropical Pacific decreases during both the last interglacial (126–115 kyr BP) and the Holocene in response to enhanced shallow-water CaCO_3 burial on shelves (Fig. 3). During the glacial period, CaCO_3 burial increased, with a peak of preservation during deglaciation at about 15 kyr BP, in line with data from the deep South Atlantic (Hodell et al., 2001; Anderson et al., 2008). A key diagnostic of ocean carbonate chemistry, the concentration of carbonate ion, $[\text{CO}_3^{2-}]$, changes in the deep ocean through the glacial

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cycle. At LGM, $[\text{CO}_3^{2-}]$ in the deep tropical Pacific increases by 10–15 $\mu\text{mol kg}^{-1}$, and peaks during deglaciation as ocean CO_2 is released to the atmosphere (Yu et al., 2010) (Fig. 4). In the deep tropical Atlantic, $[\text{CO}_3^{2-}]$ and CaCO_3 burial decrease during interglacials, but carbonate preservation peaks during LGM and the early Holocene. Contrary to the Pacific, $[\text{CO}_3^{2-}]$ in Atlantic decreases during deglaciation as a shutdown of the Atlantic meridional overturning circulation brings in Antarctic bottom waters characterized by lower $[\text{CO}_3^{2-}]$. The resumption of circulation leads to an increase in $[\text{CO}_3^{2-}]$ and enhanced preservation of CaCO_3 . The later maximum in CaCO_3 preservation in the Atlantic is consistent with sedimentary data (Broecker et al., 2001; Hodell et al., 2001), as is the Atlantic maximum in $[\text{CO}_3^{2-}]$ at the LGM (Rickaby et al., 2010).

The main forcing of atmospheric CO_2 after 10 kyr BP is a reduction of ocean alkalinity resulting from coral reef regrowth (Kleypas, 1997) and its impact on the CaCO_3 cycle (Ridgwell et al., 2003). The terrestrial biosphere ameliorates the CO_2 rise by taking about 500 GtC in between 20 and 5 kyr BP. By 0 kyr BP, the carbonate chemistry is still not fully equilibrated to interglacial conditions, as atmospheric CO_2 , now at 274 ppmv, continues to rise. Extending the model runs by 10 kyr into the future (without anthropogenic forcing) allows the model to find an interglacial equilibrium of 279 ppm at 4 kyr in the future. The main difference between this state and the previous interglacial state at 126 kyr is that the terrestrial biosphere carbon storage is about 200 GtC smaller, due to more favourable conditions of the past interglacial relative to the Holocene. Recent evidence (Zimov et al., 2009) shows that several hundred GtC might have been stored in the high latitude regions in permafrost soils during the glacial period. Therefore, although model experiments suggest that the land biosphere had 500–800 Gt less carbon at the LGM than during the Holocene (Kaplan et al., 2002), the simulated reduction in glacial land carbon storage could possibly be overestimated since these models neglect changes in permafrost carbon storage.

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3.2 “Standing volume” effect on the atmospheric CO₂ concentration

For a better understanding of the physical mechanisms responsible for the glacial CO₂ drawdown, it is useful to compare the standard experiment P, which only includes physical processes, with an even more schematic experiment P_NOI. In that experiment the ice sheet component was switched off and therefore neither climate changes due to growth and decay of the ice sheets, nor variations in the sea level and ocean volume affect the carbon cycle. Therefore, all climate changes that affect the carbon cycle in this experiment are only driven by orbital forcing and radiative forcing of the changing CO₂ concentration. Note that the direct effect of orbital forcing on climate and CO₂ is rather small. In all other respects, the experiment P_NOI is equivalent to the experiment P. Simulated CO₂ variations for the P and P_NOI simulations are shown in Fig. 5 by black and purple lines, respectively. Simulated CO₂ in the P_NOI experiment closely resembles prescribed equivalent CO₂ concentration but with about a three times smaller amplitude of variation (Fig. 5). The maximum glacial/interglacial CO₂ difference in this experiment is about 40 ppm. Since variations in the global mean SST are about 2 °C (similar changes are simulated in the global surface air temperature and the deep ocean temperatures), this implies a CO₂ sensitivity to temperature changes of about 20 ppm/°C which is too high for the solubility effect alone (Archer et al., 2004). In fact, the solubility effect explains only ca. 10 ppm, while the rest results from changes in the redistribution of major water masses and their ventilation.

The first effect was already observed in previous equilibrium glacial simulations (Brovkin et al., 2007) and was later named as “standing volume effect” (Skinner, 2009). The essence of this effect is that under glacial climate conditions, the volume of the water masses of the North Atlantic origin in the World Ocean decreases and the volume of the water masses of the southern origin increases. Since the DIC concentration in the water masses originating in the SO is higher than in the NADW, this leads to an increase of carbon storage in the deep ocean. To illustrate this effect we have added a dye tracer to the ocean which allows us to compute relative contributions of the water

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masses of each origin at each location. Figure 6 shows that in the interglacial climate state most of the deep Atlantic is filled with the water masses of the North Atlantic origin. Under glacial climate conditions, the upper cell of the AMOC becomes shallower and more water masses of the southern origin penetrate into the deep Atlantic.

5 As a result, the fraction of the northern water masses in the deep Atlantic decreases from almost 100 % to 50 %. Smaller but not negligible changes were also found in the Pacific and Indian oceans (not shown). Note that increase of the fraction of the water masses of southern origin is related not to the changes in the rate of Antarctic bottom water (AABW) formation but to a relative increase of the density of AABW compared to NADW. This, in turn, is primarily related to the enhanced sea ice formation in the Southern Ocean, stronger brine rejection and an increase of the AABW salinity (Brovkin et al., 2007). Increased salinity of AABW agrees with empirical reconstructions (Adkins et al., 2002). Since glacial/interglacial climate change in the Southern Ocean is primarily explained by changes of GHG concentrations, the increase in the sea ice area in the P_NOI experiment is similar to that in the P experiment with the interactive ice sheets. As a result, the fraction of northern water masses in the World Ocean (below 1.5 km) in the P_NOI experiment decreases at the LGM by ca. 10 % compared to the interglacial state. During the first 100 ka, the changes in the fraction of the North Atlantic water masses are similar in the P and in the P_NOI experiments, but around the LGM, the fraction of the North Atlantic water masses falls drastically in the P experiment, which is related to further weakening and frequent shutdowns of the AMOC at this period. This further reduction in the fraction of North Atlantic water masses, however, does not result in the further CO₂ drop. In fact, the CO₂ concentration in the P_NOI experiment is lower at this period of time than in the P simulation. This can be explained by the fact that the weakening of the AMOC in the P experiment results in a reduction of the CO₂ uptake in the North Atlantic which in turn leads to a decrease of DIC in the intermediate water masses and compensates for the increase in the fraction of the water masses of southern origin.

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Recently, Bouttes et al. (2011a) used another version of CLIMBER-2 model to analyse equilibrium sensitivity of the glacial carbon cycle to brine-induced stratification, stratification-dependant diffusion, and iron fertilization. They demonstrated that with certain parameterizations of these processes, the model is able to reproduce the atmospheric CO₂ level and the ocean carbon isotopic composition observed for the LGM. The results of Bouttes et al. (2011a) can possibly be interpreted in terms of the “standing volume effect”. However, such an analysis would require additional experiments which go beyond the purpose of this study.

3.3 Millennial-scale variability in atmospheric CO₂

Apart from climate changes on the orbital time scales, the CLIMBER-2 model simulates pronounced millennial scale variability associated with changes in the AMOC strength caused by the changing freshwater flux into the northern North Atlantic. The largest fluctuations of the freshwater flux are associated with the internal instability of the Laurentide ice sheet and the development of fast ice streams over the Hudson Strait (Calov et al., 2002). These events of enhanced ice flux resemble real Heinrich events and result in additional freshwater fluxes of about 0.1 Sv during several hundreds to one thousand years. Since Heinrich-like events simulated in the model result from internal instability of the ice sheet component, their timing is random and does not coincide with the timing of real Heinrich events. During these intervals of enhanced freshwater flux into the North Atlantic, the AMOC weakens considerably. The NADW formation area retreats southward and the AMOC cell shoals (Ganopolski and Rahmstorf, 2001). These changes of the AMOC are closely associated with millennial scale variations in the atmospheric CO₂ concentration (Fig. 2c–d). During the phases of weak AMOC, the CO₂ concentration rises rapidly by 10–20 ppm and equally rapidly drops after the resumption of strong AMOC. The temporal dynamics and the magnitude of the simulated millennial CO₂ variations resemble CO₂ variations in the Byrd ice core record during Heinrich stadials (Fig. 2c).

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The strong positive excursions in the CO₂ concentration during Heinrich-like events can already be seen in the P experiment. The magnitude of these variations is essentially the same in the experiment with and without terrestrial biosphere. Therefore, the rise of CO₂ concentration during the weak AMOC in our model can be primarily attributed to the ocean component of the carbon cycle. The small contribution of the terrestrial biosphere to the CO₂ change can be explained by the absence of terrestrial vegetation in the areas primarily affected by the reorganization of the AMOC under the glacial climate conditions. The detailed analysis of the mechanisms behind the millennial scale variability of the atmospheric CO₂ concentration reveals considerable complexity and a number of counteracting mechanisms involved.

During periods of weakened AMOC, the carbon uptake in the northern North Atlantic is significantly reduced and DIC in the core of NADW decreases. However, a reduction of carbon content in the upper part of the Atlantic is compensated by an increase of DIC in the deep Atlantic due to the substitution of the NADW by southern water masses which have relatively high DIC concentration. As a result, the Atlantic Ocean, although characterized by the largest changes in the atmosphere-ocean carbon fluxes, does not contribute much to the CO₂ rise during stadials. In the Southern Ocean neither the surface fluxes nor the total carbon inventory are appreciably affected by changes in the AMOC in spite of significant variations in the surface temperature and sea ice cover caused by the reduction of the northward ocean heat flux (i.e., the “bipolar seesaw” mechanism). This is because the increase of the oceanic outgassing in the area where the southern sea ice cover is retreated polarwards is counteracted by an increase in the strength of the biological pump.

There are little changes in the surface ocean-atmosphere carbon fluxes over the Indo-Pacific but, surprisingly, carbon inventory shows that the rise of the atmospheric CO₂ during periods of weakened AMOC is explained by the loss of carbon from the Indo-Pacific. The decrease in DIC occurs here primarily at intermediate depths between 1 and 3 km and is explained by a weakening of the reverse cell of the Indo-Pacific overturning circulation by about 2 Sv during periods of reduced AMOC. Therefore, the

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physical mechanisms of the millennial and of the orbital scale variability of the atmospheric CO₂ in our model are different.

Previous studies of the carbon cycle response to a weakening or a complete shut-down of the AMOC produced controversial results. While Marchal et al. (1999), Schmitner and Galbraith (2008), and Bouttes et al. (2011b) found that there was ocean releasing CO₂ during weakenings of the AMOC, Menviel et al. (2008) and Bozbiyik et al. (2011) reported opposite results. We conclude that the complexity of the response of atmospheric pCO₂ to such changes in the AMOC and a number of counteracting processes taken into account in our model may explain why this response is so strongly model-dependent.

4 Conclusions

According to our results, physical mechanisms – reduction in SSTs and changes in standing volume in response to the expansion of ice-sheets especially in the Northern Hemisphere and the resulting lower sea level – lead to the initial drop of CO₂ during glacial inception. This drop, amplified by the CaCO₃ cycle, is followed by increased buildup of ocean alkalinity during the rest of the glacial cycle, as exposed tropical shelves serve as a source of CaCO₃ to the ocean, and CaCO₃ burial is shifted from shallow waters to the deep sea. Increased nutrient utilization in the sub-Antarctic plays the dominant role in further CO₂ decline. The role of the land carbon is less certain as recently published data suggest globally smaller changes in the land carbon storage between glacial and interglacial times than previously thought.

In the second part of the glacial cycle, the model simulates millennial scale variations in the atmospheric CO₂ concentration that appear to be closely associated with AMOC changes. In response to AMOC weakenings, the CO₂ concentration rapidly rises, similarly to CO₂ variations observed during Heinrich stadials (Ahn and Brook, 2008). During these variations, Indo-Pacific Ocean serves as a source of CO₂, while land carbon storage increases in response to the elevated CO₂ concentration. Distributions of

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carbon sinks and sources in response to glacial AMOC changes differ among models due to counteracting effects of physical and biogeochemical mechanisms. A detailed analysis of carbon cycle response to AMOC changes is certainly an interesting topic for future model intercomparison.

5 During the deglaciation and the Holocene, the model simulates an increase in the atmospheric CO₂ concentration that is of the right amplitude but less abrupt than the rapid CO₂ rise recorded in the ice cores (Monnin et al., 2004). This mismatch could be due to limitations in the representation of ocean biogeochemistry, especially carbonate chemistry, in our zonally-averaged model or to abrupt processes not considered in our
10 land carbon model, such as rapid thawing of frozen organic matter stored in permafrost soils. Implementation of permafrost carbon could possibly also improve the CO₂ dynamics during glacial inception. Constraining the amplitude of such a permafrost carbon feedback from paleo simulations is essential for projecting the response of the carbon cycle at high latitudes to anthropogenic climate change over the next centuries
15 (Schaefer et al., 2011).

In this study, the carbon cycle was simulated in a diagnostic mode, as there was no feedback from simulated atmospheric CO₂ to the physical climate model. This partial decoupling between climate and carbon cycle is necessary to evaluate a role of different mechanisms in the glacial CO₂ changes within the model framework. These
20 numerical experiments serve as a basis for further simulations of glacial cycles driven solely by changes in orbital forcing.

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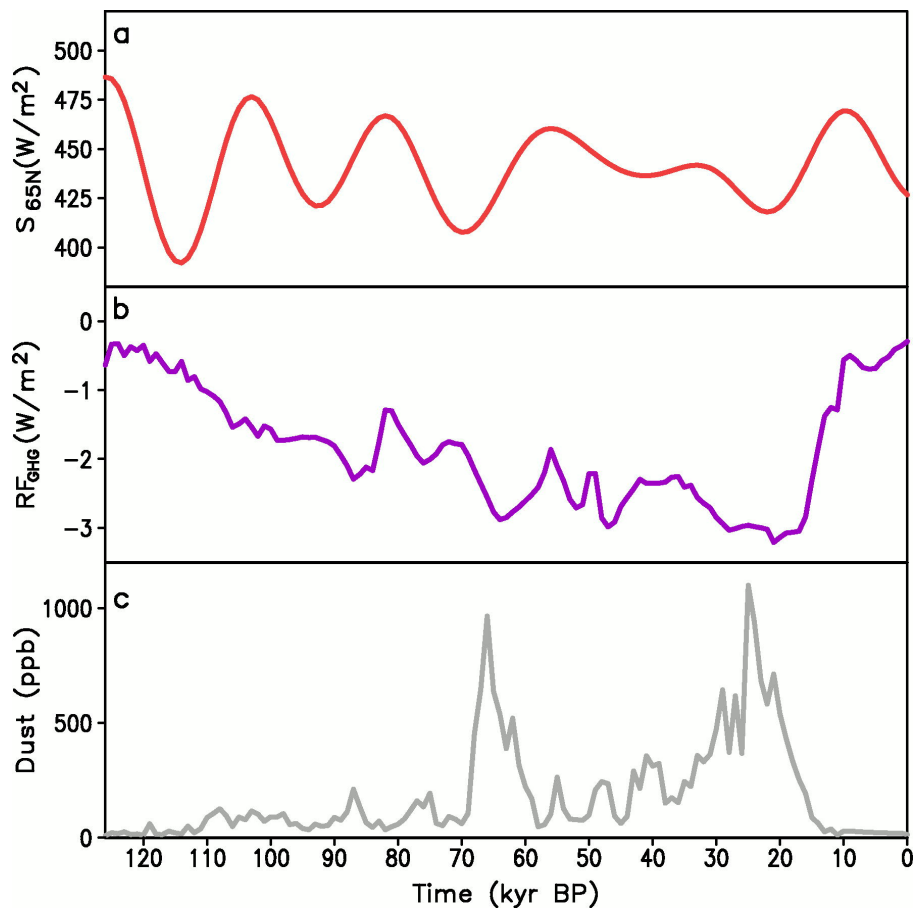


Fig. 1. Prescribed forcings of the simulation experiments of the last glacial cycle. **(a)** Orbital forcing (illustrated by the June insolation at 65° N), **(b)** radiative forcing of GHGs, **(c)** concentration of dust in the EPICA ice core.

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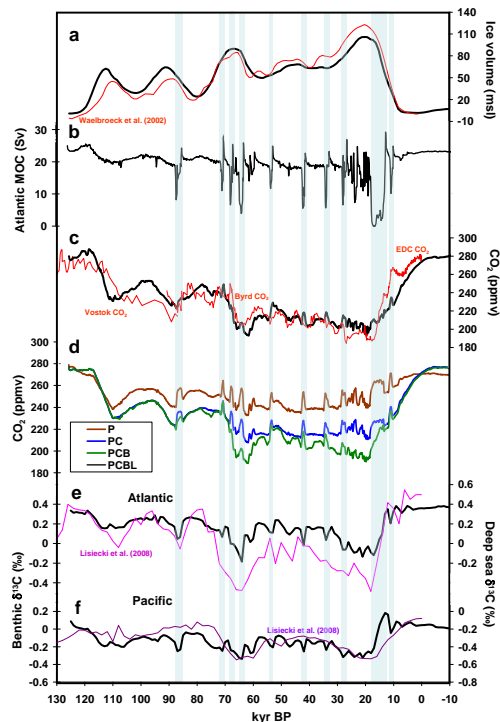


Fig. 2. Comparison of model results with geological records. **(a)** Changes in the simulated ice sheet volume (thick black line) and reconstructed sea level change (Waelbroeck et al., 2002) (thin red line), m of sea level change relative to 126 kyr BP. **(b)** Simulated Atlantic meridional overturning circulation, Sv. **(c)** Atmospheric CO₂ concentration, simulation PCBL (thin black line), and Antarctic ice core reconstructions (Ahn and Brook, 2008; Barnola et al., 1987; Monnin et al., 2004) (thin red lines), ppmv. **(d)** Atmospheric CO₂ concentration for simulations P, PC, and PCB (brown, blue, and green lines, respectively), ppmv. **(e)** Deep sea $\delta^{13}\text{C}$ for simulation PCBL (thick black line) and geological stack record for the Atlantic Ocean (Lisiecki et al., 2008) (thin pink line), ‰. **(f)** The same as **(e)** but for Pacific Ocean.

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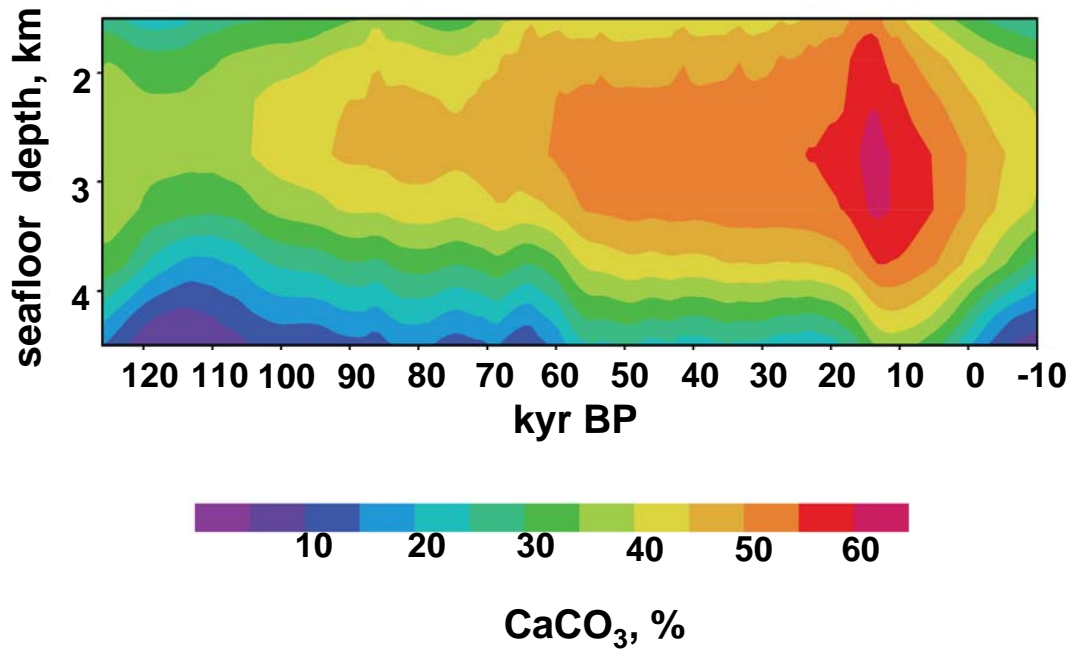


Fig. 3. Dynamics of the vertical distribution of simulated CaCO_3 fraction (%) in top layer of seafloor sediments averaged over 30°S – 30°N , Pacific Ocean.

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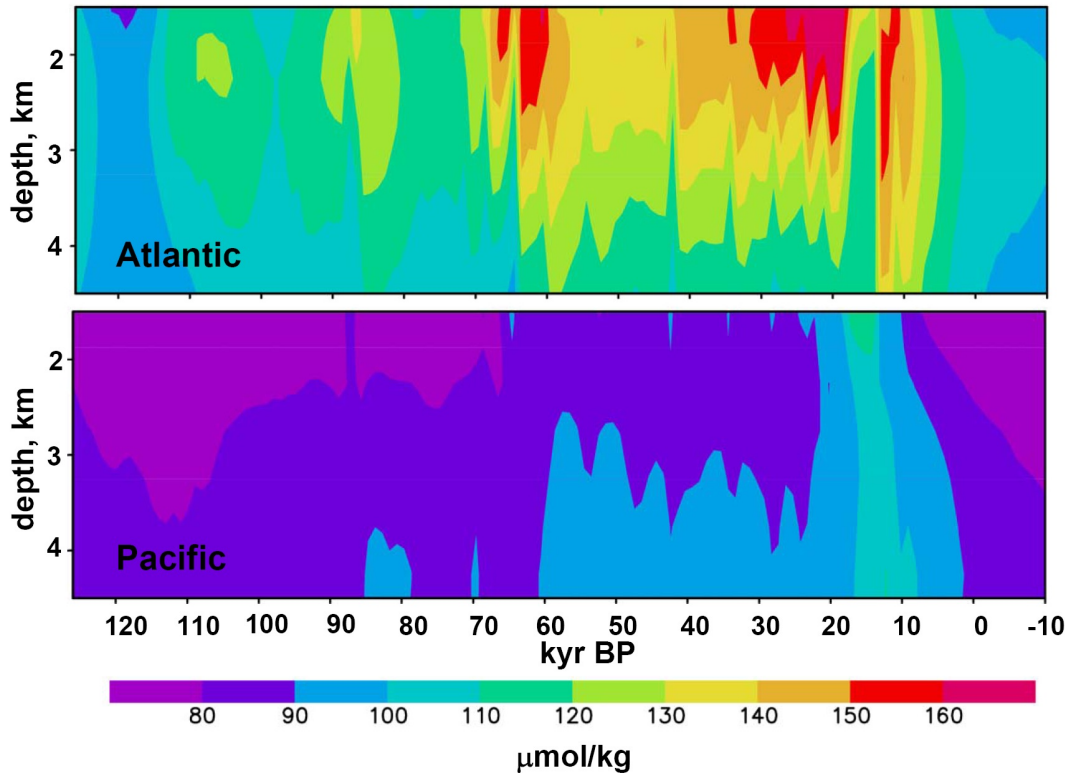


Fig. 4. Dynamics of the vertical distribution of simulated $[\text{CO}_3^{2-}]$ concentration in the deep ocean, $\mu\text{mol kg}^{-1}$, averaged over 30°S – 30°N . Top, Atlantic Ocean; bottom, Pacific Ocean.

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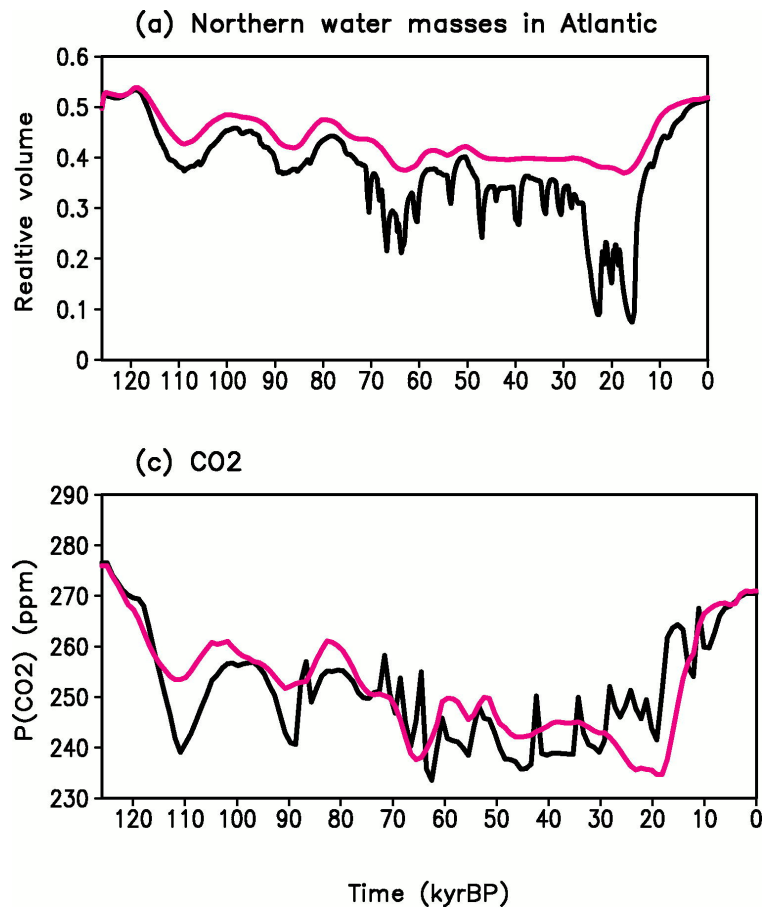


Fig. 5. Relative volume of the water masses of the northern origin in the Atlantic ocean **(a)** and simulated atmospheric CO₂ concentration **(b)**. Black line – experiment P, purple line – experiment P_NOI.

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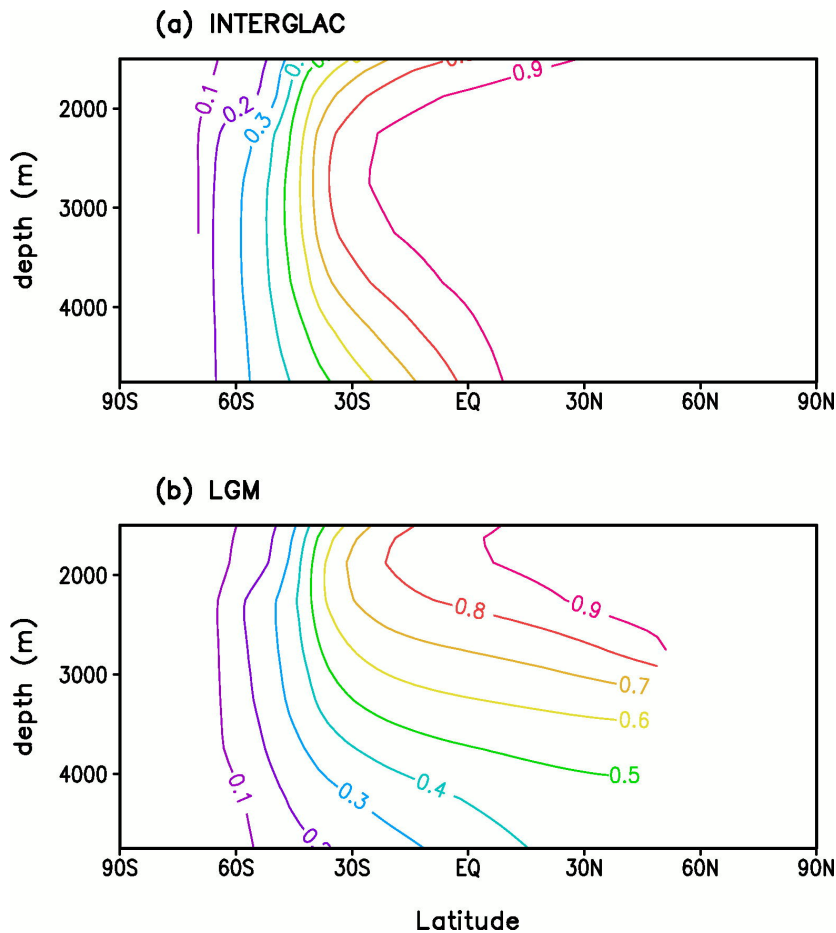


Fig. 6. “Dye” tracer concentration in the Atlantic for interglacial **(a)** and LGM **(b)** states. Northern water masses have a “dye” concentration of 1 and the southern water masses of 0.