Clim. Past Discuss., 4, 309–333, 2008
www.clim-past-discuss.net/4/309/2008/

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Influence of the Atlantic thermohaline circulation on neodymium isotopic composition at the Last Glacial Maximum – a modelling sensitivity study

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Received: 6 February 2008 – Accepted: 14 February 2008 – Published: 18 March 2008

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Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

The oceanic neodymium isotopic composition (hereafter expressed as ε_{Nd}) is modeled for the Last Glacial Maximum (LGM) using the coarse resolution Ocean Global Circulation Model NEMO–ORCA2°. This study focuses on the impact of changes in the overturning cell and circulation patterns between LGM and Holocene on ε_{Nd} in the Atlantic basin. Three different LGM freshwater forcing experiments are performed to test the variability in ε_{Nd} oceanic distribution as a function of ocean circulation. Highly distinct representations of ocean circulation are generated in the three simulations, which drive significant differences in ε_{Nd}, particularly in deep waters of the western part of the basin. However, mean Atlantic LGM ε_{Nd} values are remain half a unit more radiogenic than for the modern control run. A fourth experiment shows that changes in Nd sources and bathymetry drive a shift in the ε_{Nd} signature of Northern end-members (NADW or GNAIW glacial equivalent) that is sufficient to explain the shift in mean ε_{Nd} during our LGM simulations. None of our three LGM circulation scenarios gives a better agreement with the existing ε_{Nd} paleo-data, as the model fails in reproducing the dynamical features of the area. Therefore, this study cannot indicate the likelihood of a given LGM oceanic circulation scenario. Rather, our modeling results highlight the need for data from western Atlantic deep waters, where the ε_{Nd} gradient in the three LGM scenarios is the most important (up to 3 ε_{Nd}). This would also aid more precise conclusions concerning the north end-member ε_{Nd} signature evolution, and thus the potential use of ε_{Nd} as a tracer of past oceanic circulation.

1 Introduction

Ocean circulation plays an important role in climate change since it is suspected to be an amplifier of, or may even trigger, shifts between glacial and interglacial periods (Broecker and Denton, 1989; Charles and Fairbanks, 1992; Rahmstorf, 2002). The meridional circulation structure (Meridional Overturning Circulation – MOC) of the

North Atlantic Basin redistributes heat to different latitudes. The southward transport of cold water at depth towards the Antarctic circumpolar current as the North Atlantic Deep Water (NADW) is compensated for by the northward transport of heat from the south in surface and shallow waters. This Atlantic overturning cell is a dynamic element of the oceanic thermohaline circulation (THC), acting on the atmospheric circulation and chemistry (CO_2 in particular), which are directly involved in governing climate. An ongoing problem for climatologists is to determine if the MOC will persist in the future and therefore to determine what controls its strength and variability. Studying past climate situations can help to understand the factors involved in controlling the MOC and clear up the role of the different forcings.

The Last Glacial Maximum (LGM) occurred 21 000 years ago and lasted for at least a few millennia and was a climate drastically different to today. Assessing how changes in the different components of the climate system (atmosphere, ice, land, and in this present study, MOC) control the overall climate is a fundamental question of determining which processes lead to different climate conditions. However, understanding the behaviour of past oceanic circulation, water-mass composition and flow patterns, remains difficult due to the multiple factors that control the distribution of relevant paleo-proxies. Different geochemical and isotopic paleo-tracers often give contradictory results (Lynch-Stieglitz et al., 2007) and up to three possible MOC scenarios are considered at LGM:

1. a highly stratified basin with a water mass at a maximum depth of 2200 m (with characteristics comparable to modern NADW; often referred to as Glacial North Atlantic Intermediate Water, GNAIW), overlying a large volume of water that originates from the Antarctic (which can be viewed as a more northward version of modern Antarctic Bottom Water, AABW). This view was first suggested by cadmium to calcium ratios (Cd/Ca) and carbon isotopes ($\delta^{13}C$) preserved in the fossilised shells of benthic foraminifera, which are used as proxies of the past nutrient distribution (Marchitto and Broecker, 2006; Duplessy et al., 1988; Curry and Lohmann, 1983; Charles and Fairbanks, 1992). Measurements of

radiocarbon (^{14}C) from benthic foraminifera suggest an older age for GNAIW at the LGM (Keigwin, 2004), which implies a slower circulation, with deep water ventilations times as great as 2000 years (Keigwin and Schlegel, 2002).

2. in contrast, other carbon isotope data (Curry and Oppo, 2005) and Cd/Ca data (Oppo and Rosenthal, 1994) suggest vigorous overturning was maintained, with a penetration of AABW as far north as 60° N. Protactinium and thorium isotopes ($^{231}Pa/^{230}Th$) seem to indicate shorter residence times for waters in the Atlantic basin, which would confirm vigorous overturning at shallower depth and weakened deep water ventilation (Yu et al., 1996; McManus et al., 2004). However, a large Pa/Th database suggests no large differences in the MOC between modern and LGM (Asmus et al., 1999).

3. oxygen isotopes ($\delta^{18}O$) from benthic foraminifera allow the reconstruction of water density at a given depth and suggest that during the LGM, the east-west $\delta^{18}O$ gradient was at least reduced or even reversed (Lynch-Stieglitz et al., 2006). These observations are consistent with a very weak GNAIW cell, but contradict the scenarios of circulation proposed by the other proxies that are mentioned above (Lynch-Stieglitz et al., 1999). On the other hand, alternative Cd/Ca data support a strong slowdown in the LGM MOC (Oppo and Fairbanks, 1987; Charles and Fairbanks, 1992; Broecker, 2002).

All these conclusions have to be taken cautiously due to the scarcity of the data characterizing the LGM and to the fact that the behaviours of these proxies in the past remain partially misunderstood (Lynch-Stieglitz et al., 2007). Overall, there is currently no consensus on the structure of the LGM ocean circulation. Apart from the necessary acquisition of more field observations, a better knowledge of the processes forcing glacial circulation would be aided by the representation of different circulatory proxies in numerical models. This has motivated the recent modelling of proxies at LGM (Henderson et al., 1999; Siddall et al., 2005), in order to better constrain the behaviour that results in the observed variability in the temporal distribution of such proxies.

Nd isotopic composition (hereafter expressed as $\varepsilon_{Nd}=((Nd^{143}/Nd^{144})_{sample}/(Nd^{143}/Nd^{144})_{CHUR}-1)*10\,000$, where CHUR is the Chondritic Uniform Reservoir, which represents the present day average earth value, $(Nd^{143}/Nd^{144})_{CHUR}=0.512638$ (Jacobsen and Wasserburg, 1980)), behaves conservatively in the ocean, aside from any lithogenic inputs, and is unaffected by biological cycles. Variations in ε_{Nd} have been measured in different water masses of the same water column, and this parameter has been used as a water mass tracer (Piepgras and Wasserburg, 1982; Jeandel, 1993; von Blanckenburg, 1999; Lacan and Jeandel, 2004; Amakawa et al., 2004). The modern Atlantic basin is characterized by two well-identified end-members: namely, a negative signature ($-13.5\pm0.5\,\varepsilon_{Nd}$; Piepgras and Wasserburg, 1980; Piepgras and Wasserburg, 1987; Lacan and Jeandel, 2005a) of NADW acquired in the Nordic and Labrador seas and also a less negative signal from the southern water masses (AAIW and AABW, $\varepsilon_{Nd}=-8\pm1$; Piepgras and Wasserburg, 1982; Jeandel, 1993). The evolution of ε_{Nd} along the modern day THC, from negative values in the north Atlantic, to positive values in the Pacific, makes ε_{Nd} a good candidate as a tracer of paleocirculation, and of the THC in particular. Planktonic or benthic foraminifera, benthic ferromanganese nodules and crusts, as well as iron-manganese oxides coatings are carrier phases that record the variations in ε_{Nd} from surface and bottom water mass signatures over different time scales (Elderfield et al., 1981; Vance and Burton, 1999; Albarede et al., 1997; Abouchami et al., 1999; Rutberg et al., 2000; Bayon et al., 2002; Van De Flierdt et al., 2004; Piotrowski et al., 2004). For example, Piotrowski et al. (2004) measured the Nd isotopic composition (*IC*) preserved in Fe-Mn oxides from LGM to mid Holocene in the South Atlantic, providing the first determination of circulation variations using Nd isotopic data at LGM. Nevertheless, current knowledge of Nd oceanic cycle is far from complete and uncertainties remain concerning the ability of ε_{Nd} to trace paleo-circulation. Indeed, this last study assumes no change in the ε_{Nd} in both northern and southern end-members at the LGM. Accordingly, the temporal variations observed are interpreted as changes in circulation and the relative contribution of the two end members. However, Lacan and Jeandel (2005a) demonstrated that the change in water mass mixing during the formation of the northern water mass component may directly affect the IC of this end-member. These authors also suggested that exchange of Nd between the sediment deposited on the margins and the seawater occurred. This process, named "Boundary Exchange" (dissolved/particulates interaction along the continental margin, hereafter referred as BE) is likely efficiently modifying the signature of water masses flowing along these margins (Jeandel et al., 1998; Lacan and Jeandel, 2005b; Arsouze et al., 2007). Therefore, the change in the continental weathering regime (which will alter sediment fluxes and the type of material deposited along the continental margin; Vance and Burton, 1999; Reynolds et al., 2004), as well as a shift in the sites of deep water formation to lower latitudes (Ganopolski et al., 1998) are likely to affect the Nd isotopic composition of the end-members. Therefore, a hypothesis of Nd end-member signatures that are invariant in time may not be robust.

In this study, we use a modelling approach to reconstruct the global scale distribution of ε_{Nd} at the LGM. Our aim is to investigate the extent to which the observed temporal variation in ε_{Nd} data reflects changes in either the thermohaline circulation, or the signatures of the two end-members, or some combination of the two. We therefore test the evolution of ε_{Nd} distribution under different representations of the LGM oceanic circulation generated by the IPSL (Institut Pierre-Simon Laplace) coupled model. Firstly, the characteristics of this model and the dynamical features of the different simulations performed are described. We then present ε_{Nd} distributions for each run, in order to compare the changes in simulated circulation between LGM and modern state. Finally, we compare the output with the available data and evaluate the importance of the different processes that generate ε_{Nd} distribution at the LGM.

CPD

4, 309–333, 2008

2 Simulation description

All atmosphere-ocean simulations used in the present study are performed with the IPSL_CM4 model (Ocean Atmosphere Global Circulation Model – OAGCM) developed at Institut Pierre Simon Laplace (Marti et al., 2006). The modules of this coupled model are LMDz.3.3 (LMD; Hourdin et al., 2006) in a 3.75°×2.5° resolution for the atmosphere circulation, the ocean part is the NEMO model in its coarse resolution ORCA2, (LOCEAN; Madec, 2008), and the associated sea ice component is represented by the LIM model (UCL-ASTR, (Fichefet and Maqueda, 1997; Goosse and Fichefet, 1999). The three constituent parts of the OAGCM (ocean, atmosphere and sea ice) are coupled using the OASIS coupler (CERFACS; Valcke, 2006).

The control simulation (modern run, cf. Table 1 and Fig. 2) is the pre-industrial simulation run for the recent IPCC exercise (IPCC, 2007, http://ipcc-wg1.ucar.edu/wg1/wg1-report.html). The MOC in this simulation is relatively weak (≈10 Sv) compared to the most recent evaluations (Swingedouw et al., 2007). This shortfall in the modern MOC is partly explained by the lack of convection in the Labrador Sea.

For the LGM simulations, the land-sea mask, topography and ice-sheet extent are prescribed according to the Peltier ICE5G reconstruction (Peltier, 2004). Accordingly, ice sheets cover Hudson Bay, the Baltic Sea, the Bering Strait and the Barents Sea. In addition, sea level is reduced by 120 m, relative to the modern run, which causes consequent changes in topography (extension of the Patagonian and New Foundland plateau) (Fig. 1). The boundary conditions have been set by reducing atmospheric concentrations of CO_2, CH_4 and N_2O to 185 ppm, 350 ppb and 200 ppb respectively, and using the 21 ky BP orbital parameters, according to the PMIP2 protocol (http://pmip2.lsce.ipsl.fr). Three runs are been performed that use a river routing scheme adapted for LGM conditions, i.e. addressing the impact of ice-sheets on river basins (Alkama et al., 2006; Alkama et al., 2008). The first simulation (LGMA) is the reference LGM run, obtained with the boundary conditions listed above. In LGMA, snow accumulates on the ice sheets and to close the freshwater budget, snow is redistributed over three latitude bands, with limits at 90S/50S/40N/90N. The 40N limits corresponds to the southernmost latitudes reached by icebergs during ice ages. In each latitude band, the excess freshwater, which we define as calving, is integrated and supplied to the ocean in the same latitude band. For the northern band, freshwater fluxes due to calving are delivered to the Atlantic and Arctic Oceans, but not to the Pacific. This simulation is characterised by a vigorous Atlantic MOC (18 Sv), 8 Sv stronger than the control run (cf. Table 1). The second simulation, LGMB, is obtained by redistributing all the freshwater that arises from calving into the northern latitude band and results in a MOC of 14 Sv. Additional artificial calving is added in this northern latitude band to last simulation, LGMC, which reduces the MOC to only 6 Sv. The additional freshwater added to the northern latitude band corresponds to 0.2, 0.28 and 0.4 Sv for LGMA, LGMB and LGMC, respectively (cf. Table 1). Even though changes in fresh water forcing remain relatively small, a variety of representations of the MOC result. LGMB is generally similar LGMA, where a dominant and vigorous water mass from the north fills the basin, but in contrast to LGMA, bottom water from the south enters to more northerly latitudes in LGMB. The MOC in LGMC can be viewed as a reproduction of one of the proposed LGM circulation scenarios (scenario 1), cf. above), wherein the influence of southern component water is increased and the northern component water flows south at shallower depth than current NADW (thus corresponding to the GNAIW). In addition, we also performed a simulation with modern boundary conditions but retaining the LGM land-sea distribution. Since a change in bathymetry induces a change in the definition of continental margin, and thus a change in Nd inputs, this simulation, hereafter referred as modernM, tests the sensitivity of the ε_{Nd} distribution to such factors (cf. next section). Circulation changes induced by using LGM bathymetry are not significant. AABW is slightly weaker than during the control simulation (≈3 Sv compared to 5 Sv), but the main structures and characteristic depths are conserved.

All LGM scenarios of ocean circulation were generated at the "Laboratoire des Sciences du Climat et de l'Environnement" (LSCE), and a detailed description of these simulations is to be published soon. A summary of the simulation characteristics and the overturning sections are provided in Table 1 and Fig. 2.

3 ε_{Nd} modelling

ε_{Nd} is simulated following the approach developed in Arsouze et al. (2007), which provides a detailed description of the parameterisation. The oceanic ε_{Nd} distribution is generated by a passive tracer model, using pre-computed advective and diffusive fields (Ethé et al., 2006). The only source/sink term taken into account is BE, which is parameterized as a relaxing term towards the continental margin ε_{Nd} value. The relaxing time used varies from six months in surface to 10 years at 3000 m depth, since this vertical configuration provides the best results for the modern ocean (Arsouze et al., 2007).

Since the decay of the radioactive isotope ^{147}Sm to ^{143}Nd takes much longer (half life of 106 Gy) than the studied time interval (about 20 ky), we assumed that there was no evolution in the isotopic signature of the margin due to natural radioactive decay between LGM and Holocene. In addition, since no major tectonic events have occurred since the LGM, the overall margin ε_{Nd} distribution was likely very similar to that of today. Therefore, we apply the margin ε_{Nd} composition established by Jeandel et al. (2007) for our LGM simulations. Finally, we assume that the vertical parameterisation and relaxing time that characterizes BE is unchanged at the LGM. This last hypothesis has to be taken cautiously, since terrigenous fluxes were possibly higher at the LGM (Franzese et al., 2006). However, until the impact of these fluxes on BE can be better constrained, we base our hypothesis on Tachikawa's results (Tachikawa et al., 2003), which found a minor impact of terrigenous flux variations on the ε_{Nd} signatures of deep water masses.

We acknowledge that this simplistic modelling parameterization can only resolve the first order representation of the oceanic ε_{Nd} distribution and we do not attempt to simulate the full oceanic Nd cycle. However, it successfully reproduces the modern ε_{Nd}

composition of the major ocean water masses (Fig. 2; and Fig. 5 in Arsouze et al., 2007) and thus it appears to be a potential tool with which to investigate the impact of variations in circulation and bathymetry on the distribution of ε_{Nd} and the end-member signature in the Atlantic.

4 Results

The section used here to present the ε_{Nd} distribution in the simulations has been chosen to fit the classical western basin section, and to compare with the only available data in the south (Piotrowski et al., 2004).

The modern simulation produces a ε_{Nd} distribution that is in good agreement with the existing data, even if slightly too radiogenic and has a ε_{Nd} composition of −11.5 and −7 for NADW and AABW, respectively (which favourably compare to −13.5 and −8 respectively, for the data, cf. Fig. 3). ModernM yields a more radiogenic mean ε_{Nd} distribution than the control run (+0.5 ε_{Nd}). The largest anomalies are observed in surface waters and at depth for the formation site of the northern end-member (+1 and +2 ε_{Nd}, respectively) (Figs. 3 and 4). This anomaly is subsequently propagated southward via the deep western boundary current.

The three LGM simulations produce a ε_{Nd} distribution that is somewhat similar to the modern distribution. LGM surface, intermediate and bottom waters are radiogenic (−5.5 to −8 ε_{Nd}), and sandwich a more negative ε_{Nd} (−8 to −10 ε_{Nd}) deep water mass from the north. Despite significant changes in ocean circulation during our scenarios, the global mean LGM simulated ε_{Nd} in the Atlantic basin are only 0.5 ε_{Nd} more radiogenic than the modern scenario (Table 1, Fig. 2). Nevertheless, important changes are observed locally. For example, during simulations LGMA and LGMB (which exhibit a strong NADW), as well as for simulation LGMC (typified by weak GNAIW), the northern end-member is up to 3 ε_{Nd} more radiogenic than for the control simulation (ε_{Nd}=−12.5 for control run, ε_{Nd}=−10.5 for LGMB and ε_{Nd}=−9.5 for both LGMA and LGMC). Further south (20° S), this northern end-member still remains 1 ε_{Nd} more radiogenic, with

CPD
4, 309–333, 2008

Title Page
Abstract | Introduction
Conclusions | References
Tables | Figures
◀| |▶
◀ ▶
Back | Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

very few differences between each LGM simulation.

The ε_{Nd} composition of the bottom water masses of the Atlantic basin is directly related to the relative influence of the northern and southern end-members. LGMA, simulating a deep and robust NADW cell with a negative ε_{Nd} signature, displays the lowest bottom values whereas LGMC, which represents a dominant radiogenic AABW in the basin, has the highest values. LGMB, with a water-mass composition that is intermediate between LGMA and LGMC, unsurprisingly has an intermediate ε_{Nd} distribution at the bottom. It can also be noted that the overturning circulation in LGMA and LGMB is so vigorous that it propagates the positive signature from southern surface and intermediate waters to northerly latitudes (up to $+1.5\,\varepsilon_{Nd}$, Fig. 3) and therefore influences the signature of the northern end-member. In addition, LGMC simulates large differences in ε_{Nd} between the western and eastern parts of the Atlantic basin (Fig. 4). Since AABW preferentially flows in the western part of the Atlantic basin, the negative ε_{Nd} signature of GNAIW during LGMC is reflected only in the eastern part. This east-west gradient in ε_{Nd} is not observed in the other LGM simulations, due to the predominant influence of northern water masses across the entire Atlantic basin.

The isotopic composition of the Atlantic sector of the Southern Ocean (south of 30° S) is also between 0.5 and 1 ε_{Nd} more radiogenic during the LGM, relative to the modern control run. This is due to an increased influence of the more radiogenic south Pacific water masses when the importance of the Antarctic Circumpolar Current (ACC) is greater. LGMA is typified by a more vigorous ACC circulation than LGMC, while LGMB is intermediate. Accordingly, the most positive signature of AABW in the Atlantic is obtained for the LGMA and LGMB simulations (Fig. 4).

We finally note that the core studied by Piotrowski et al. (2004) is located in a region of important ε_{Nd} gradient (Fig. 4).

CPD
4, 309–333, 2008

5 Discussion

The simulated ε_{Nd} never differs by more than ±3 units between the three LGM simulations. In addition, when ε_{Nd} is averaged over the whole basin, the resulting difference between each set of two simulations never exceeds $+0.1\ \varepsilon_{Nd}$ (Table 1). This shows that the average isotopic composition is not drastically influenced by the different circulation schemes under our parameterization scheme. However some clear tendencies can be underlined from these simulations.

Firstly, the isotopic composition of the whole Atlantic basin is $+0.5\ \varepsilon_{Nd}$ more radiogenic for modernM simulation than for the control run. Hence, the ε_{Nd} signature of the end-members is likely affected by changes in bathymetry and the modification of Nd sources. This is probably due to the presence of ice sheets situated over Barents Sea, Hudson Bay and even Baltic Sea, that could prevent BE and shorten residence time with some very negative ε_{Nd} margins (respectively −15, −25 and −18, Jeandel et al., 2007). The closure of Bering Strait, that prevents radiogenic waters from the North Pacific from entering the Arctic basin, appears to play a negligible role as the flux involved is only important locally. LGM topography and the presence of ice sheets in the north causes variability in the exchange of Nd between water masses and continental margins and therefore drives a change in ε_{Nd} for the northern end-member. In contrast, the southern end-member remains virtually insensitive to any change in bathymetry (less than $+0.2\ \varepsilon_{Nd}$ variation, during simulation modernM), while the three LGM scenarios result in AABW that is up to $0.7\ \varepsilon_{Nd}$ more radiogenic (Fig. 3). This difference might be explained by changes in circulation, and more particularly by the ACC strength that mixes both the Atlantic and the more radiogenic Indo-Pacific waters to compose the southern end-member signature, as suggested by previous works, (Duplessy et al., 1988; Charles and Fairbanks, 1992; Oppo and Rosenthal, 1994).

While the measured LGM ε_{Nd} absolute value of −6.5 (Piotrowski et al., 2004) is almost reproduced in all three LGM simulations, the measured gradient between modern and LGM ($+3\ \varepsilon_{Nd}$ in the data) is not fully represented (only $+1\ \varepsilon_{Nd}$ at maximum, Fig. 3).

CPD
4, 309–333, 2008
Neodymium modelling
T. Arsouze et al.

It is encouraging that we simulate the correct trend in ε_{Nd} between modern and LGM, but the amplitude of the variation remains too low. We can ascribe this deficiency to the poor model resolution, which is therefore unable to reproduce all the processes of AABW propagation and formation along the Antarctic margin. This means the model falls short in characterizing the ε_{Nd} composition of AABW. It is a classical problem of coarse resolution OGCM (Dutay et al., 2002).

The location of the core studied by Piotrowski et al. (2004) is situated in an area of rough topography, and is influenced (in the model) by a variety of factors aside from the relative influence of AABW and GNAIW. For example, ACC strength and associated sediment transport, variation of Indo-Pacific waters entering the basin via the Agulhas current (Franzese et al., 2006), are not fully resolved in the model. A higher degree of confidence in the LGM circulation changes based on the distribution of ε_{Nd}, could be obtained by studying temporal variations of bottom waters west of the mid Atlantic ridge, where the AABW preferentially flows. This area is more representative of the water mass influences in the basin, with a gradient directly dependant on the end-members contribution. Indeed, the signature of bottom waters in LGMA is representative of the northern end-member (ε_{Nd}=−10.5), whereas southern component signature (ε_{Nd}=−7.5) can be observed in simulation LGMC.

There is an urgent need to constrain the end-member isotopic composition at the LGM. This critical question is the key in evaluating the potential of ε_{Nd} as a water mass tracer as function of time. This would also provide complementary information if used together with other chemical and isotopic paleotracers. The present study clearly suggests a difference of at least +0.5 ε_{Nd} when compared to present. However, Van de Flierdt et al. (2006), measured Nd variations in deep-sea coral from the New England Seamounts during Younger Dryas (short time period at 11/13 kyr BP with a suspected similar oceanic circulation regime as LGM; Keigwin and Schlegel, 2002; Keigwin, 2004) and found signature of surface (−14.5 ε_{Nd}) and deep waters (−13/−13.5 ε_{Nd}) similar to present surface water and NADW signature, respectively. They therefore concluded that there was likely no variation in the composition of end-members with time.

CPD
4, 309–333, 2008

A depth profile of ε_{Nd} at LGM in the western part of the Atlantic basin would allow a clear statement about the signature of the NADW change and bottom measurements would provide a confirmation concerning the penetration of AABW in the basin, since it is suspected to flow northward as far as 60° N (Curry and Oppo, 2005).

6 Conclusions

Using a simple parameterisation that resolves the first order global ε_{Nd} composition in an OGCM, we have tested the impact on ε_{Nd} of different circulation scenarios in the Atlantic for the LGM, relative to a modern control run.

Run modernM (LGM land-sea mask with modern forcings) shows that the presence of ice sheets, without significant circulation variation, affects the IC of the basin, and generates changes in mean ε_{Nd} of +0.5 units. This is due to a shift in the composition of the end-members. The northen end member, in particular, does not acquire its modern negative signature, due to prevented exchange with highly negative margins protected by ice sheets at the LGM. On the other hand, the southern end-member is not significantly more radiogenic in modernM than during the control run.

The main characteristics of the three LGM simulations are the same (isotopic composition of the end-members, structure of the water masses that compose the basin), even though the circulation changes drastically from one to another. However, important changes in ε_{Nd} of the bottom waters in the western basin can be observed, that are coherent with the penetration of southern water mass and relative influence of components. As in modernM, mean ε_{Nd} is 0.5 more radiogenic than modern run, suggesting that although circulation changes do not play a key role in end-members ε_{Nd} acquisition, they are important in redistributing the characteristics of the basin.

So far, very few data concerning ε_{Nd} at LGM are available. The change in isotopic composition between present and LGM observed in the data provided by Piotrowski et al. (2004) is partially reproduced by the model, but the location of the core is not the most relevant place for a clear comparison with the coarse resolution model used in this study. This is because of a very complex bathymetry and dynamics within the area of the core. Rather, we propose a comparison between model and data in the north of the western part of the basin, where the composition of the end-members can be easily constrained, and the only factor controlling the distribution of ε_{Nd} is the relative influence of the northern and southern components.

Substantial progress has to be made in: 1) the understanding of the Nd oceanic cycle, so as to better reproduce the features that drive the temporal evolution in the ε_{Nd}, 2) in LGM modelling of ocean circulation, since simulations LGMB and LGMC were produced via the artificial addition of freshwater, and moreover, other scenarios suggested by other studies (for example scenarios 2) and 3) mentioned in the introduction) cannot be tested because they have not produced by OAGCMs; and 3) there is a need to obtain more data for Nd isotopes in order that Nd can be a consistent tool for intercomparison with other paleo proxies; and modelling effort still has to be provided.

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CPD
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CPD
4, 309–333, 2008

Table 1. Main characteristics of the simulations. Modern run is the reference run obtained with pre-industrial run forcings. Simulations LGMA, LGMB and LGMC are all produced with LGM forcings and boundary conditions (orbital parameters, ice sheets coverage and subsequent sea ice level drop, realistic river routing, atmosphere chemistry composition). The three LGM simulations are obtained with different calving fluxes (treatment of snow accumulating on the northern mid latitude ice sheets, which excess is redistributed over the ocean to close the freshwater budget). ModernM simulation is obtained with modern forcing and LGM land-sea mask.

Experience Name	Calving (freshwater from ice sheets melt redistributed north of 40° N)	North component water flow	South component water flow	Mean ε_{Nd} of the basin
Modern	–	9 Sv	5 Sv	−9.1
ModernM	–	9 Sv	3 Sv	−8.7
LGMA	0.2 Sv	18 Sv	1 Sv	−8.6
LGMB	0.28 Sv	14 Sv	2 Sv	−8.7
LGMC	0.4 Sv	6 Sv	4 Sv	−8.7

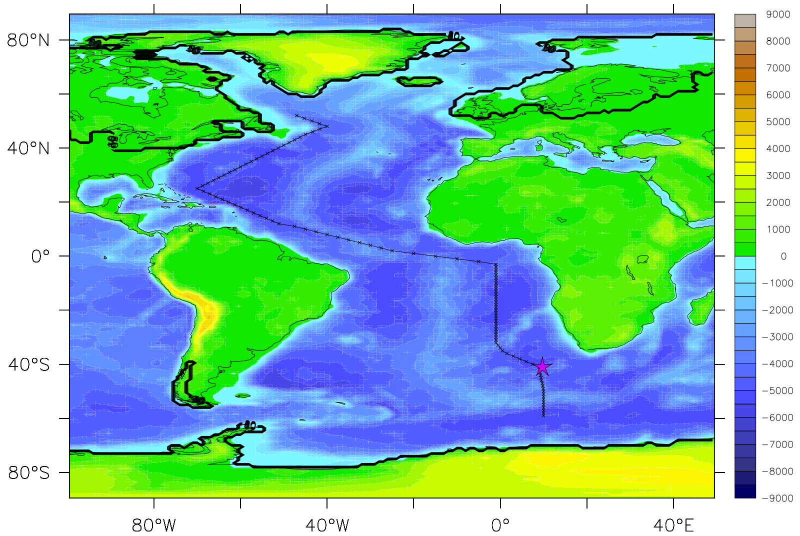

Fig. 1. Bathymetry and topography map at LGM, with modern continent contours. Sea level was 120 m lower at LGM than at Holocene (inducing larger Patagonian or New Foundland Plateau), with present oceanic regions covered by ice sheets (enclosed by thick line) over the Barent Sea, Hudson Bay or Nordic Sea, and the closure of the Bering Strait. The black line represents the trajectory of the vertical section in Fig. 3. Seawater ε_{Nd} available data at LGM is represented by a star (provided by and Piotrowski et al., 2004).

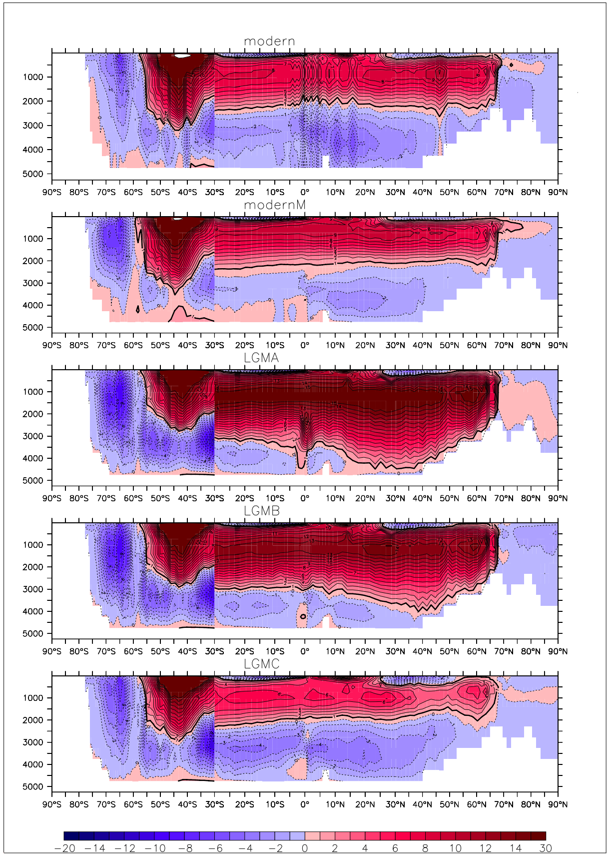

Fig. 2. Meridional overturning streamfunction (in Sv) of the Atlantic basin, north of 30° S, and of global ocean south of 30° S, for all simulations.

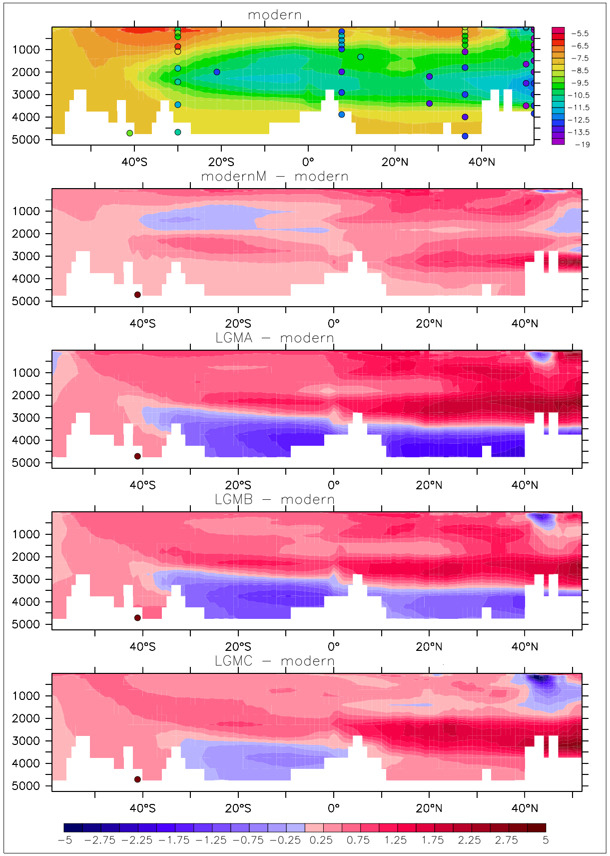

Fig. 3. Vertical ε_{Nd} section along the track represented in black in Fig. 1 for modern simulation (top panel). Variations between this reference modern run and modernM, LGMA, LGMB and LGMC simulations. Data are superimposed in circles with the same colour code than simulation output. LGM data are provided by Piotrowski et al. (2004). The color scale is non linear.

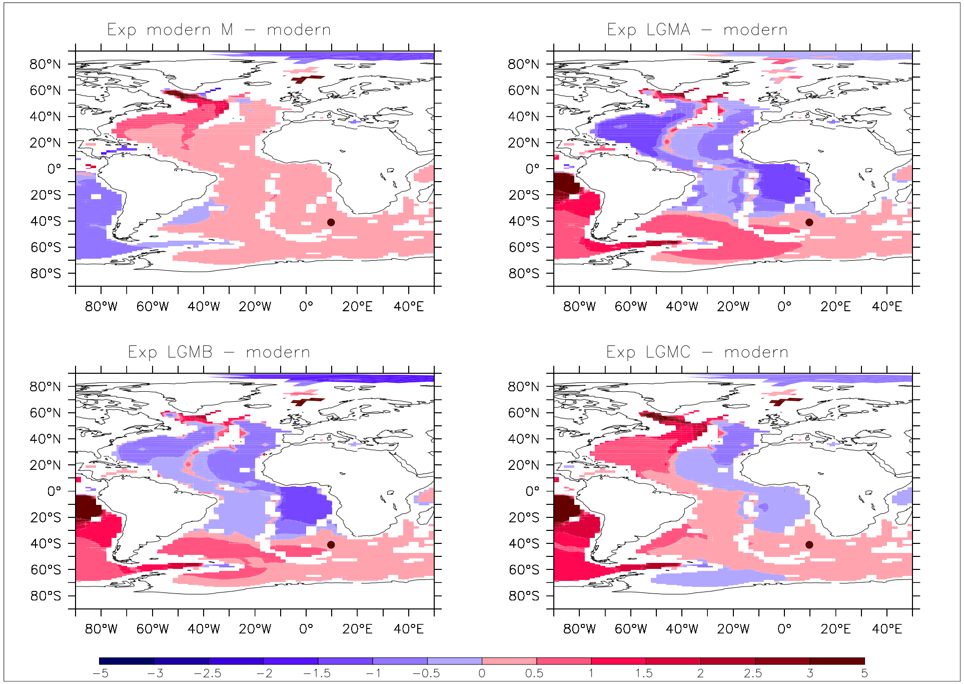

Fig. 4. Map of ε_{Nd} difference between reference run (modern run) and the four other runs (modernM, LGMA, LGMB, LGMC), averaged between 3000 and 5000 m. Data are superimposed in circles with the same colour code than simulation output. LGM data are provided by Piotrowski et al. (2004). The color scale is non linear.

CPD
4, 309–333, 2008