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Impact of strong deep ocean stratification on the glacial carbon cycle

N. Bouttes,¹ D. M. Roche,^{1,2} and D. Paillard¹

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[1] During the Last Glacial Maximum, the climate was substantially colder and the carbon cycle was clearly different from the late Holocene. According to proxy data deep oceanic $\delta^{13}\text{C}$ was very low, and the atmospheric CO₂ concentration also reduced. Several mechanisms have been proposed to explain these changes, but none can fully explain the data, especially the very low deep ocean $\delta^{13}\text{C}$ values. Oceanic core data show that the deep ocean was very cold and salty, which would lead to enhanced deep ocean stratification. We show that such an enhanced stratification in the coupled climate model CLIMBER-2 helps get very low deep oceanic $\delta^{13}\text{C}$ values. Indeed the simulated $\delta^{13}\text{C}$ reaches values as low as $-0.8\text{\textperthousand}$ in line with proxy data evidences. Moreover it increases the oceanic carbon reservoir leading to a small, yet robust, atmospheric CO₂ drop of approximately 10 ppm.

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1. Introduction

[2] During the Last Glacial Maximum (LGM (about 21,000 years ago)) the climate was characterized by colder global temperatures compared to the late Holocene (LH) [Adkins *et al.*, 2002], as well as larger ice sheets [Peltier, 2004]. The carbon cycle, which is intimately linked to the climate, was also different, with lower atmospheric CO₂ concentrations of 180–200 ppm inferred from measurements of air bubbles trapped in the Antarctic ice core [Neffel *et al.*, 1982; Barnola *et al.*, 1987; Petit *et al.*, 1999; Indermühle *et al.*, 2000; Monnin *et al.*, 2001]. This coevolution of the atmospheric CO₂ and climate has long been studied and several hypotheses proposed to explain it. On the glacial-interglacial timescale, the carbon cycle is composed of three main reservoirs: the atmosphere, the terrestrial biosphere, and the ocean. During the LGM the atmospheric inventory was reduced, and so was the terrestrial one. Indeed pollen-based reconstructions indicate an approximately 760–1040 GtC terrestrial biosphere decrease [Crowley, 1995] and marine sediments a 270–720 GtC decrease [Bird *et al.*, 1994]. Thus both the atmospheric and terrestrial reservoirs were lowered. As the total carbon is approximately constant on the timescale considered, it leaves the ocean, the biggest of the three reservoirs, as the potential increased one during the LGM. Consequently, the various hypotheses have focused on ways of increasing the ocean carbon storage.

[3] Since the ocean was likely responsible for the LGM CO₂ concentration, a common tool used to track the

changes in the carbon partitioning is the oceanic carbon-13, which is represented by the oceanic $\delta^{13}\text{C}$ defined as

$$\delta^{13}\text{C} = \left(\frac{R}{R_{\text{ref}}} - 1 \right) * 1000$$

with $R = \frac{^{13}\text{C}}{^{12}\text{C}}$

R_{ref} is the PDB (Peedee belemnite) carbon isotope standard, which corresponds approximately to average limestone [Craig, 1957].

[4] The carbon-13 variations reflect the changes of the carbon cycle in the various reservoirs, influenced by fractionation and oceanic circulation. While atmosphere-ocean gas exchange can influence oceanic $\delta^{13}\text{C}$ through modification of the end-member value at the deep-water formation site, the main source of fractionation comes from photosynthesis. The biosphere preferentially consumes the light carbon-12 over carbon-13, increasing the $\delta^{13}\text{C}$ in the surrounding environment. In the ocean, transport is added to the biological fractionation so that the horizontal distribution also depends on the circulation of the water masses. The $\delta^{13}\text{C}$ can then be interpreted as an indicator of the different water masses. As it can be measured in marine sediment cores, it represents a powerful tracer of the past circulation. Therefore understanding changes in the $\delta^{13}\text{C}$ distribution helps toward a better view of the LGM circulation and its impact on the climate. During the LGM, while the atmospheric CO₂ concentration was lower, the $\delta^{13}\text{C}$ distribution in the ocean differed from the present one, in particular with extremely lower deep values that could reach $-0.8\text{\textperthousand}$ [Curry and Oppo, 2005]. Even though various hypotheses have been proposed to account for the carbon cycle LGM changes, none has given complete satisfaction yet. In particular, the very low $\delta^{13}\text{C}$ remain unexplained. Therefore we propose and test a different hypothesis, the

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deep ocean enhanced stratification, aimed primarily at reaching those deep ocean low values.

[5] Before adding this mechanism, we have to take into account two well known and documented LGM characteristics that impact the carbon cycle: the temperature drop and the vegetation decline [Archer *et al.*, 2000]. As the temperature decreases, the ocean CO₂ solubility increases and the ocean can take more CO₂ from the atmosphere. This effect implies a CO₂ drop of around 10 ppm during the LGM without the sediment compensation. The alongside terrestrial biosphere reduction releases around 30 ppm of CO₂ into the atmosphere. This terrestrial biosphere decrease also leads to a -0.4‰ shift in the global oceanic mean δ¹³C. Yet, without any other change, combining these two effects cannot account alone for the reconstructed distribution of δ¹³C in the ocean, and would lead to an approximately 20 ppm increase of atmospheric CO₂ concentration, far away from the 80–100 ppm decrease observed. Hence it underlines the need for other mechanisms impacting the carbon cycle.

[6] To explain the remaining differences with the observed LGM δ¹³C and atmospheric CO₂, most of the mechanisms proposed so far have focused on enhancing the biological pump. Those ideas imply changes in the marine biological cycle, such as larger nutrient availability [Matsumoto *et al.*, 2002], higher C/P ratio [Broecker and Peng, 1982], iron fertilization [Martin, 1990], a shift of dominant plankton species [Archer and Maier-Reimer, 1994], or surface stratification [François *et al.*, 1997]. However, data proxies indicate that neither of these mechanisms is sufficient [Kohfeld *et al.*, 2005]. This result is corroborated by model simulations [Bopp *et al.*, 2003; Brovkin *et al.*, 2007] showing that even with extreme and unrealistic conditions these hypotheses do not manage to sufficiently decrease the deep oceanic δ¹³C as well as the atmospheric CO₂. On the other hand, other hypotheses tested using box models and 3-D oceanic models have focused more on the oceanic transport such as vertical chemical rearrangement [Boyle, 1988], reduced North Atlantic overturning circulation [Winguth *et al.*, 1999], or reduced Southern ventilation [Toggweiler, 1999; Toggweiler *et al.*, 2006].

[7] In line with the latter studies exploring the role of modified ocean circulation on the glacial carbon cycle we propose to test another mechanism in the ocean: enhanced deep stratification through low deep vertical diffusion. Indeed, according to sediment core proxy data [Adkins *et al.*, 2002], the deep glacial ocean should have been colder and much saltier, thus acquiring a higher density. This, in turn, should result in a huge, less ventilated, carbon reservoir. The deep ocean stratification therefore appears as a plausible mechanism to store carbon in the ocean and ultimately decrease the atmospheric CO₂ concentration [Paillard and Parrenin, 2004].

2. Methods

2.1. Model

[8] To investigate this mechanism and the impact on the carbon cycle we use the CLIMBER-2 fully coupled model. CLIMBER-2 is an intermediate complexity model

[Petoukhov *et al.*, 2000], well suited for the long-term simulations we run. It has a coarse resolution of 10° in latitude by 51° in longitude in the atmosphere, and 21 depth levels by 2.5° latitude in the zonally averaged ocean, which is precise enough to take into account geographical changes, while allowing the model to be fast enough to run long simulations. The model is composed of various modules simulating the ocean, the atmosphere, and the continental biosphere dynamics. Three LGM conditions are simultaneously imposed in the control simulation: the ice sheets [Peltier, 2004], the solar insolation [Berger, 1978], and the atmospheric CO₂ concentration for the radiative code (190 ppm, not used in the carbon cycle part of the model). CLIMBER-2 has already been used and evaluated in previous studies [Ganopolski *et al.*, 2001; Brovkin *et al.*, 2002a, 2002b, 2007].

[9] Although the carbon cycle is highly simplified in CLIMBER-2, it represents the main mechanisms. The carbon is distributed between the three reservoirs: the atmosphere, the continental biosphere and the ocean. The atmosphere exchanges carbon with the oceanic reservoir which contains a biogeochemical model. The latter simulates two plankton groups: phytoplankton and zooplankton, as well as dissolved organic and inorganic carbon, calcium carbonate, carbon-13, carbon-14, nitrate, and phosphate. Phytoplankton production depends on light, temperature and is phosphate limited. Fixed ratios uptake are employed for nitrate and phosphate. On top of production, grazing, transport, remineralization and dissolution are the fundamental mechanisms controlling the evolution of these tracers. As the model version used explicitly computes the evolution of the carbon cycle and carbon isotopes (such as δ¹³C) in every reservoir, it allows us to compare the model output with data from sediment cores.

[10] To account for all the mechanisms, we use a version of CLIMBER-2 with a closed carbon cycle. In this version carbon is dynamically exchanged between the atmospheric reservoir and the continental biosphere, which is explicitly computed at every time step. The total amount of carbon defined as the sum of carbon in all three reservoirs is fixed to get an atmospheric CO₂ concentration of 280 ppm during the late Holocene, and kept constant in all the simulations. Simulations are started without terrestrial vegetation, and run during 20,000 model years insuring an equilibrium for the carbon cycle. Although it represents an important mechanism, carbonate compensation was not taken into account in this study as the model did not include a sediment module. Its potential impact is discussed in section 3.4.

2.2. Results for the Control Simulations

[11] For the late Holocene and the standard Last Glacial Maximum simulations the results obtained are similar to previous studies. The Global mean temperature is 14°C for the LH and 11°C for the LGM. The atmospheric CO₂ increases from 280 ppm at LH to 302 ppm with LGM conditions. The continental biosphere declines from 2392 GtC at LH to 1704 GtC at LGM, i.e., a 688 GtC loss, in the range of proxy data reconstructions around 270–1040 GtC [Bird *et al.*, 1994; Crowley, 1995] and previous studies

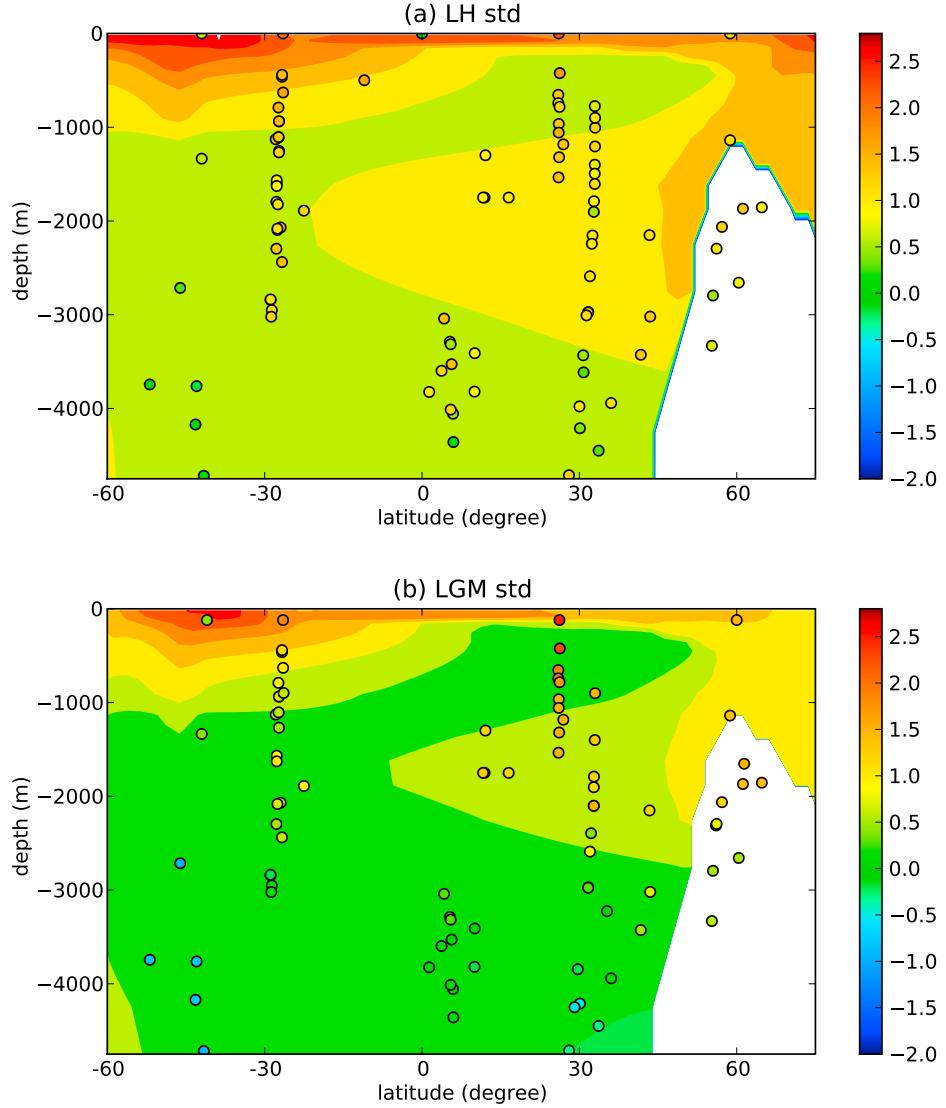


Figure 1. The $\delta^{13}\text{C}$ distribution in the Atlantic Ocean. The $\delta^{13}\text{C}$ distribution (a) during the late Holocene (LH) in the western Atlantic Ocean as reconstructed from proxy data [Kroopnick, 1985] (circle plot) and in the zonally averaged Atlantic simulated by CLIMBER-2 (contour plot) and (b) during the Last Glacial Maximum (LGM) in the western Atlantic Ocean as reconstructed from proxy data [Curry and Oppo, 2005] (circle plot) and in the zonally averaged Atlantic simulated by CLIMBER-2 (contour plot).

around 640 GtC [Brovkin *et al.*, 2002b]. The $\delta^{13}\text{C}$ distribution at LH represents fairly well the reconstruction obtained with the data (Figure 1), highlighting the different water masses, especially NADW (relative high values) and AABW (relative low values). The LGM distribution does not match as well the data, but is similar to previous studies [Brovkin *et al.*, 2007]. The loss of carbon from the terrestrial biosphere appears in the ocean through the shift in global mean $\delta^{13}\text{C}$ value falling from 0.57‰ at LH to 0.24‰ at LGM, i.e., a 0.33‰ drop. The supplementary uptake of carbon by the ocean due to the increased solubility and modified circulation roughly counterbalances part of the release of carbon from the continental biosphere (the ocean takes 642 GtC out of the 688 GtC loss from the vegetation). But it is by far not enough to explain the 80–100 ppm difference in the

atmospheric CO₂ concentrations measured between the glacial and interglacial periods (the 46 GtC loss from the vegetation not taken by the ocean increases the atmospheric reservoir, resulting in a 22 ppm increase of the atmospheric CO₂ concentration). Yet even with drastic changes in the marine biology, as previously simulated [Brovkin *et al.*, 2007], the results still differ from the data, the deep Southern Ocean $\delta^{13}\text{C}$ and atmospheric CO₂ remaining too high.

2.3. Stratification

[12] In order to achieve low $\delta^{13}\text{C}$ values in the deep ocean, we investigate the enhanced deep ocean stratification mechanism through sensitivity experiments that test various ways of stratifying the deep ocean during the LGM. In order to

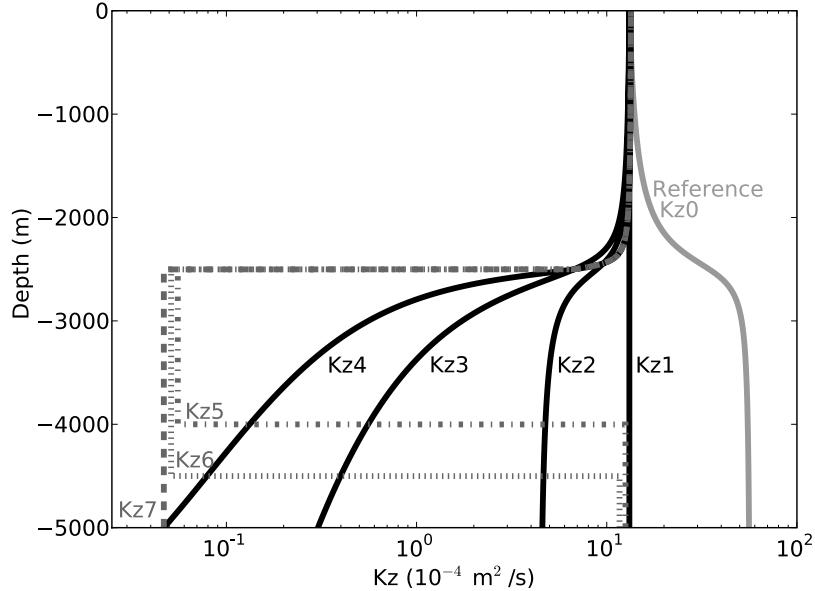


Figure 2. Vertical diffusion coefficient profiles used to enhance the stratification in the simulations.

modify the deep stratification we prescribe several vertical diffusion coefficients K_z (Figure 2). The diffusion coefficient K_z is a fixed parameter in CLIMBER-2 and represents vertical mixing. In a general circulation model (GCM) with a more complex 3-D ocean, it would depend on the water density in addition to another fixed coefficient. In reality the stratification impacts on the vertical mixing which also depends on turbulence and the tidal energy. In a fast model such as CLIMBER-2, this simple representation with a single parameter K_z allows us to easily modify the diffusion and thus the stratification. We can notice that contrary to most models, the diffusion coefficient in CLIMBER-2 is the same in the physical module and in the biogeochemical one, impacting both in a simple and above all coherent way. In a nutshell, by decreasing the K_z coefficient we reduce the vertical diffusion, leading to deep ocean stratification. It is worth noting that K_z is fixed and does not depend on stratification in the model, which allows us to easily alter it. Additional development is needed in order for the K_z to physically depend on local stratification. On top of progressively vertically decreasing coefficients (K_z 1, 2, 3, and 4) we define three extremely idealized coefficients to explore the relative importance of very deep (below 4000 m) stratification (K_z 5, 6 and 7).

3. Results and Discussion

3.1. Impact of the Deep Stratification on the Carbon-13

[13] Enhancing the deep ocean stratification appears as a very effective way of increasing the surface-deep $\delta^{13}\text{C}$ gradient (the $\delta^{13}\text{C}$ vertical gradient is defined as the gradient between the surface (0 to -2000 m) and deep (-3000 to -5000 m) oceanic $\delta^{13}\text{C}$ values at latitudes between 60°S and 30°S (Figure 3)), making the $\delta^{13}\text{C}$ distribution more similar to proxy data. Indeed, the $\delta^{13}\text{C}$ gradient was greatly higher during the LGM compared to the LH [Curry and Oppo, 2005]. The simulations clearly show that the more

the vertical diffusion coefficient K_z is diminished, the higher the vertical $\delta^{13}\text{C}$ gradient becomes. In other words, with a strong enhanced deep stratification the surface-deep $\delta^{13}\text{C}$ gradient is improved. The general $\delta^{13}\text{C}$ distribution is closer to the data reconstruction (Figure 4) with enhanced stratification. Extremely low values down to $-0.8\text{\textperthousand}$ are simulated, in line with observations [Curry and Oppo, 2005]. Yet some discrepancies remain, especially concerning the latitudinal distribution. In the data, $\delta^{13}\text{C}$ gets more negative further south, which is in contrast to our results (the minimum value is in the north in the simulations).

[14] At intermediate depth in the northern part of the Atlantic, Antarctic Intermediate Water (AAIW) seems to have inaccurate light $\delta^{13}\text{C}$ values. At this point we have to underline the fact that CLIMBER-2 is a 2-D model in the ocean which can therefore not represent perfectly all the mechanisms, even if it provides a good representation of the climatic and carbon variables. In particular CLIMBER-2 simulates a zonally averaged ocean which does not correspond to a simple latitudinal section and can therefore not be considered as such in comparisons with data. As discussed by Roche [2005], the zonally averaged Atlantic from CLIMBER-2 represents the thermohaline circulation, and does not represent the tropical gyres in surface, nor the eastern deep ocean. We can thus relatively easily compare the deep ocean simulated by CLIMBER-2 to the western Atlantic data, but the surface is not entirely representative of the western Atlantic. The main difference is the none representation of the tropical gyres which are low biological production regions. The CLIMBER-2 surface is representative of the high-production regions, thus regions with a high remineralization in the water column. This remineralization leads to low intermediate $\delta^{13}\text{C}$ values, as the organic carbon is preferentially composed of carbon-12. This can explain the low $\delta^{13}\text{C}$ values observed in all the simulated North Atlantic intermediate waters (late Holocene and glacial

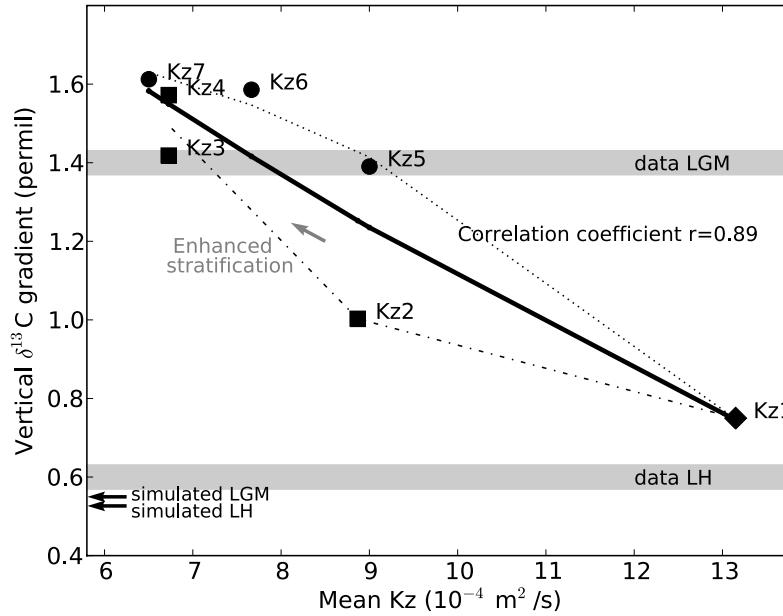


Figure 3. The $\delta^{13}\text{C}$ vertical gradient versus the mean vertical diffusion coefficient K_z . The $\delta^{13}\text{C}$ vertical gradient is the gradient between the surface (0 to -2000 m) and deep (-3000 to -5000 m) oceanic $\delta^{13}\text{C}$ values at latitudes between 60°S and 30°S . Simulated LH and simulated LGM refer to the $\delta^{13}\text{C}$ gradient values from the simulations with the original K_z ($K_z 0$), i.e., without enhanced stratification. Data LH and data LGM are the $\delta^{13}\text{C}$ gradients inferred from the present and LGM proxy data [Kroopnick, 1985; Curry and Oppo, 2005]. The result obtained with $K_z 1$ is a diamond; those obtained with $K_z 2, 3$, and 4 are squares; and those obtained with $K_z 5, 6$, and 7 are circles.

simulations, as well as with or without stratification simulations). Hopefully, it does not interact with the results of the sensitivity experiments we carry on when changing the deep ocean stratification since it is an independent mechanism. It will therefore not be further discussed.

[15] The lowering of the $\delta^{13}\text{C}$ values in the deep ocean is explained by the AABW lowering (Figure 5). The diffusion

coefficient K_z directly impacts the ocean circulation, and a K_z change, such as a lowering of K_z , modifies the ocean circulation. In this case a low K_z tends to reduce the ocean circulation as it diminishes the vertical mixing, and leads to a greater stratification. As the stratification becomes greatly enhanced the deep ocean is more isolated. The biological particles still sink and are degraded deeper in the ocean,

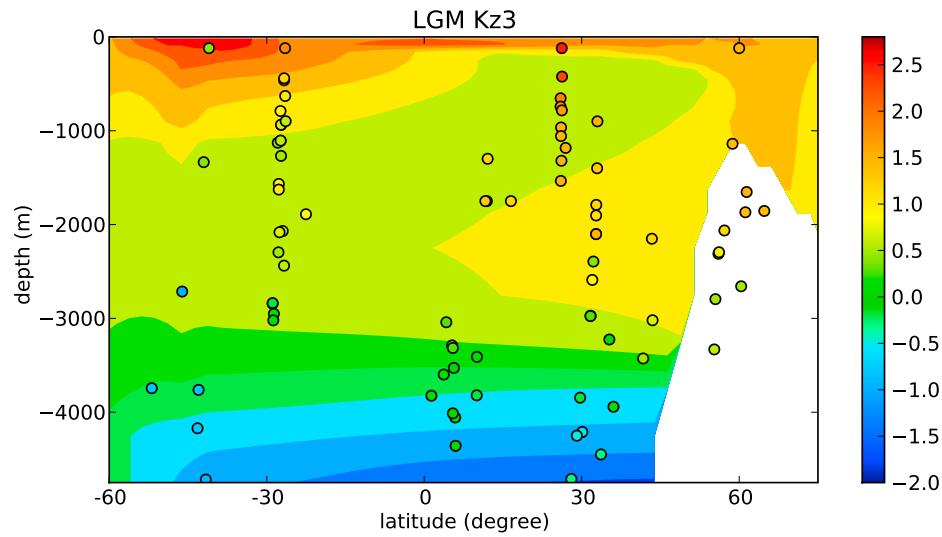


Figure 4. The $\delta^{13}\text{C}$ distribution during the LGM in the western Atlantic Ocean as reconstructed from proxy data (as in Figure 1b [Curry and Oppo, 2005] (circle plot)) and in the zonally averaged Atlantic simulated by CLIMBER-2 with an enhanced stratification $K_z 3$ (contour plot).

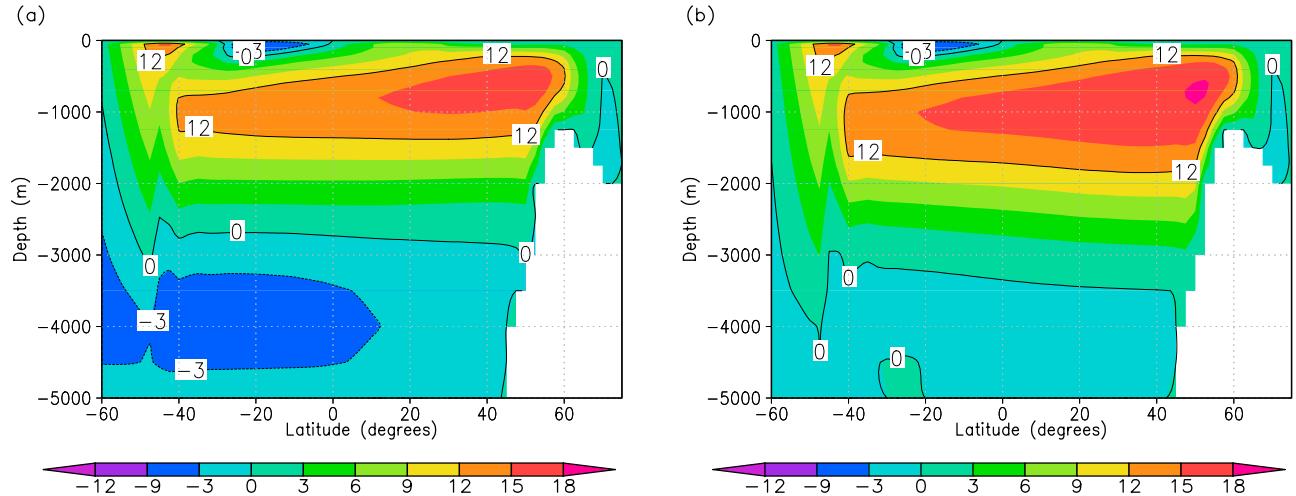


Figure 5. Vertical overturning stream function (Sv) in the Atlantic: (a) in the standard LGM simulation and (b) in the stratified LGM simulation with Kz 3.

releasing carbon-12 which decreases $\delta^{13}\text{C}$. Since the deep ocean is more isolated the deep carbon has a greater residence time and thus the $\delta^{13}\text{C}$ values are lower.

3.2. Impact of the Deep Stratification on the $p\text{CO}_2$

[16] The enhanced deep stratification also impacts the atmospheric CO_2 concentration. Indeed, the simulations emphasize the fact that the lower Kz gets, the greater the atmospheric CO_2 drawdown becomes (Figure 6). A very low Kz, i.e., a very strong deep stratification, favors a net atmospheric CO_2 drop, with a linear response of the deep

ocean. This effect on the atmospheric $p\text{CO}_2$ is not very large, but can be up to around 10 ppm which would represent 10% of the glacial $p\text{CO}_2$ drop.

[17] This CO_2 drop can be explained by the changes in dissolved inorganic carbon (DIC) and alkalinity (ALK) distribution in the ocean (Figure 7). Above approximately 3000 m deep DIC and ALK both decrease, while they both increase below. The decrease in surface and increase in the deep ocean of DIC explains the atmospheric CO_2 drop. Because of the enhanced stratification the biological carbon sinking can stay in the deep ocean, increasing the deep DIC.

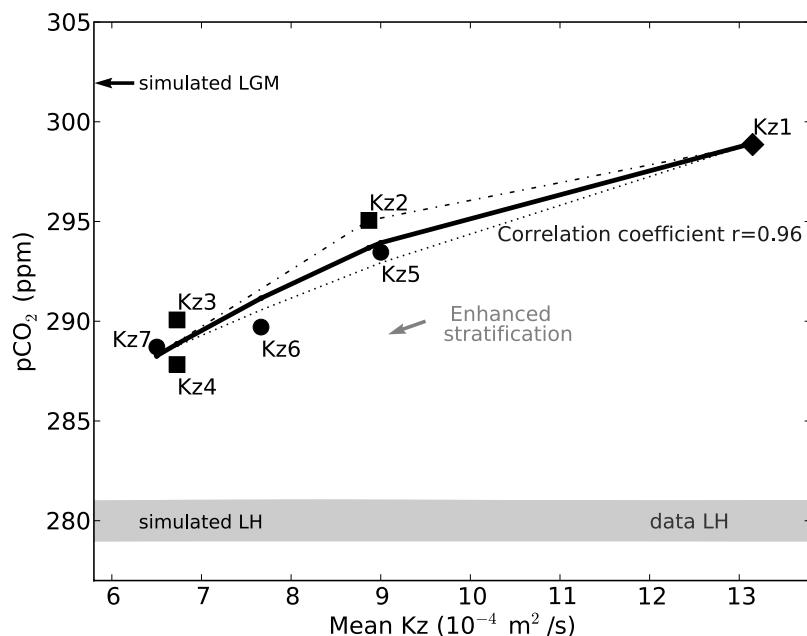


Figure 6. Atmospheric $p\text{CO}_2$ versus the mean vertical diffusion coefficient Kz. The result obtained with Kz 1 is a diamond; those obtained with Kz 2, 3, and 4 are squares; and those obtained with Kz 5, 6, and 7 are circles.

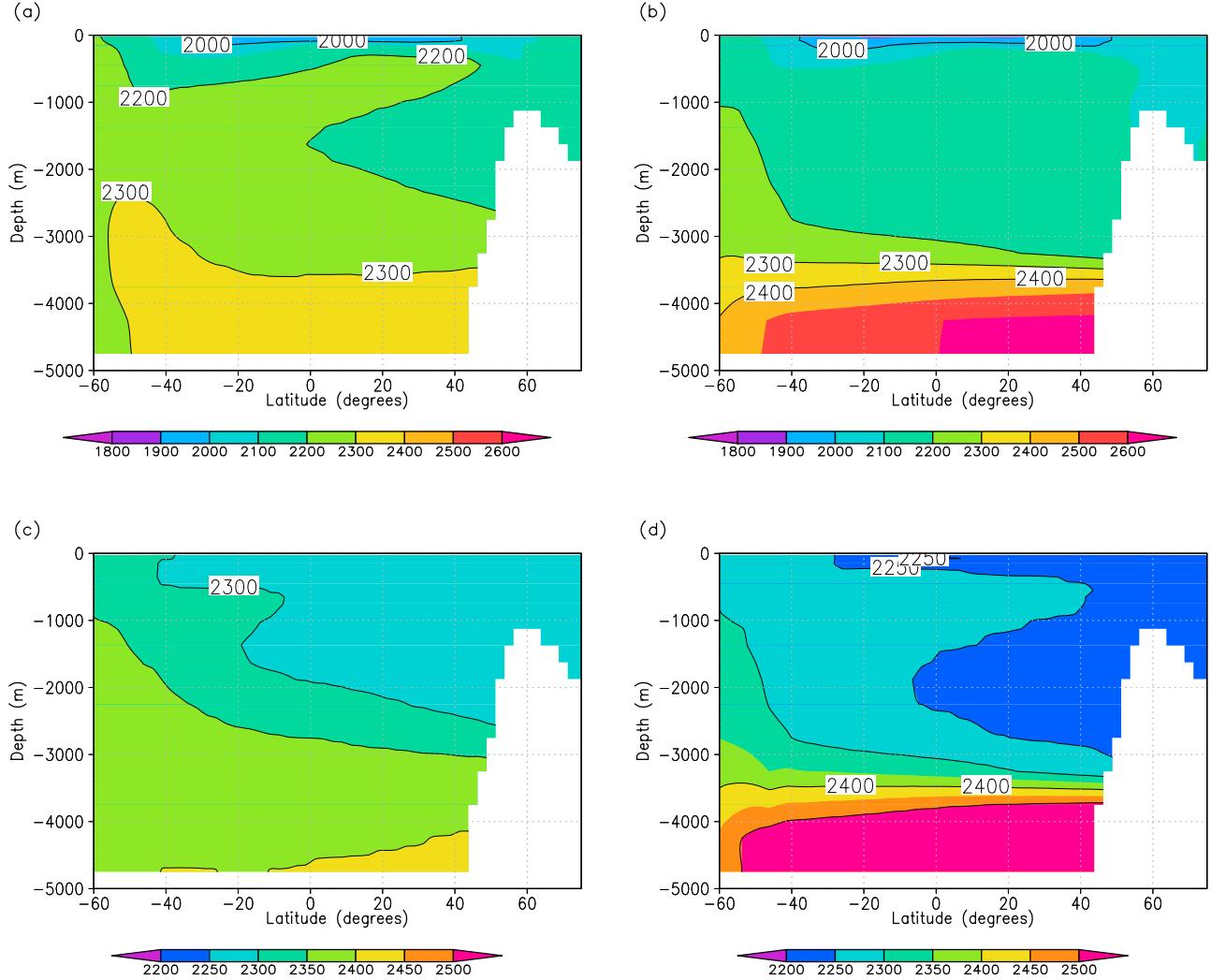


Figure 7. Dissolved inorganic carbon distribution in the Atlantic ($\mu\text{mol/kg}$): (a) in the standard LGM simulation and (b) in the stratified LGM simulation with K_z 3. Alkalinity distribution in the Atlantic ($\mu\text{mol/kg}$): (c) in the standard LGM simulation and (d) in the stratified LGM simulation with K_z 3.

The surface DIC is diminished since more carbon stays in the deep ocean and is not returned to the surface because of the stratification. When the surface DIC is lowered, the oceanic CO_2 becomes lower compared to the atmospheric CO_2 concentration and the ocean can get more CO_2 from the atmosphere, decreasing the atmospheric CO_2 until a new equilibrium is reached. The surface decreased DIC thus allows to take more CO_2 from the atmosphere. However, the same changes in alkalinity and nutrients partly counterbalance this effect. Indeed, the decreased surface alkalinity prevents from taking more carbon, since the CO_2 concentration in the ocean inversely depends on the alkalinity. When the alkalinity is lower the oceanic CO_2 is higher and the atmospheric CO_2 rises. Moreover, the reduced nutrient supply limits biological production as the phytoplankton cannot grow without nutrients. With less nutrients the production is then diminished and less carbon is taken by the marine biology, which also increases the surface CO_2 . This counteracts its initial decrease and leads to a smaller

atmospheric CO_2 uptake by the ocean. Overall, the net impact of enhancing the deep ocean stratification is to lower the atmospheric CO_2 , but can clearly not explain the 80–100 ppm drop by itself.

3.3. Spacial Distribution of the Stratification

[18] In this study we have presented idealized sensitivity cases of changes of the stratification obtained through modifications of the model vertical diffusion coefficient K_z . Such an enhanced stratification leads to better results in terms of $\delta^{13}\text{C}$ distribution, underlining the importance and necessity of this mechanism. Yet some discrepancies appear when the model results are compared to the data-based reconstructions. Those can be related to the model limitations, as we have seen with the zonally averaged oceans, but also to the specific study cases chosen, with the same coefficient K_z imposed everywhere. Contrary to these idealized cases, the stratification is unlikely to be the same everywhere, and the K_z coefficient should depend on

the region, according to the geography and physics of the considered region. Let us first consider the horizontal and then the vertical distribution of the stratification and their respective impacts.

[19] Concerning the horizontal distribution, the key region seems to be the Southern Ocean, as shown by pore fluids measurements used to reconstruct the LGM temperature and salinity [Adkins *et al.*, 2002]. Such data show that the LGM stratification was dominated by salinity variations, with the saltiest waters in the deep Southern Ocean. The salinity at the LGM was approximately 1 practical salinity unit (psu) more in the Southern Ocean than in the North Atlantic. The Southern Ocean is thus where the stratification mechanism should be the most important. Accordingly it is not surprising that the $\delta^{13}\text{C}$ latitudinal distribution mismatches the data since we have applied the same low diffusion for every latitude. Yet according to proxy data the changes should probably come from the south where the deep ocean was the saltiest during the LGM, thus very dense and stratified. Beyond the sensitivity tests run in this study the deep stratification should be looked at more closely depending on the localization in order to match more closely the data. It could then help getting a better $\delta^{13}\text{C}$ deep horizontal gradient with lower $\delta^{13}\text{C}$ values in the south.

[20] We can then consider the vertical distribution of the stratification. To investigate the impact of the deep stratification according to the depth we analyze three simulations with very idealized Kz coefficients (Kz 5, 6, and 7). They take low values from 2500 m and then high values below in the very deep ocean from 4000 m to the bottom (Kz 5), 4500 m to the bottom (Kz 6), or stay at low value (Kz 7). It appears that when Kz becomes greater in the very deep ocean (Kz 5 and Kz 6) it does not change drastically the results. The surface-deep $\delta^{13}\text{C}$ gradient is slightly smaller compared to the fully stratified case (Kz 7), but not significantly. Indeed, despite the higher Kz in the very deep part of the ocean, as long as the ocean above has a small Kz the deep waters below remain isolated, which is the fundamental mechanism, and therefore does not greatly change the results. Results are similar for $p\text{CO}_2$, though the greater very deep Kz has a more effective impact on $p\text{CO}_2$, and lessens the $p\text{CO}_2$. Atmospheric $p\text{CO}_2$ has a more linear response to the deep Kz. A greater deep ocean Kz also improves the horizontal deep ocean $\delta^{13}\text{C}$ gradient (Kz 5 and Kz 6). As can be seen in Figure 8, the abyssal ocean becomes more homogeneous with a high Kz (Kz 5) than with a low Kz (Kz 7), because a high Kz implies a higher vertical mixing than with a low Kz, i.e., a better mixing. The $\delta^{13}\text{C}$ horizontal gradient is diminished as it is better mixed, and gets closer to the data reconstruction [Curry and Oppo, 2005]. Such a higher Kz in the very deep ocean could physically come from the geothermal heating which can play a substantial role in bringing energy in the deep ocean, as recently pointed out [Adcroft *et al.*, 2001; Emile-Geay and Madec, 2008]. Finally, The main mechanism is clearly the isolation of the deep ocean that can store carbon with low $\delta^{13}\text{C}$, which is obtained by the deep stratification, and not drastically changed when the very deep ocean is better mixed as long as it is highly stratified just above.

3.4. Carbonate Compensation

[21] The carbonate compensation mechanism, which has not been explicitly taken into account in this study, is believed to be responsible for approximately 30 ppm of the glacial-interglacial CO_2 drop of 90 ppm [Broecker and Peng, 1987]. This additional mechanism tends to compensate CO_3^{2-} changes in the ocean on the timescale of a few thousand years by increasing or decreasing the accumulation of CaCO_3 in deep sea sediments. For instance when CO_3^{2-} concentration decreases in the ocean it tends to return slowly toward its initial value by an increase in the dissolution of CaCO_3 .

[22] A first attempt to represent this effect can be carried out by adjusting the whole ocean DIC and ALK inventories to restore the deep sea carbonate ion concentration. DIC and ALK are both modified, yet the ALK changes are greater than the DIC because a mole of CO_3^{2-} represents a mole of carbon ($\text{DIC} = +1$) but brings two negative charges ($\text{ALK} = +2$). As the ALK increase in the surface prevails it decreases $p\text{CO}_2$ in the surface and atmospheric CO_2 is reduced. By altering in this way the DIC and ALK concentrations for the low-diffusion simulation with Kz 3 the simulated CO_2 is 256 ppm. The additional CO_2 decrease due to carbonate compensation is thus 35 ppm, which adds up to the 11 ppm decrease due to low diffusion (with Kz 3) and represents a significant contribution to the CO_2 drop. The $\delta^{13}\text{C}$ effect is negligible (the vertical $\delta^{13}\text{C}$ gradient in the South Atlantic is 1.42‰, a value very similar to the LGM-Az3 $\delta^{13}\text{C}$ gradient of 1.44‰).

[23] Another test has been realized by changing the CaCO_3 remineralization profile in the model. CaCO_3 remineralization is increased in the upper part of the ocean (above 2500 m) and decreased below by changing the Suess [1980] function which sets the in and out fluxes of CaCO_3 at a given depth: $J \sim \exp(\frac{-Z}{Z_0})$ with Z the depth and Z_0 a characteristic depth which was originally 3000 m and was set to 2000 m in the modified remineralization profile. This new remineralization profile also leads to a CO_2 decrease of 21 ppm because the prevailing geochemical change is the ALK increase in the surface ocean. Here again, the $\delta^{13}\text{C}$ effect is negligible (the vertical $\delta^{13}\text{C}$ gradient is 1.40‰).

[24] These sensitivity tests both give a CO_2 decrease of about 30 ppm as a rough estimate with no significant effect on the $\delta^{13}\text{C}$ oceanic distributions. Still, the effect of carbonate compensation remains to be tested with an explicit model in presence of low vertical diffusion.

4. Conclusion

[25] In summary, we use the coupled climate model CLIMBER-2 to test the deep ocean stratification during LGM and its implications on the carbon cycle. The CLIMBER-2 model has already been used in several studies before and validated [Petoukhov *et al.*, 2000; Ganopolski *et al.*, 2001; Brovkin *et al.*, 2002a, 2002b, 2007]. In order to get the complete response of the system in term of carbon distribution, the model version used has a fully interactive carbon cycle module that takes into account the feedbacks between the climate and the carbon cycle. To test the impact of the deep ocean stratification we take advantage of the

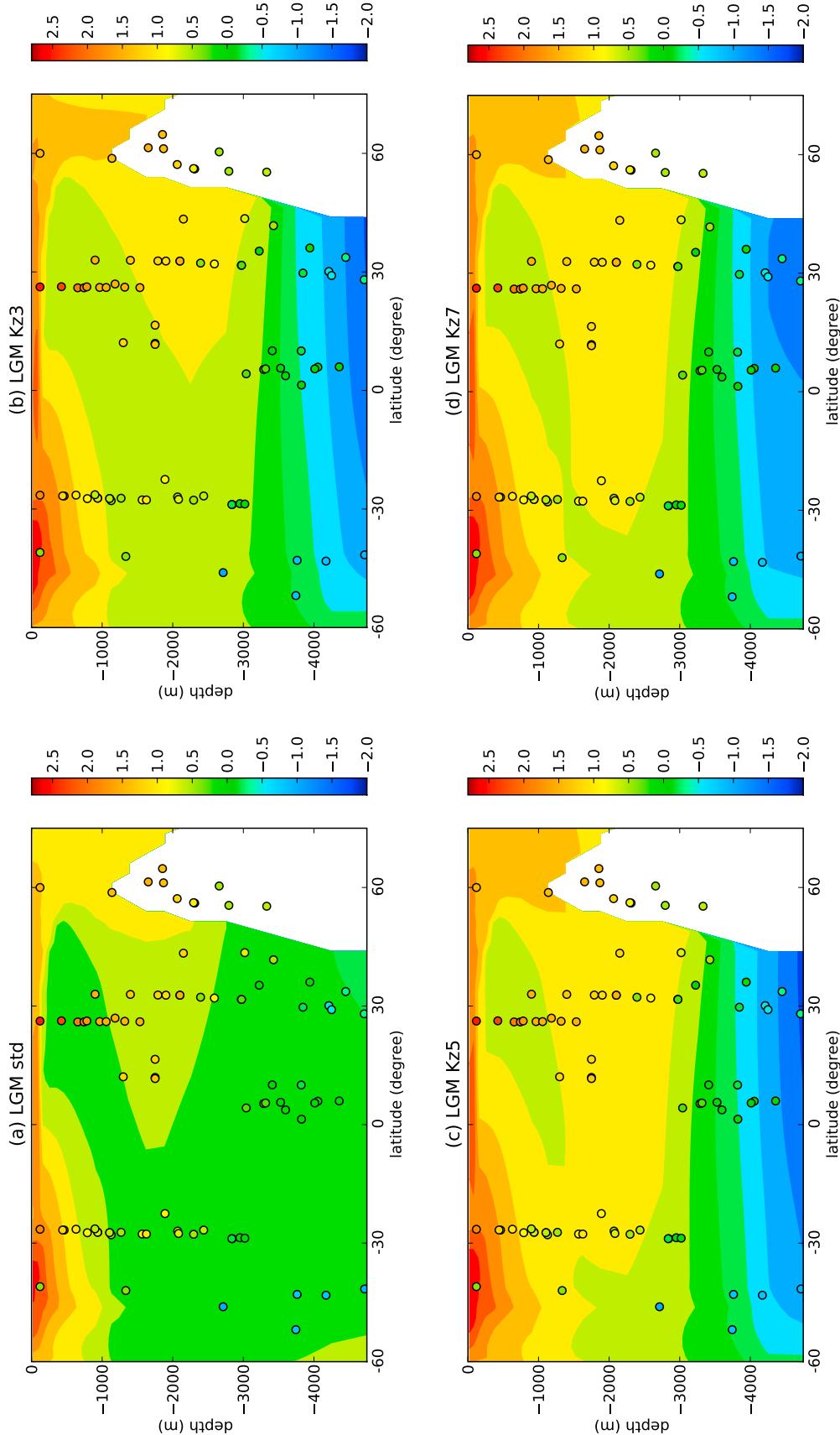


Figure 8. The $\delta^{13}\text{C}$ distribution in the Atlantic Ocean as inferred from proxy data [Curry and Oppo, 2005] (circles plot) and simulated by CLIMBER-2 (contour plot) during the LGM: (a) standard simulation, (b) with an enhanced stratification obtained with the diffusion stratification $Kz\,3$, (c) with a very idealized diffusion stratification coefficient $Kz\,5$, and (d) with a very idealized diffusion stratification coefficient $Kz\,7$.

simple modelization of the vertical diffusion in CLIMBER-2 based on a single fixed parameter K_z . We prescribe several profiles of K_z and analyze the oceanic $\delta^{13}\text{C}$ distribution and atmospheric $p\text{CO}_2$ modifications implied.

[26] Thanks to this stratification we manage to simulate very negative abyssal $\delta^{13}\text{C}$ values down to $-0.8\text{\textperthousand}$ during the LGM. These very low values were reached by greatly enhancing the deep ocean stratification, thus isolating the bottom waters. It therefore appears necessary to be in such a deep isolated ocean state to reach these very low $\delta^{13}\text{C}$ values similar to the reconstructed ones [Curry and Oppo, 2005]. The deep ocean stratification also impacts on the atmospheric $p\text{CO}_2$, as it leads to an approximately 10 ppm drop. The CO_2 decline is explained by the DIC increase in the deep ocean and simultaneous decrease in the surface. But the CO_2 uptake is counteracted by the same evolution of ALK and nutrients: the alkalinity decrease in the surface leads to a CO_2 release in the atmosphere. At the same time the smaller availability of nutrients in the surface lowers the biological production and thus also leads to a release of CO_2 in the atmosphere. The atmospheric CO_2 drop is therefore

robust though not very large because of the same changes in DIC, ALK and nutrients, yet accounting for carbonate compensation could amplify significantly the impact of deep stratification. A greater change requires to decorrelate DIC on the one hand and ALK and nutrients on the other hand.

[27] Furthermore, the proxy data show that the key region in this process is the Southern Ocean, where the salinity was highest during the LGM [Adkins et al., 2002]. Future work would need to focus on this region with a stratification different in space.

[28] Finally we have seen that the deep stratification is necessary, yet the very deep ocean does not need to be as stratified. As long as the ocean above is isolated because of the stratification the very deep ocean can be better mixed, as it is probably the case because of mechanisms such as the geothermy.

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References

- Adcroft, A., J. R. Scott, and J. Marotzke (2001), Impact of geothermal heating on the global ocean circulation, *Geophys. Res. Lett.*, 28(9), 1735–1738.
- Adkins, J. F., K. McIntyre, and D. P. Schrag (2002), The salinity, temperature, and $\delta^{18}\text{O}$ of the glacial deep ocean, *Science*, 298, 1769–1773.
- Archer, D., and E. Maier-Reimer (1994), Effect of deep-sea sedimentary calcite preservation on atmospheric CO_2 concentration, *Nature*, 367, 260–263, doi:10.1038/367260a0.
- Archer, D., A. Winguth, D. Lea, and N. Mahowald (2000), What caused the glacial/interglacial $p\text{CO}_2$ cycles?, *Rev. Geophys.*, 38, 159–189.
- Barnola, J. M., D. Raynaud, Y. S. Korotkevich, and C. Lorius (1987), Vostok ice core provides 160,000-year record of atmospheric CO_2 , *Nature*, 329, 408–414.
- Berger, A. L. (1978), Long-term variations of daily insolation and Quaternary climatic changes, *J. Atmos. Sci.*, 35, 2362–2368.
- Bird, M. I., J. Lloyd, and G. D. Farquhar (1994), Terrestrial carbon storage at the LGM, *Nature*, 371, 566.
- Bopp, L., K. E. Kohfeld, C. Le Quéré, and O. Aumont (2003), Dust impact on marine biota and atmospheric CO_2 during glacial periods, *Paleoceanography*, 18(2), 1046, doi:10.1029/2002PA000810.
- Boyle, E. A. (1988), The role of vertical chemical fractionation in controlling late Quaternary atmospheric carbon dioxide, *J. Geophys. Res.*, 93, 15,701–15,714.
- Broecker, W. S., and T.-H. Peng (1982), *Tracers in the Sea*, Lamont-Doherty Geol. Obs. of Columbia Univ., Palisades, New York.
- Broecker, W. S., and T.-H. Peng (1987), The role of CaCO_3 compensation in the glacial to interglacial atmospheric CO_2 change, *Global Biogeochem. Cycles*, 1, 15–29.
- Brovkin, V., J. Bendtsen, M. Claussen, A. Ganopolski, C. Kubatzki, V. Petoukhov, and A. Andreev (2002a), Carbon cycle, vegetation, and climate dynamics in the Holocene: Experiments with the CLIMBER-2 model, *Glob. Biogeochem. Cycles*, 16(4), 1139, doi:10.1029/2001GB001662.
- Brovkin, V., M. Hofmann, J. Bendtsen, and A. Ganopolski (2002b), Ocean biology could control atmospheric $\delta^{13}\text{C}$ during glacial-interglacial cycle, *Geochim. Geophys. Geosyst.*, 3(5), 1027, doi:10.1029/2001GC000270.
- Brovkin, V., A. Ganopolski, D. Archer, and S. Rahmstorf (2007), Lowering of glacial atmospheric CO_2 in response to changes in oceanic circulation and marine biogeochemistry, *Paleoceanography*, 22, PA4202, doi:10.1029/2006PA001380.
- Craig, H. (1957), Isotopic standards for carbon and oxygen and correction factors for mass-spectrometric analysis of carbon dioxide, *Geochim. Cosmochim. Acta*, 12, 133–149.
- Crowley, T. J. (1995), Ice age terrestrial carbon changes revisited, *Global Biogeochem. Cycles*, 9, 377–389.
- Curry, W. B., and D. W. Oppo (2005), Glacial water mass geometry and the distribution of $\delta^{13}\text{C}$ of ΣCO_2 in the western Atlantic Ocean, *Paleoceanography*, 20, PA1017, doi:10.1029/2004PA001021.
- Emile-Geay, J., and G. Madec (2008), Geothermal heating, diapycnal mixing and the abyssal circulation, *Ocean Sci. Discuss.*, 5, 281–325.
- François, R., M. A. Altabet, E.-F. Yu, D. M. Sigman, M. P. Bacon, M. Frank, G. Bohrmann, G. Bareille, and L. D. Labeyrie (1997), Contribution of Southern Ocean surface-water stratification to low atmospheric CO_2 concentrations during the last glacial period, *Nature*, 389, 929–935, doi:10.1038/40073.
- Ganopolski, A., V. Petoukhov, S. Rahmstorf, V. Brovkin, M. Claussen, A. Eliseev, and C. Kubatzki (2001), CLIMBER-2: A climate system model of intermediate complexity, part II: Model sensitivity, *Clim. Dyn.*, 17, 735–751.
- Indermühle, A., E. Monnin, B. Stauffer, T. F. Stocker, and M. Wahlen (2000), Atmospheric CO_2 concentration from 60 to 20 kyr BP from the Taylor Dome ice core, Antarctica, *Geophys. Res. Lett.*, 27(5), 735–738.
- Kohfeld, K. E., C. Le Quéré, S. P. Harrison, and R. F. Anderson (2005), Role of marine biology in glacial-interglacial CO_2 cycles, *Science*, 308, 74–78.
- Kroopnick, P. M. (1985), The distribution of $\delta^{13}\text{C}$ of ΣCO_2 in the world oceans, *Deep Sea Res., Part A*, 32, 57–84.
- Martin, J. H. (1990), Glacial-interglacial CO_2 change: The iron hypothesis, *Paleoceanography*, 5, 1–13.
- Matsumoto, K., J. L. Sarmiento, and M. A. Brzezinski (2002), Silicic acid leakage from the Southern Ocean: A possible explanation for glacial atmospheric $p\text{CO}_2$, *Global Biogeochem. Cycles*, 16(3), 1031, doi:10.1029/2001GB001442.
- Monnin, E., A. Indermühle, A. Dännenbach, J. Flückiger, B. Stauffer, T. F. Stocker, D. Raynaud, and J.-M. Barnola (2001), Atmospheric CO_2 concentrations over the last glacial termination, *Science*, 291, 112–114.
- Nefel, A., H. Oeschger, J. Schwander, B. Stauffer, and R. Zumbrunn (1982), Ice core sample measurements give atmospheric CO_2 content during the past 40,000 yr, *Nature*, 295, 220–223, doi:10.1038/295220a0.
- Paillard, D., and F. Parrenin (2004), The Antarctic ice sheet and the triggering of deglaciations, *Earth Planet. Sci. Lett.*, 227, 263–271.
- Peltier, W. R. (2004), Global glacial isostasy and the surface of the ice-age Earth: The ICE-5G (VM2) model and GRACE, *Annu. Rev. Earth Planet. Sci.*, 32, 111–149, doi:10.1146/annurev.earth.32.082503.144359.
- Petit, J. R., et al. (1999), Climate and atmospheric history of the past 420000 years from the Vostok ice core, Antarctica, *Nature*, 399, 429–436.
- Petoukhov, V., A. Ganopolski, A. Eliseev, C. Kubatzki, and S. Rahmstorf (2000), CLIMBER-2: A climate system model of intermediate complexity, part I: Model description and performance for present climate, *Clim. Dyn.*, 16, 1–17.
- Roche, D. M. (2005), Isotopes de l'oxygène en paléoclimatologie: L'apport d'un modèle de

complexité intermédiaire, Ph.D. thesis, Lab. des Sci. du Climat et de l'Environ., Univ. Pierre et Marie Curie, Gif-sur-Yvette, France.

Suess, E. (1980), Particulate organic carbon flux in the oceans—Surface productivity and oxygen utilization, *Nature*, 288, 260–263, doi:10.1038/288260a0.

Toggweiler, J. R. (1999), Variation of atmospheric CO₂ by ventilation of the ocean's

deepest water, *Paleoceanography*, 14, 571–588.

Toggweiler, J. R., J. L. Russel, and S. R. Carson (2006), Midlatitude westerlies, atmospheric CO₂, and climate change during the ice ages, *Paleoceanography*, 21, PA2005, doi:10.1029/2005PA001154.

Winguth, A. M. E., D. Archer, J.-C. Duplessy, E. Maier-Reimer, and U. Mikolajewicz (1999), Sensitivity of paleonutrient tracer distributions

and deep-sea circulation to glacial boundary conditions, *Paleoceanography*, 14, 304–323.

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