Answer to reviewers

We thank both reviewers for their comments, which we address point by point below. We have highlighted the corresponding changes in the manuscript in red.

Reviewer #1

General comments:

In this study, Bouttes et al. describe and evaluate the new land and ocean carbon cycle components coupled to the iLOVECLIM model. The manuscript is well written and easy to follow. They showed that the model is capable of broadly reproducing the observed distribution patterns of key ocean biogeochemical tracers. Statistically, the model also performed comparable or better than some more complex CMIP5 models. This is very encouraging and highlight the potential of using the computationally efficient EMIC model to study past climate variability involving the global carbon cycle interactions. The study fits very well within the scope of GMD, through documenting the new development and analyzing the new model performance. Below I have some suggestions which I thought could clarify improve the manuscript prior to publication. Most of them are quite straight forward to address.

Specific comments:

1) Land/Ocean imbalance. The title give away the impression of land and ocean carbon cycle, but there is an obvious imbalance in the content, e.g., 1 fig for land and 21 figs for ocean. I am not a terrestrial modeler and not qualified to give a fair judgement whether the presented land evaluation is sufficient for the purpose of future studies using this model.

The focus of this paper is indeed on the ocean since the model for the terrestrial biosphere already existed and has been described in Goose et al, 2010. We have added onexly minor changes to the terrestrial biosphere to include the isotopes in all carbon pools. As suggested we have modified the title to: "Including an ocean carbon cycle model into i LOVECLIM (v1.0)".

2) How well does the model conserve tracer's mass? If there is no sediment, do you assume all expor t, PIC/POC, are remineralized back into the water column before reaching the ocean floor? If not, is there any riverine fluxes? Does the model has any drift on the DIC/nutrients/O2/etc budget?

As first guessed by the reviewer, everything is remineralized/dissolved back in the water column. There is no riverine flux and no drift, all mass is conserved. We have made this clearer in section 2.2.1.

3) Biological production. I feel that the discussion around primary/expor t production can be improved. Many of the tracers (nutrients, O2, pCO2, ALK, DIC) shown here depend on the spatial distribution of surface primary and expor t production, as the author correctly noted in Fig. 1. Since the model includes an NPZD type ecosystem model, it would be useful also to show the surface distribution of NPP and export, and compare that with estimates from e.g., remote sensing data. How good is the annual

globally integrated NPP and expor t? What is the PIC/POC ratio in the model? Are the vertical remineralization profile is the same for PIC as for POC?

Unfortunately we haven't been able to plot the primary production due to issues with the output of the model, but it will be done in further work. The dissolution/remineralisation profile for POC and $CaCO_3$ is similar but with different coefficients.

On P3943, L17-18: what is the motivation for modifying the ver tical remineralization rate? Was it to improve model-data fit?

Yes

4) Air-sea oxygen fluxes. P3943, L4 states O_2 is prescribed to saturation values, but on P3948, L18: you mentioned there is an exchange with the atmosphere. How do you prescribed surface O_2 and at the same time prognostically simulate O_2 sources (to photosynthesis) at surface? Please reformulate the sentence. Can the authors elaborate why O_2 gas exchange is not implemented in the model? It should be relatively similar to the CO_2 fluxes and won't take much computational time.

It was a mistake on page 3948 l18, the oxygen is indeed prescribed to the saturation value at the surface as stated before in the manuscript. This has been corrected in the text.

Preliminary results to implement exchange of O_2 have shown that it is not completely straightforward and will require more work, therefore it will be done in further work.

5) Air-sea CO_2 flux. Which formulation is used? How is the spatial distribution compare to obser ved estimates? What is the annual CO_2 flux globally (also for land)?

The air-sea carbon flux depends on the difference between the partial pressure of CO_2 in the air and in the surface ocean with the gas exchange coefficient of 0.06 mol m-2 yr-1 (Brovkin et al., 2002). We have made this clearer in the text. We have compared pCO_2 with data (the air sea CO_2 flux data is computed from ocean pCO_2 measurements).

6) Oxygen is quite low in the deep North Atlantic. Is this because of the prescribed to the saturation state of O_2 at surface? If so, it is possible that bias in SST/SSS translates to/enhances this bias. Have you compared the saturated O_2 computed using WOA SST/SSS with surface O_2 data from WOA?

As suggested by the reviewer we have compared the saturated O_2 using WOA SST/SSS and surface O_2 data from WOA (see figure below). In the North Atlantic, the saturated O_2 computed from data SSS and SST is in better agreement with data compared to the O_2 from iLOVECLIM, showing that the low O_2 values in the North Atlantic in iLOVECLIM are due to the errors from the ocean model as discussed in section 3.2. The low values are then transported to the deep North Atlantic where it explains the disagreement with data, which we have added in the text in this section.
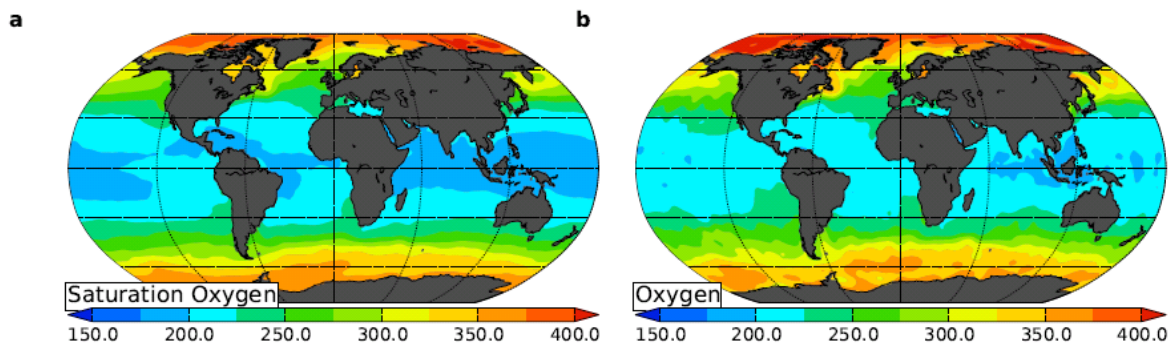
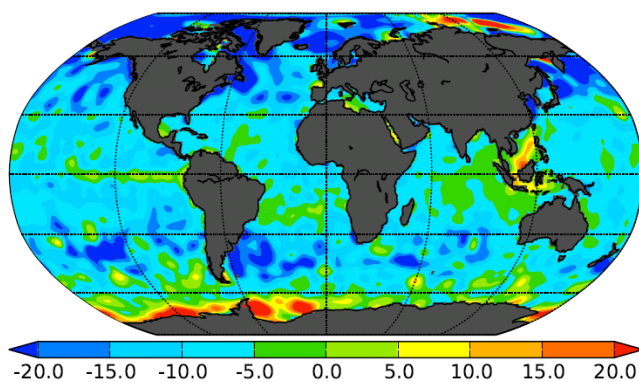Figure: oxygen distribution at the surface (a) computed at saturation using data SST and SSS and (b) from data.

Figure: difference between the oxygen at saturation and the oxygen from data.

7) Related to P3948, L20, in much of the high latitude Southern Ocean, O2 is also underestimated despite reasonable SST. Seems to contradict the statement on the NW Atlantic and Benguela upwelling regions. Maybe clarification on the biological coupling to the surface oxygen (point no.3 above) can explain this. The temperature is slightly overestimated by the model in parts of the Southern Ocean (Figure 4), leading to the underestimation of oxygen in these regions. This has been added to the text.

8) P3949„ L2: This is unconvincing to me. How much of this bias is due to the, say too strong remineralization rate at depth or lower surface productivity? It would be useful to plot the preformed vs regenerated PO4 (Duteil et al., 2012, Biogeosciences).

As suggested by the reviewer we have plotted preformed phosphate as in Duteil et al. (2012) (figure below). At depth the concentration of preformed phosphate is lower in the model compared to data, implying that there is not enough preformed phosphate and too much regenerated phosphate. This indicates either that there is too much production and remineralisation or that the ocean circulation is too sluggish. This is a very similar indication to carbon isotopes, hence we haven't included it in the manuscript.
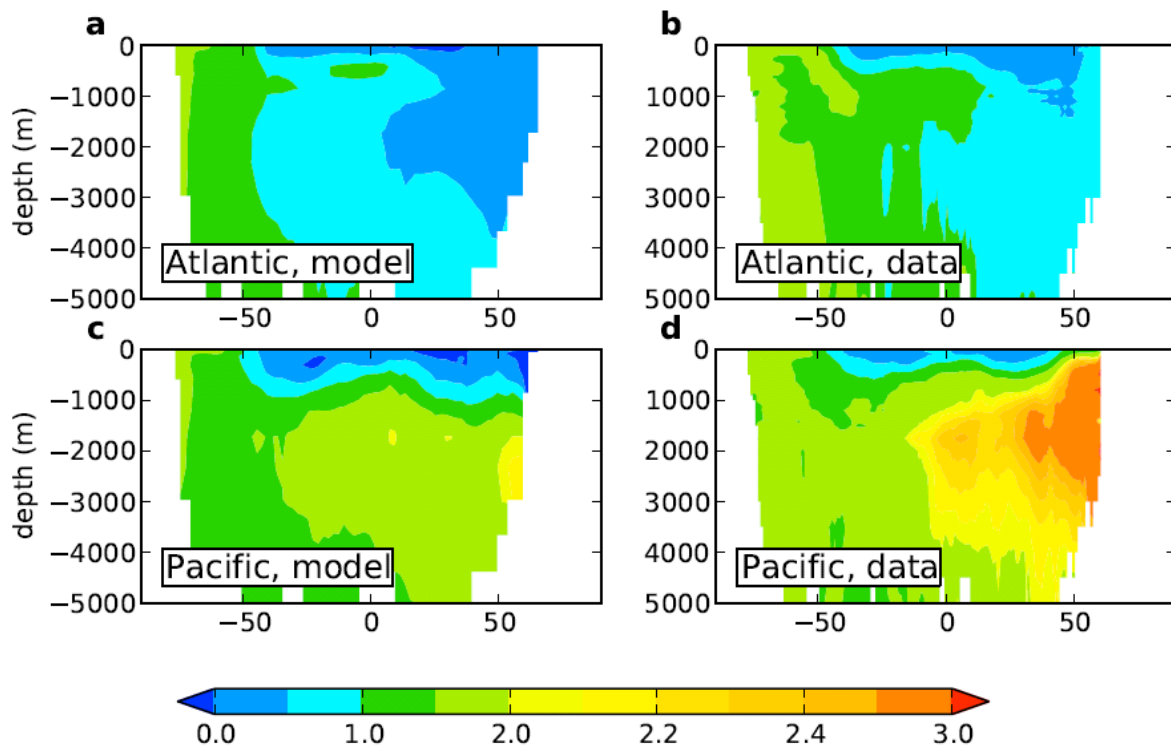

Figure: preformed phosphate computed as in Duteil et al. (2012): (a and c) for the model and (b and d) for the data.

9) P3950, L4: Maybe i misunderstood this, but by including iron cycle (limitation), production should decrease, less CO2 uptake by photosynthesis, thus even higher simulated pCO2.
As noted by the reviewer iron cycle is unlikely to explain the mismatch. The latter could be due to not enough convection, but more work will be needed to decipher the cause of the disagreement and to improve the model results. This has been modified in the manuscript.

10) For the disagreement in the delta13C values in the interior N. Atlantic, the authors attribute this to the too much diffusion (P3950, L15). It appears to me that the biology in the equatorial Atlantic also plays a critical role: too much export t production, which lead to too much remineralization, also seen in the O2 and PO4 signals. So the export t/surface production map would be useful here. What is the role of too weak AMOC, as noted in the manuscript? I understand that it is challenging to isolate the reason for this bias, but it just seems more than simply 'diffusion'.

As noted by the reviewer, it is difficult to separate the different effects, which should be done in future work.

Technical comments:
P3939, L28 is the same as L26. Remove or rephrase one.
This has been changed.

P3940, L19: consider revising the last par t of the sentence. E.g., replace "and improve our understanding and model simulations" with "to calibrate model simulations and improve our understanding."
Done.

P3941, L3: developed
Done

P3942 L26: some description on the carbon chemistry would be useful (e.g., OCMIP protocol?).
We have added more details in section 2.2.1 on the carbon chemistry, which is computed exactly as in Brovkin et al. (2002).

P3945, L25: clarify what is meant by "adjusted" here.
The total amount of carbon is set to obtain the correct atmospheric CO_2 using iterative runs. This has been modified in the text.

P3950,L13: north
Done

P3950, L14: do you mean the 'high' delta13C values?
Yes, this has been added.

P3951, L12: too much nor thward
Done

P3952, L12: Additionally
Done

P3952, L17: add "statistically" : statistically perform best . . .
Done

P3954, L6: reproduces
Done

P3959: add space in oceanmodels
Done

P3960, L23: replace 'e. a.' with 'et al.'
Done

Units missing from Figs 9-19. Adding latitude labels on the surface maps would be useful.
We have added the units and the latitude labels.

P3979, Fig 17 caption: 'distribution' and 'Takahashi et al.'
Done

P3982, Figs. 20/21/22 captions: clarify if this is surface/3D fields, area-weighted or not?
We have modified the captions to make it clearer (it is for the zonally averaged distribution).

Reviewer #2

This paper describes the introduction of an ocean carbon cycle model, including the carbon isotopes 13C and 14C, into the iLOVECLIM model (which already contained a terrestrial carbon cycle). I think this is a good paper that is acceptable with minor revisions.
The English is fairly good but could use a thorough editing by someone entirely fluent in English (the running title contains an unnecessary 'the' and there are lots of these in the text). The writing is adequate but could have benefited from a bit more thorough editing before submission. There are numerous references to processes not included in the model, with comments like "this could be included in future versions" or "this could be improved to be more realistic". I would prefer to see these statements deleted, and the paper focused on what was done, what was demonstrated, and what the strengths and weaknesses of the current model are. If they want they can include a paragraph in the discussion that speculates about what missing processes might account for some of the discrepancies between model and observations, but I would prefer that these not be sprinkled throughout the text.
There is a paragraph at the end summarizing future development, but it seems useful to us to explain the possible cause of a problem when it is identified, and what could be done in the future to improve it, especially in a paper on model development.

Main points:
(1) In general a lot of the claims about model skill in this paper are qualitative and subjective, and have a bit of an arm-waving quality about them. It's hard to tell exactly what is meant by "The thermocline is well represented" (3947/14), for example (see also 3948/16, 3948/25). The discussion of alkalinity on 3949/11-15 is not very convincing. Overall the model does not look very much like the observations, especially in the Atlantic, other than that it reproduces the global spatial pattern to first order (as does almost any model). The maxima associated with the subtropical gyres result from excess evapouration, and before one can even begin to speculate about the role of biological processes in generating the model errors one should look at the salinity-normalized alkalinity, because surface alkalinity to first order

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follows salinity. The salinity map in Figure 5 shows a pattern similar to the alkalinity but generally with a low bias in the Atlantic whereas alkalinity has a high bias. But I think it is wor th examining fur ther the role of the freshwater balance in generating the errors in modelled alkalinity in the Atlantic (contrary to what is claimed in the text, surface alkalinity is too high almost everywhere in the Atlantic and in some places these errors are quite large). I don't in general find subjective comparisons of maps very useful, and Figure 21 shows that when quantitative metrics of model skill are invoked, skill for alkalinity in the Atlantic is indistinguishable from zero.

In the Taylor diagrams (Figures 20-22), it appears that what is being compared here are the basin zonal means from Figures 12, 15 and 16. This is (obliquely) stated in the text (3952/5-7) but needs to be stated in the figure captions.

These are the only quantitative assessments of model skill in this paper and the captions do not state whether the Taylor diagrams are comparing ver tical or horizontal distributions or both, or over what latitude and longitude ranges. (Note that Figures 18-19 also appear to show basin zonal means but again this should be stated in the captions.)

I would also like to know why there are no Taylor diagrams or other quantitative skill assessments for 13C and 14C. Again all of the assessment is qualitative and subjective.

We have modified the captions of figures 18 to 22 to include the information about what is compared in the Taylor diagrams.

There is no Taylor diagram for 13C and 14C because there is no 13C data and no published 14C data available from OCMIP.

From looking at Figures 18 and 19 it looks like the model is doing a reasonably good job of simulating the distributions of 13C and 14C, but I would like to see a bit more discussion of what one expects, i.e., what sor t of model performance would we consider to be good enough, or what would we consider poor? Is there any literature to draw upon where similar experiments have been conducted with other models? Some of the statements regarding 13C strike me as questionable. For example, "13C also depends on circulation, so that NADW is characterized by relatively high values and AABW by lower values, in agreement with data." (3950/10-12). Surface water is enriched in 13C because that's where photosynthesis occurs, and deep water is depleted where there is a lot of cumulative remineralization (as in the deep Pacific), as described in the text at the beginning of this paragraph. So anywhere surface water is advected downward 13C will be high relative to surrounding waters at that depth. I don't see any reason for NADW and AABW to differ in this respect, and I don't see any evidence in the graphics to suppor t this statement. The 'column' of low 13C rising from the bottom at around 50S in the observations is probably upwelling (see e.g. Karsten and Marshall 2002 JPO 32: 3315) of water transported from the Pacific by the ACC.

The figure was probably misleading because of the interpolation and the small number of data. We have modified the d13C figure to use an improved dataset that contains the data that we were using before (GLODAP) in addition to other data (Schmittner et al., 2013). We show the observed values directly and don't do an interpolation anymore. The distinctive values of d13C for NADW and AABW are more easily discernible with the new figure. The difference between the d13C values is due to the fact that NADW is more ventilated than AABW and therefore has higher d13C values.

A discussion of the performance of the simulated C14 compared to other C14 models has been added in the paper.
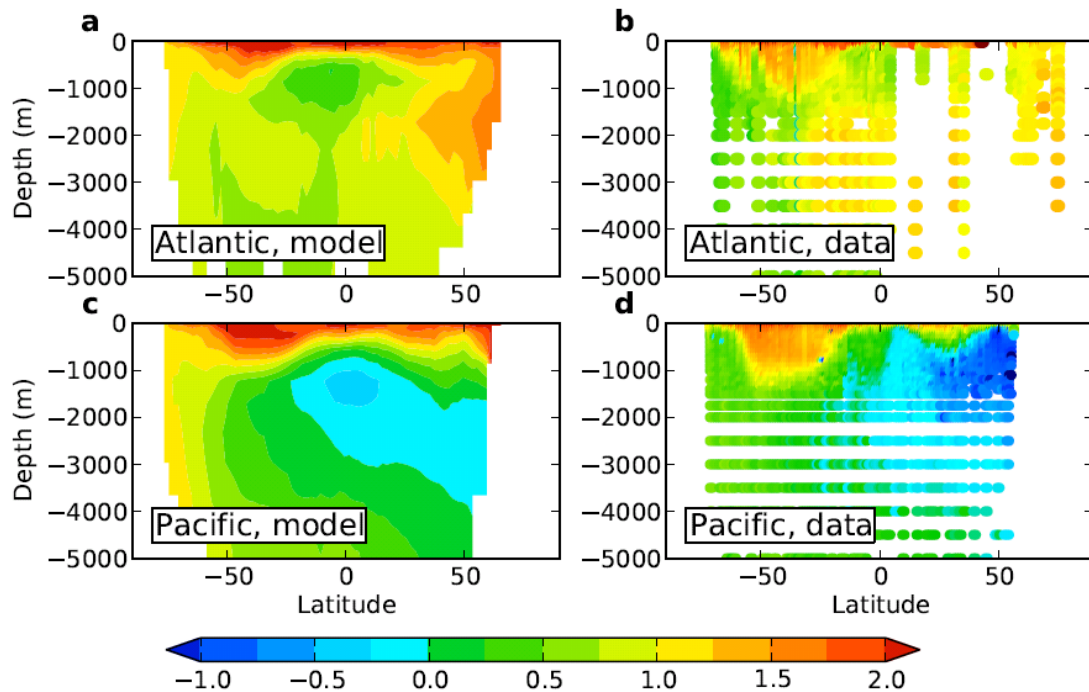
New figure for d13C (Figure 18) using data from Schmittner et al. (2013) and no interpolation.

(2) The pur pose of the whole exercise is presented as developing a model that can be used to run longer simulations to examine past (e.g. glacial) climates, but the description of the exact time scale intended is ambiguous. There are frequent references to a time scale of 100s to 1000s of years. This is long enough to simulate a climate with glacial boundar y conditions, but not to simulate glacial-interglacial cycles (e.g., 3939/29, 3940/21), and is not outside the capabilities of more complex models like IPSL-CM5A-LR. It would be possible, for example, for such a model run with glacial boundary conditions to be run to something fairly close to equilibrium (say 3000 years). Given time and resource constraints most climate modelling labs would probably not do such an experiment, but it is not outside the capabilities of modern supercomputers. If the intent is to simulate glacial-interglacial cycles or even glacial-interglacial transi-tions, a time scale of 100s to 1000s of years is inadequate. On 3941/12 it is stated that "The general goal of the new developments within iLOVECLIM is to include the suite of processes needed for climate simulations on the Milankovic timescale", which is a lot longer than 100s to 1000s of years. (Note also that the simulation was said to have been run for ~10000 years or "until it reached an equilibrium for deep ocean variables" (3946/1) but the criterion for convergence is not stated.)
The objective is to use the model for timescales of at least a few thousand years, this has been made clearer in the text.

(3) The errors in simulation of DIC and alkalinity are often attributed to the remineralization/dissolution profiles, but the description of what was done here is inadequate. Figure 2 refers only to POC and the text implies but does not actually state that the same profile is used for CaCO3 dissolution (3943/15-18). CaCO3 dissolution is frequently referred to as remineralization (e.g., 3949/14, 20, 23) which is not appropriate as CaCO3 is an inorganic compound.
We have replaced "remineralization" by "dissolution" when it refers to CacO3.

(4) The text is ambiguous as to what was done regarding the terrestrial carbon cycle. At some points it says that the terrestrial carbon cycle was already included and the purpose of the present research was simply to incorporate the ocean carbon cycle, but at other points it refers to modifications to the terrestrial carbon cycle model but does not state exactly what these are or what their relationship to the research presented is. The title refers to the "full" carbon cycle but could probably be changed to say simply "ocean" carbon cycle.
If nothing has meaningfully changed in the terrestrial carbon cycle model it is not clear to me that the descriptions of terrestrial vegetation at the beginning of section 3, or Figure 3, are necessary, especially given that there is no feedback from the ocean carbon cycle to the climate. "The total vegetation cover simulated by the model (Fig. 3) is in agreement with the one from another version of LOVECLIM" doesn't seem like a very useful analysis as it is essentially the same model. On 3951/2-3 it states that "The terrestrial biosphere has only been slightly modified to include the carbon reservoirs", but I find it hard to believe that these were not already par t of the model.
As discussed in question 1 of reviewer #1, we agree that the title might be misleading. The terrestrial carbon was already modelled and has been described previously, the only modification that has been made here concerns the carbon isotopes. This has been made clearer in the text and the title has been modified, see response to reviewer #1, question #1.

The description is also ambiguous regarding the unvegetated fraction. It is initially referred to as "bare soil" and later as "deser t" (3943/24). The latter term is not defined and in my view not appropriate as most deser t areas are not entirely devoid of vegetation. It appears to be equivalent to the "bare soil" fraction but this is not stated.
The desert fraction is the term used by Brovkin et al., 2002 (GBC) in the description of the terrestrial biosphere module.

(5) The description of the atmospheric CO2 boundary condition is also ambiguous. It is clear that the atmospheric radiation code does not use the simulated concentration and a constant concentration of 280 ppm is stated (3945/12). There are several subsequent references to the "data" or "measured" values, but no observational data set is referenced. I assume the "data value" means the specified constant value of 280 but this needs to be made clear. The rate of cosmogenic production of 14C is stated (3945/23), but it is not stated if this occurs at all levels of the atmosphere is only in the top one. This probably doesn't matter that much as only an areal rate is stated and the atmospheric circulation will redistribute it among the layers fairly quickly, but I think it is appropriate to specify whether or not it is initially injected into all levels or just the topmost one. Section 2.3 is rather verbose, without being very informative regarding the critical details of the model setup.

Regarding CO2, we have modified the text to make it clearer when we refer to the pre-industrial value of 280 ppm.

Regarding carbon 14, in the iLOVECLIM model the atmosphere component consists in just one box. The rate of cosmogenic production of 14C is stated in atom 14C cm-2 s-1 in the paper to be comparable with the value measured by Masarik and Beer (2009). In the model, this number is then integrated over the Earth surface. This integrated flux is thus applied once to the atmosphere box. This has been made clearer in the text.

(6) The simulation (which uses preindustrial boundary conditions) was compared with a contemporar y (1990-1999) climate from the CMIP5 historical experiment (3952/1- 12), but it is not stated why this was done. Preindustrial simulations are available for most of the CMIP5 models. Perhaps there should be a section in the Methods that describes the validation data sets (I consider the CMIP5 models to be among these, as comparing to higher-resolution ESMs is standard for evaluating an EMIC). Then such decisions could be explained up front instead of appearing out of nowhere in the Results or Discussion.

We agree with the reviewer that it would be better to compare our results with a closer period from the CMIP5 model simulations. Therefore we are now comparing with the model results for years 1890-1899 instead of 1990-1999. Because we're comparing the vertical distribution, which takes a few hundred years to be modified, it results only in very minor changes in the comparison.

As suggested, we have added a section on the data (new section 2.4).

(7) I don't understand the assumptions made regarding biological fractionation (or not) of 14C. Because 14C is the heavier isotope, I would assume that the fractionation effect is stronger than with 13C. In the present-day world the effect of isotope fractionation is negligible because a massive pulse of excess (bomb) 14C was put into the atmosphere over a ver y shor t period (effectively instantaneous relative to the time scale of ocean circulation). So biological fractionation plays a very small role in the redistribution of this excess 14C in the ocean. This does not necessarily mean that the biological pump as a whole plays a negligible role, although it is often assumed that it does. But I don't see why this should a priori be assumed to be the case in the preindustrial world where inputs of 14C to the atmosphere are much more gradual. At least, it should be clearly stated that biotic uptake and sedimentation are neglected. The present text simply states that fractionation is neglected (3944/23-25).

The reviewer is right: biological fractionation of 14C is considered to be twice the biological fractionation of 13C because 14C isotope is heavier than 13C. Thus, compared to 13C biological fractionation, 14C biological fractionation is clearly not negligible. The sentence in the text was misleading as it was written and it has been made clearer.

The reviewer is still right saying that fractionation should not be *a priori* assumed negligible when establishing a 14C representation in an Earth System Model. We made the choice of neglecting fractionation processes so that they can be directly compared to Δ14C values provided in observations (e.g. Key et al. 2004) and 'paleo' reconstructions (e. g. Burke and Robinson 2012). Δ14C are indeed equal to δ14C corrected for fractionation. Basically, the 14C concentrations are corrected in order to be directly convertible in time through the radioactive decay relation.

I can't make any sense of "14C in a reservoir is mainly driven by the distance between this reservoir and the atmosphere where 14C is formed by cosmogenic production.
The more distant this reservoir is from the atmosphere, the more depleted in 14C it will be, because of radioactive decay." (3945/1-4) This seems to imply that the 14C activity should decrease monotonically with depth, which is not the case since intermediate ocean waters are generally older than deep waters.
The sentence was indeed not clear. What we meant is that the age of a reservoir of 14C reflects the time since when this reservoir has been in direct contact with the atmosphere. Thus, for water masses, this implies that the longer they have been separated from a direct contact with the atmosphere, the older they are, which is compatible with some intermediate ocean waters being older than deep waters due to certain ocean circulation pathways, like in the Pacific Ocean for example. This has been made clearer in the text.

They might also want to state something about the range and precision of estimates of preindustrial del14C (3946/5-6). I think a model value of 1.5 permil vs an observed (?) value of 0 is very good, but not all readers will know this. Many readers might have no idea what the range or precision are. Also why is there a literature reference given for CO2 concentration and del13C but not for del14C?
For the range and precision of the model preindustrial value, it has been improved in the text. The reference for 14C has been also added.

(8) The boundary condition for oxygen could also be better explained. From what I can tell, there is no air-sea exchange: O2 is simply set to the saturation value in the surface layer of the ocean. But I find this hard to reconcile with Figure 11, which shows that the observed oxygen concentration is larger than the modelled over large areas of ocean. This in turn implies that the mean concentration over all of these areas is significantly supersaturated, which doesn't make sense to me. Supersaturation can develop when cold water warms rapidly or through excess photosynthesis, but at the surface gas exchange should limit supersaturation to within a few umol/kg of the saturation concentration. This could be due to errors in model SST but these don't look very large (Figure 4). It would be worth trying to explain this apparent anomaly because this figure looks suspicious to me.
The O2 is set to the saturation level at the surface, this has been made clearer, see response to reviewer #1 question #4 and 6.

Some details:
3938/7 "anthropogenic carbon emissions are predicted to continue in the future". Probably true but not consistent with the way the word "predict" should be used in climate science.
This has been changed to: Are likely to continue

3938/15 "The model can thus be used for long-term past and future climate–carbon studies." Again, probably true, but I don't see how this follows from what is stated in the previous sentence.
This has been removed.

3939/4 "influences the development of terrestrial biosphere and its decomposition" influences the development of the terrestrial biosphere and decomposition of terrestrial organic matter

This modification has been done.

3939/6 "climate-carbon retroactions" interactions
Done

3939/16 "range from simple box models to Global Climate Models" I don't recall there being any box models in C4MIP. I'm also not sure there were any models in C4MIP with a terrestrial N cycle as seems to be implied by 3939/17-19.
The sentence with box models in it was not referring to C4MIP. To make it clearer we have created a new paragraph there.

3940/8 and elsewhere change "values" to "concentrations" (e.g. 3948/27-28, 3949/1-3, 13, 28, 3951/16)
We have changed values to concentrations.

3940/13 "the measure of 13C and 14C in sediment cores" measurement
Done

3940/26 "avoid increasing the computing time exceedingly" excessively
Done

3940/28 "Processes linked to sediments" Sedimentary processes
Done

3941/8 "a code fork" a jargony term that could be avoided
It has been made easier to understand by using "new development branch". We have nonetheless also kept "code fork" as it is the correct description.

3941/11 "oxygen water isotopes" water oxygen isotopes
Done

3942/2 "a parameterisation of downsloping currents" downslope
Done

3943/13 "par t of it is exuded to DOC (and DOCs) and par t to POC" The term "exuded" is not appropriate here as it refers specifically to losses to the dissolved phase. The arrow in Figure 1 shows exudation going to DOC only.
This has been modified.

3946/24 "the distribution of the variables in the ocean" tracers? chemical species? biogeochemical fields?
Changed to tracers

3947/1 "The ocean dynamics, which depend on temperature and salinity gradients" is this really necessar y?
This is just a reminder.

3947/8 delete "perfectly"
Done

3947/13-26 I'm not sure it is appropriate to refer to all of the water masses named in the text without identifying them on any of the plots. For a reader familiar with oceanography it's not much of a burden, but in general it's not good practice.
We have added NADW and AABW on figure 6a.

3947/28 "compared to values between 14 and 31 Sv" This sentence gives no indication of what this range of estimates is for. Is it for CMIP5 models? Please clarify.
This refers to CMIP5 models studied in Weaver et al., 2012., This has been made clearer in the text by adding: "for CMIP5 models".

3948/9 "their use by marine biology" biota
Done

3950/17 "deconvoluate"
Changed to "separate".

3952/21 "most models simulate badly" poorly (see also 3952/24)
Done

Fig. 17 caption "distribution" misspelled
Done

Should the legends to Figures 20-22 refer to iLOVECLIM instead of LOVECLIM?
Done

Including an ocean carbon cycle model into *i*LOVECLIM (v1.0)

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Abstract. The atmospheric carbon dioxide concentration plays a crucial role in the radiative balance and as such has a strong influence on the evolution of climate. Because of the numerous interactions between climate and the carbon cycle, it is necessary to include a model of the carbon cycle within a climate model to understand and simulate past and future changes of the carbon cycle. In particular, natural variations of atmospheric CO_2 have happened in the past, while anthropogenic carbon emissions are likely to continue in the future. To study changes of the carbon cycle and climate on timescales of a few hundred to a few thousand years, we have included a simple carbon cycle model into the *i*LOVECLIM Earth System Model. In this study, we describe the ocean and terrestrial biosphere carbon cycle models and their performance relative to observational data. We focus on the main carbon cycle variables including the carbon isotope ratios $\delta^{13}C$ and the $\Delta^{14}C$. We show that the model results are in good agreement with modern observations both at the surface and in the deep ocean for the main variables, in particular phosphates, DIC and the carbon isotopes.

1 Introduction

The carbon cycle is a key component of climate and environmental sciences, both because CO_2 is a greenhouse gas (Tyndall, 1861) and has a direct impact on climate, but also because it plays an important role in ocean acidification (Orr et al., 2005) which directly impacts marine life. The three main carbon reservoirs involved on the timescale of a few thousand years are the atmosphere, the ocean and the land biosphere. The ocean is the biggest of the three reservoirs with around 39 000 GtC, while the atmosphere contains around 589 GtC and the terrestrial biosphere

between 1950 and 3050 GtC for the pre-industrial (Ciais et al., 2013). The climate also impacts the carbon cycle and hence the concentration of atmospheric CO_2 through various dynamical, chemical and biological processes. For example, changes in the ocean temperature will modify the solubility of CO_2: the warmer the ocean the less soluble CO_2 becomes, which decreases the carbon stock in the ocean and increases atmospheric CO_2. Temperature, as well as humidity, also influences the development of the terrestrial biosphere and decomposition of terrestrial organic matter. Low temperature and dry conditions tend to favor lower rates of decomposition. The various climate-carbon interactions involve all three carbon reservoirs. Therefore it is necessary to include a model of the carbon cycle within a climate model to understand past changes and anticipate the future evolution of the carbon cycle and climate.

Such models have been developed during the last decades (Cox et al., 2000; Friedlingstein et al., 2001) and a subset of coupled models used in CMIP5 (Coupled Model Intercomparison Project 5) now include a complete description of the ocean and land carbon cycles. Eleven models have been compared within the framework of the fourth Coupled Carbon Cycle Climate Model Intercomparison Project (C4MIP) (Friedlingstein et al., 2006). They include models of both the ocean and the land carbon cycle.

Climate models range from simple box models to Global Climate Models (GCM). The carbon models have gradually become more complex by including more types of plankton in the ocean and more plant functional types on land, as well as more nutrients, such as iron in the ocean or nitrogen on land (Anav et al., 2013). The number of additional tracers directly impacts the computing time, therefore such complex models are well suited to study the climate-carbon evolution on timescales of a few decades to hundreds of years, but are too computentially expensive for longer simulations. Sim-

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pler carbon models such as the ocean carbon models based on NPZD (nutrient-phytoplankton-zooplankton-detritus) ecosystems, and simple terrestrial biosphere models with a few plant functional types, associated with intermediate complexity climate models, are thus more convenient for the study of long timescales of more than a few thousand years.

Intermediate complexity models are well suited for long term studies of a few thousand to hundred of thousand years, and in particular the glacial-interglacial cycles. The carbon cycle varies greatly during the glacial and interglacial periods, with atmospheric CO_2 concentrations of around 190 ppm during the relatively colder glacials periods and around 280 ppm during the warmer interglacials (EPICA community members, 2004). Although such periods have already been studied with intermediate complexity models (Brovkin et al., 2007; d'Orgeville et al., 2010; Bouttes et al., 2010; Tschumi et al., 2011; Menviel et al., 2012), large uncertainties remain concerning the processes responsible for the changes of the carbon cycle.

Besides understanding and simulating CO_2 concentrations in the past and future, the carbon cycle also provides indirect yet valuable information about changes of the ocean dynamics and biology, as well as the land vegetation, through carbon isotopes changes (Duplessy et al., 1988; Crowley, 1995). Indeed, there is no direct data of ocean circulation changes in the past (except for the last decades, see for example (Mielke et al., 2013)), but the measurement of $\delta^{13}C$ and $\Delta^{14}C$ in sediment cores can help constrain the ocean and land vegetation changes. Moreover, the measure of atmospheric $\delta^{13}C$ in ice cores (Lourantou et al., 2010; Schmitt et al., 2012) and the calibration curves of atmospheric $\Delta^{14}C$ (Reimer et al., 2013, 2009) provide additional data and constraints. By explicitly simulating the carbon isotopes within the carbon cycle model, as we have done in the *i*LOVECLIM model, it is possible to directly compare model results with data to calibrate model simulations and improve our understanding.

Our long-term objective is to study past and future carbon cycle changes over timescales of a few thousand to hundred of thousand years, typical of glacial-interglacial changes. The *i*LOVECLIM model is perfectly suited to such studies since it includes the relevant physical and dynamical components of the ocean, atmosphere and terrestrial biosphere while running fast enough to simulate thousands of years in a reasonable amount of time (500 simulated years per day). To avoid increasing the computing time excessively, the ocean carbon cycle that we included in *i*LOVECLIM is based on a NPZD ecosystem which provides the main mechanisms relevant on the timescales of hundreds to thousands years, and includes the carbon isotopes. Sedimentary processes would also be relevant to such timescales. However, the introduction of a sediment model is beyond the scope of this study and remains to be done in future work. The terrestrial biosphere already included in *i*LOVECLIM has been further developed to add the carbon pools and carbon isotopes. Here, we evaluate the results obtained by including the model of ocean carbon in *i*LOVECLIM. We focus on the main variables from the carbon cycle and on the ocean carbon isotopes ($\delta^{13}C$ and $\Delta^{14}C$).

2 Model description and experiment

2.1 *i*LOVECLIM

The *i*LOVECLIM model is a new development branch (code fork) of the LOVECLIM model in its version 1.2, as presented in Goosse et al. (2010). It is identical to the latter with respect to its base components: Atmosphere, Ocean and Vegetation (AOV). It has been modified in a number of aspects to include water oxygen isotopes (Roche, 2013) and an interactive ice-sheet model (Roche et al., 2013a). The general goal of the new developments within *i*LOVECLIM is to include the suite of processes needed for climate simulations on the Milankovic timescale. We summarize in the following the main characteristics of the AOV components as described in Roche et al. (2007); Goosse et al. (2010). The following paragraph is taken from Roche et al. (2013b)

"The atmospheric component ECBilt was developed at the Dutch Royal Meteorological Institute (KNMI) (Opsteegh et al., 1998). Its dynamical core is based on quasi-geostrophic approximation with additional ageostrophic terms added to improve the representation of the Hadley cell dynamics. It is run on a spectral grid with a T21 truncation ($\simeq 5.6°$ in latitude/longitude in the physical space). ECBilt has three vertical layers at 800, 500 and 200 hPa. Only the first layer contains humidity as a prognostic variable. The time step of integration of ECBilt is 4 hours. The oceanic component (CLIO) is a 3-D oceanic general circulation model (Goosse and Fichefet, 1999) based on the Navier-Stokes equations. It is discretized on an Arakawa B-grid at approximately $3°\times3°$ resolution. The vertical discretization follows a "z-coordinate" on 20 levels. It has a free surface that allows the use of real freshwater fluxes, a parameterisation of downslope currents (Campin and Goosse, 1999) and a realistic bathymetry. CLIO includes a dynamical-thermodynamical sea-ice component that is an updated version of Fichefet and Morales Maqueda (1997, 1999). The dynamic land vegetation model (VECODE) was specifically designed for long-term computation and coupling to coarse resolution models (Brovkin et al., 1997). VECODE consists of three sub-models: (1) a model of vegetation structure (bioclimatic classification) calculates plant functional type (PFT) fractions in equilibrium with climate; (2) a biogeochemical model computes net primary

productivity (NPP), allocation of NPP, and carbon pool dynamics (leaves, trunks, litter and soil carbon pools); (3) a vegetation dynamics model. The latter computes two Plant Functional Types (PFT : trees & grass) and a dummy type (bare soil). The vegetation model is resolved on the atmospheric grid (hence at T21 resolution) and allows fractional allocation of PFTs in the same grid cell to account for the small spatial scale needed by vegetation. The different modules exchange heat, stress and water."

For the sake of clarity, it shall be reminded that the carbon cycle model described here do not have any relationship with the LOCH model as described in Goosse et al. (2010).

2.2 Carbon cycle model

2.2.1 Carbon cycle in the ocean

The ocean carbon cycle model is originally based on the NPZD ecosystem model described in Six and Maier-Reimer (1996) (Figure 1). It is the same model as the one included in the CLIMBER-2 model of intermediate complexity (Brovkin et al., 2002a,b, 2007) using the same parameter values, except for the remineralisation profile and the atmospheric ^{14}C, which are described below.

The carbon cycle is divided into an inorganic and an organic parts. The inorganic carbon is simulated as Dissolved Inorganic Carbon (DIC) and alkalinity (ALK). Both tracers are advected and mixed in the ocean by the advection-diffusion scheme of *i*LOVECLIM. As in Brovkin et al. (2002a), the flux of carbon at the air-sea surface is computed from the difference between the partial pressure of CO_2 in the atmosphere and ocean (with a gas exchange coefficient of 0.06 mol m^{-2} yr^{-1}). The sea surface pCO_2 is computed from temperature, salinity, DIC, and ALK following Millero (1995). The O_2 concentration is prescribed to saturation in the surface cell of the ocean.

The organic carbon pool includes 6 additional tracers on top of inorganic carbon pool, O_2 and the nutrients (phosphate and nitrate, which is diagnostically deduced from phosphate by the Redfield ratio): phytoplankton, zooplankton, Dissolved Organic Carbon (DOC), slow Dissolved Organic Carbon (DOCs), Particulate Organic Carbon (POC) and calcium carbonate ($CaCO_3$). The phytoplankton synthesizes carbon using the light and nutrients available in the first 100 meters of the ocean (euphotic zone). It then either dies and sinks or is grazed by zooplankton. Part of the plankton is remineralized to DIC, while part of it is exuded to DOC (and DOCs) and the rest is allocated to POC. The $CaCO_3$ production is linearly dependent on the organic carbon production with a fixed coefficient. Both POC and $CaCO_3$ are heavy enough to sink and are instantly remineralized and dissolved at depth. All POC and $CaCO_3$ are remineralised

and dissolved in the water column and there is no riverine input. The remineralization profile follows an exponential law as in Brovkin et al. (2002a), but this profile has been slightly modified to have less remineralization in the upper levels and more below (Figure 2). All the tracers (except for the particulate pools $CaCO_3$ and POC) are also transported by the advection-diffusion scheme of *i*LOVECLIM.

2.2.2 Carbon cycle in the terrestrial biosphere

The VECODE terrestrial biosphere model (Brovkin et al., 1997) was already included in *i*LOVECLIM (Goosse et al., 2010). The model simulates two types of plants: trees and grass, as well as desert. The plants are divided into 4 compartments that exchange carbon: leaves, wood, litter and soil. Photosynthesis depends on the local climate (precipitation and temperature) and on the atmospheric CO_2 (CO_2 fertilization). We have added the isotopes of carbon to this pre-existing version of VECODE in every carbon compartment as was done in CLIMBER-2.

2.2.3 Carbon isotopes

Following the original CLIMBER-2 version of the carbon cycle model (Brovkin et al., 2002a,b, 2007), the carbon isotopes ^{13}C and ^{14}C are simulated in the ocean and terrestrial biosphere. The ^{13}C is modeled as in Brovkin et al. (2007), while the numerical code has been modified for the ^{14}C which is now interactively dependent on cosmogenic production and carbon cycling in the atmosphere instead of having a fixed atmospheric value (Mariotti et al., 2013).

The ^{13}C simulated in the model is then used as the ratio of ^{13}C on ^{12}C to compare to the δ^{13}C data from sediment cores. The δ^{13}C is defined as follows:

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

$$\text{with } R = \frac{^{13}C}{^{12}C} \tag{2}$$

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

The ^{13}C distribution in the ocean depends on the air-sea exchange, the transport by the ocean circulation (by the advection-diffusion scheme), and the marine biology fractionation. In the terrestrial biosphere, it only depends on the exchange with the atmosphere and the biological fractionation. Indeed, both the marine and terrestrial organisms preferentially use the lighter ^{12}C over ^{13}C during photosynthesis, which tends to increase the δ^{13}C in the surrounding environment. When the remineralization occurs, the ^{12}C

rich carbon is released, which decreases the δ^{13}C in the atmosphere or ocean.

In the model, in contrast to the ^{13}C, the simulated ^{14}C is not subject to any isotopic fractionation (neither biological nor through air-sea exchanges). This formulation allows to compare directly with observations and reconstruction data from the sediment cores that are expressed in Δ^{14}C without performing a fractionation correction. The content of ^{14}C in a reservoir reflects the time since when this reservoir has been in direct contact with the atmosphere. Thus, ocean Δ^{14}C gives a good estimate of the age of water masses, which provides useful indications on ocean circulation pathways. This is particularly interesting in paleocenanography in order to reconstruct past ocean circulation changes. Moreover, the ^{14}C representation in the model can take into account temporal changes in atmospheric Δ^{14}C, which has been the case for example during the historical bomb period or the last deglaciation characterized by changes in the production rate. This aspect of the ^{14}C representation will thus be particularly useful on future paleo-simulations.

2.3 Reference simulation

The model is run under control boundary conditions set to the pre-industrial values for the orbital parameters, ice sheet reconstruction and atmospheric gas concentrations (CO_2=280ppm, CH_4=760ppb and $N_2$0=270ppb). There are indeed two different CO_2 variables in the model: the CO_2 used for the radiative code and set to 280 ppm, and the one computed by the carbon model. The CO_2 used for the radiative code is set to 280 ppm for simplicity and to make sure that the climate is correctly simulated by avoiding feedbacks arising from the wrong CO_2 computed by the carbon cycle model. For the reference simulation, as the CO_2 concentration simulated by the model is close to 280 ppm it is possible to set the radiative CO_2 equal to the CO_2 computed in the carbon cycle module, but it would be important to keep them separate for other boundary conditions such as the Last Glacial Maximum as long as the computed CO_2 concentration is not equal to the data value of the period studied. Hence the two variables are considered separately in this version of the model, but they could be the same value in future studies. The cosmogenic production of ^{14}C is set to 2.19 atom ^{14}C/cm^2/s, which is in the preindustrial data errorbar (Masarik and Beer, 2009). This production flux is then integrated over the Earth surface and added to the ^{14}C concentration of the atmosphere box. The simulation starts from an equilibrium run for the climate, and uniform distribution of tracers in the ocean. The total amount of carbon has been adjusted in iterative runs to reach a value close to the the pre-industrial CO_2 level of 280 ppm in the atmosphere. The simulation was run until it reached an equilibrium for deep ocean variables ($\simeq 10,000$ years), and the mean of the last 100 years is used to compare the results with existing data.

2.4 Data

We compare the model results with existing observations and CMIP5 models simulations. We use temperature, salinity, phosphate and oxygen data from the World ocean atlas 2009 (Locarnini et al., 2010; Antonov et al., 2010; Garcia et al., 2010a,b). For the dissolved inorganic carbon, alkalinity and Δ^{14}C we compare results with data from GLODAP (Key et al., 2004). The pCO$_2$ data come from Takahashi et al. (2009) and the δ^{13}C data from Schmittner et al. (2013).

The global climate models considered from the Coupled Model Intercomparison Phase 5 (CMIP5) are CESM1-BGC, CMCC-CESM, GFDL-ESM2G, GFDL-ESM2M,HadGEM2-ES, IPSL-CM5A-LR, IPSL-CM5A-MR, MPI-ESM-LR, MPI-ESM-MR and NorESM1-ME. For each variable, the models for which the data were available are listed in Table 1. For more detailed information on the models see Bopp et al. (2013). The results are averaged over the period 1890-1899 from the "historical" simulation. The end of the 19th century is chosen because it is more similar to the *i*LOVECLIM simulation. It can be noted that very similar results for the ocean interior are obtained when considering the end of the 20th century instead, due to the long timescale of the deep ocean (a few hundred years).

Table 1. CMIP5 models considered for each variable ("x" for yes, and "-" for no).

Model	DIC	ALK	O$_2$
CESM1-BGC	x	x	x
CMCC-CESM	x	x	x
GFDL-ESM2G	x	x	x
GFDL-ESM2M	x	x	x
HadGEM2-ES	-	-	x
IPSL-CM5A-LR	x	x	x
IPSL-CM5A-MR	x	x	x
MPI-ESM-LR	x	-	-
MPI-ESM-MR	-	x	x
NorESM1-ME	x	x	x

3 Results

After equilibrium, the atmospheric CO_2 concentration is 287 ppm, the atmospheric δ^{13}C value -6.4 permil and the atmospheric Δ^{14}C value 1.5 permil, close to the pre-industrial values of respectively 279 ppm, -6.4 permil (Elsig et al., 2009) and 0 permil (Reimer et al., 2009). In the case of Δ^{14}C , the simulated -1.5 permil is a particularly

good estimate of the observed 0 permil, because uncertainty on preindustrial $\Delta^{14}C$ values is on the order of 10 permil (Reimer et al., 2009). The ocean contains 39019 GtC and the terrestrial biosphere 2142 GtC.

The total vegetation cover simulated by the model (Figure 3) is in agreement with the one from another version of LOVECLIM (Goose et al., 2010, Figure 14). Likewise, it is similar to the data but with an overestimation of the cover in the tropics because of too much precipitation. In terms of carbon content, *i*LOVECLIM simulates low carbon contents in the regions of low vegetation cover, and particularly high carbon contents in the southern and eastern parts of North America, the north-eastern part of South America, the south-eastern part of Africa and on the maritime continent. This results in 2142 GtC globally, corresponding to 863 GtC for vegetation and 1279 GtC for soils (and litter). This is in the range of other model estimates which vary between around 320 and 930 GtC for vegetation and between around 500 and 3100 GtC for soils (Anav et al., 2013), as well as close to data estimates although with an overestimation of vegetation carbon content and underestimation of soil carbon content (respectively 450 GtC to 650 GtC for vegetation, (Prentice et al., 2013) and 1500 to 2400 GtC for litter and soils (Batjes, 1996)).

Because the objective of this coupling is to study the climate and carbon cycle on timescale of more than thousands of years, and because the terrestrial biosphere has already been studied (apart from the isotopes) (Goosse et al., 2010), we focus mainly on the distribution of the tracers in the ocean, both at the surface and in the interior. We also compare the carbon isotopes results with data as they constitute an important constraint for past climates.

3.1 Ocean dynamics

The ocean dynamics, which depend on temperature and salinity gradients, play an important role for the carbon cycle because they partly determine the distribution of the tracers that are transported. The *i*LOVECLIM model simulates relatively well the distribution of temperature and salinity both at the surface and in the ocean interior.

In the surface, the simulated temperature field is similar to the observations (Figure 4), with higher temperatures at the low latitudes and lower at high latitudes. Some local discrepancies can be observed in the boundary currents which are not well represented in the model due to its low resolution. The salinity distribution is in agreement with the data in most places (Figure 5), except in the two bands of higher salinity in the Pacific and Indian oceans around 30°N and 30°S and in the North Western part of the Indian Ocean where the simulated salinity is too low compared to observations.

In the ocean interior, the major oceanic water masses display similar features as in the data (Figure 6 and 7). The thermocline is well represented in both the Atlantic and Pacific oceans. The Antarctic Bottom Water (AABW), which forms around Antarctica and sinks to the bottom of the ocean, is characterized by very cold temperature and low salinity in the model as in the observations. The North Atlantic Deep Water (NADW) which forms in the North Atlantic high latitudes, has relatively warmer and saltier water, in agreement with data. The low salinity tongue of the Antarctic Intermediate Water (AAIW), which spreads northward at intermediate depth of around 1000m in the Southern Hemisphere, is also well represented in the model. In the Pacific, the penetration of the North Pacific Intermediate Water (NPIW) with low salinity is similar to the observational data. However two main discrepancies can be seen. In the Southern Ocean, AABW is too cold, so that most of the bottom ocean is slightly too cold compared to the data. In the North Atlantic the water that sinks with NADW is too salty because the surface water is also slightly too salty (Figure 5).

The simulated streamfunction (Figure 8) is in the range of other models, with a maximum AMOC value of 21 Sv, compared to values between 14 and 31 Sv for CMIP5 models (Weaver et al., 2012). Comparing to observation of the AMOC strength (e.g. Srokosz et al. (2012) and references therein), we find an upper limb transport at 26° N of about 15 Sv, lower that the 17 to 22 Sv estimates (Kanzow et al., 2010; Srokosz et al., 2012) from direct measurements. At 16° N, we obtain a lower limb of about 19 Sv, in good agreement with observations (Send et al., 2011; Srokosz et al., 2012) that infer a transport of 17 ± 3.5 Sv.

3.2 Nutrients and oxygen

The distribution of nutrients depends on the transport by the diffusion-advection scheme of the ocean model, their use by marine biota (net productivity) and remineralization at depth. In the euphotic zone in the first 100m below the surface, nutrients are consumed by phytoplankton during photosynthesis, while oxygen is produced. There is thus less nutrients in the surface than in the deep ocean, which can be seen on simulated phosphate concentrations, in agreement with data (Figures 9 and 10). The surface distribution of simulated phosphates tends to underestimate the intensity of boundary currents and upwellings as already seen in the surface temperature field, nonetheless the low-to-high latitudes gradient observed in data is well represented (Figure 9). In the surface the oxygen is set to the saturation level (Figures 11 and 12). The simulated surface distribution of oxygen tends to be underestimated in the North West

Atlantic and in the Benguela upwelling, as well as in parts of the Southern Ocean (Figure 11) but this is due to the too warm temperatures in these areas compared to data (Figure 4), which decreases the solubility of atmospheric oxygen in the surface water. In the North Atlantic, this error then propagates in the interior resulting in too low oxygen values in the deep North Atlantic. In the ocean interior, the remineralization of plankton consumes oxygen and releases nutrients. This explains the minimum of oxygen and maximum of nutrients around 500-1000 meters which is relatively well represented in the model compared to data (Figures 10 and 12).

The differences between the Atlantic and Pacific basins are also well represented. In the North Atlantic, the NADW sinks with lower phosphate values (Figure 10 a and b) and higher O_2 values (Figure 12 a and b) from the surface where the waters are enriched in O_2 and where nutrients are consumed for the photosynthesis. The O_2 values in the ocean interior where NADW penetrates are slightly too small in the model because the surface values are too low. In the Pacific, the water is progressively enriched in PO_4 (Figure 10 c and d) while it becomes depleted in O_2 (Figure 12 c and d) during its transport from the south to the north, because of the constant remineralisation which enriches the water in PO_4 and uses O_2.

3.3 Carbon

The simulated distribution of dissolved inorganic carbon and alkalinity is in relative agreement with the data in the oceans. At the surface, DIC is higher at high latitudes and lower at low latitudes like in the data (Figure 13), although the DIC levels in the tropics are slightly too low compared to the data. The alkalinity values are similar to the data, but with some small differences especially in the Atlantic where the data display two zones of higher values in the middle of the tropical gyres which are not very well represented by the model (Figure 14). This could be due to the dissolution profile of $CaCO_3$ which is a function of depth, as for POC but with different values, and could be improved to be more realistic.

In the ocean interior, NADW is characterized by relatively low DIC values in the model as in the data, although the model values are slightly too high (Figure 14). In the Pacific, the water becomes progressively enriched in DIC and alkalinity as it goes from the south to the north because of remineralization (Figures 15 and 16). This is well represented in the model for DIC, however the alkalinity distribution is less well represented in the model, which could be due to the simple linear relation between the production of $CaCO_3$ and the production of organic matter,

or the fixed vertical profile of remineralisation.

The regions of high and low pCO_2 are generally well represented in the model compared to the data (Figure 17). In particular, the pCO_2 values are higher around the equator, where the upwelling brings water with a high carbon content that is lost to the atmosphere, even if the model underestimates these high values. At high latitudes, especially in the North Atlantic and Arctic regions, the pCO_2 values are low where the ocean takes up carbon from the atmosphere. However, in the Southern Ocean the data indicate low values, even if they are sparse, which are not shown by the model, but the cause of this mismatch are unknown.

3.4 Carbon isotopes

During photosynthesis, the organisms preferentially use the relatively light ^{12}C over ^{13}C. This leads to higher $\delta^{13}C$ values in the surface and lower values deeper in the ocean where remineralization takes place and ^{12}C is released. This is well represented in the model (figure 18), as well as the minimum value in the subsurface equatorial Atlantic due to higher remineralization in that region. The $\delta^{13}C$ also depends on circulation, so that NADW is characterized by relatively high values and AABW by lower values, in agreement with data. In the Pacific, the water is progressively enriched in ^{12}C from remineralization from south to north, resulting in the low $\delta^{13}C$ values. However, the high $\delta^{13}C$ values in the North Atlantic do not penetrate south enough, which could be due to too much diffusion.

As opposed to simulated $\delta^{13}C$, simulated $\Delta^{14}C$ does not depend on biology effects, so it allows to separate biological and circulation effects registered by $\delta^{13}C$. The general structure of oceanic $\Delta^{14}C$ is well simulated by the model (Figure 19) and reects the penetration of water masses in the interior of the ocean: from north to south in the Atlantic Ocean and from south to north in the Pacic Ocean. The model performs well compared to other ocean general circulation models (Mariotti et al., 2013; Tschumi et al., 2011; Franke et al., 2008; Matsumoto et al., 2004), especially in the intermediate to deep Pacific Ocean. The model values seem nonetheless to decrease too rapidly following the penetration of NADW in the North Atlantic, similarly to PO_4, which could indicate that the diffusion is too strong in that region. In the Pacic, the water becomes too depleted in $\Delta^{14}C$ in the nothern part, possibly due to an underestimate of the mixing in that region.

4   Discussion

Because the main feature added to *i*LOVECLIM for the carbon cycle concerns the ocean, we only discuss the results for the oceanic variables. The terrestrial biosphere has only been slightly modified to include the carbon reservoirs, but could benefit from further improvements such as more plant functional types, as well as additional modules such as permafrost, which is work under progress (Kitover et al., 2013).

4.1   Model-data comparison

The *i*LOVECLIM model simulates most of the variables in agreement with data, especially the main characteristics of the water masses. However, a number of discrepancies exist. Some are due to errors in the simulation of surface regional features which then propagate in the ocean interior, such as the North Atlantic where the high salinity from the tropics is transported too much northward compared to the data. This could be partly due to the resolution of the model which limits the representation of small scale features. The misrepresentation of temperature has a direct impact on oxygen, for example again in the North Atlantic where the temperatures are too high, which leads to too small values of oxygen in the surface and in the ocean interior. Another source of error could come from the diffusion which seems too strong in the North Atlantic where the characteristic values of NADW for salinity, PO_4, DIC and carbon isotopes decrease too rapidly while it penetrates southward. This highlights the crucial role of a correct representation of temperature and salinity and the associated ocean circulation in setting the distribution of the biogeochemical variables. The distribution of the variables strongly depends on salinity and temperature distribution: if it is improved it should also improve the carbon cycle.

4.2   Inter-model comparison

We compare the *i*LOVECLIM results with other models using the data from the Coupled Model Intercomparison Phase 5 (CMIP5). We focus on 3 key variables (dissolved inorganic carbon, alkalinity and oxygen) using the average over years 1890-1899 of the "historical" simulation (see section 2.4). The data are zonally averaged for the Atlantic and Pacific basins (including the Southern Ocean). Note that the simulations that are compared are not exactly the same: the *i*LOVECLIM simulation is a long simulation of a few thousand years under pre-industrial conditions, whereas the CMIP5 simulations are run under evolving boundary conditions of the historical period since 1850 starting from spin-up simulations of a few hundred to one thousand years. Additionally, the spatial resolution is higher in the CMIP5 models which are fully coupled GCMs. Nevertheless, we show here that the skill scores of *i*LOVECLIM are similar to those of more complex Earth System Models used in CMIP5.

For most variables, *i*LOVECLIM is in the range of other models performance. For DIC the models that statistically perform best in both the Atlantic and Pacific are the IPSL-CM5A-LR and IPSL-CM5A-MR models (Figure 20). *i*LOVECLIM is less accurate than the IPSL models, but still reproduces most of the pattern and gives better results than other models such as NorESM1-ME, CMCC-CESM, GFDL-ESM2G or MPI-ESM-LR in terms of correlation and root mean square error. For alkalinity, most models simulate poorly the distribution especially in the Atlantic basin, where *i*LOVECLIM is performing particularly poorly (Figure 21). In the Pacific, which represents a larger volume, the models yields better results and so does *i*LOVECLIM which lies in the middle of the ensemble. This highlights the need of better understanding the processes responsible for the change of alkalinity to improve its distribution in models. For the oxygen, *i*LOVECLIM lies behind most models in the Atlantic but is in the middle of the range in the Pacific (Figure 22). In the Atlantic basin, this is partly due to the representation of the high O_2 values penetrating in the North Atlantic with NADW that is not well reproduced in *i*LOVECLIM because the O_2 values are too low at the surface. Future work will focus on understanding the causes of the mismatch to improve the O_2 distribution. In the Pacific basin *i*LOVECLIM has a good correlation at around 0.8 like most models. This is not as good as a few models with correlations higher than 0.9 such as CESM1-BGC, MPI-ESM-MR and MPI-ESM-LR, but relatively good and better than NorESM1-ME with a correlation of only 0.5.

4.3   Future developments

Overall, *i*LOVECLIM does a relatively good job compared to the data and other models and usually lies in the middle of the CMIP5 range. This is a good performance given that *i*LOVECLIM is an EMIC and has a less complex and comprehensive representation of the different processes than the CMIP5 GCMs. The GCMs usually simulate better the ocean circulation which yields better distribution of the geochemical variables. There are however a few points that need to be improved in *i*LOVECLIM namely the O_2 representation in the Atlantic and the alkalinity distribution (like in all other models).

Some limitations arise from the simplicity of the NPZD model which does not include iron nor silicate. This could be added in future work. The air-sea flux of oxygen has not yet been parameterised depending on the difference between the atmosphere and surface water values and the wind, but this will be explored in future studies. It could improve the regional distribution of oxygen values, and would also modify the temporal evolution of oxygen values

in transient simulations. Work has been done in other models showing the importance of remineralisation on the carbon cycle Schneider et al. (2008); Kwon et al. (2009). The profile, which depends on depth, is currently fixed, but the effect of changing the values depending on the temperature or other variables should be evaluated. The production and dissolution of $CaCO_3$ could also be improved, which would yield better results for the alkalinity distribution. In particular, $CaCO_3$ production is currently proportionnal to the production of organic matter, which could be modified, and the vertical dissolution profile is fixed, which could be changed to take into account the saturation state.

5 Conclusions

We have described the implementation of a carbon cycle module in the *i*LOVECLIM model, including the carbon isotopes ^{13}C and ^{14}C. Comparison with modern data show that the model performs well for the main carbon cycle variables, and reproduces the most important features of the different water masses. In particular, the good representation of the ^{13}C and ^{14}C in the ocean interior paves the way for past studies for which they represent most of the available data. Therefore the *i*LOVECLIM model with the carbon cycle is well suited for long term simulations of a few thousand years in the past but also in the future. Some improvements will be considered in future work, such as the inclusion of iron and silicate, a better parameterization of the O_2 air-sea exchange with wind and better parameterization of the remineralization and dissolution profiles. Finally, a sediment model remains to be coupled to include all relevant oceanic components of the carbon cycle on timescales of a few thousand years.

Code Availability

The *i*LOVECLIM source code is based on the LOVE-CLIM model version 1.2 whose code is accessible at http://www.elic.ucl.ac.be/modx/elic/index.php?id=289 The developments on the *i*LOVECLIM source code are hosted at https://forge.ipsl.jussieu.fr/ludus but are not publicly available due copyright restrictions. Access can be granted on demand by request to D. M. Roche (didier.roche@lsce.ipsl.fr).

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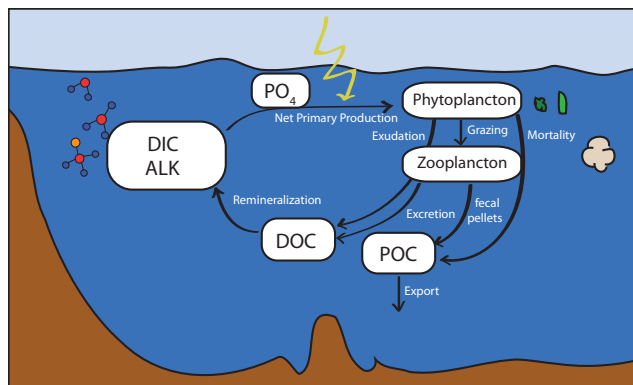
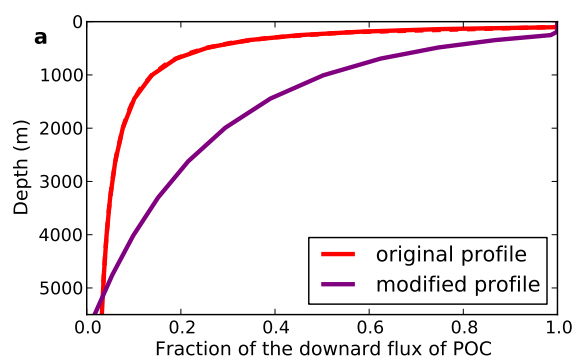**Fig. 1.** Schematic of the ocean carbon cycle in *i*LOVECLIM.

Fig. 2. Fraction of Particulate Organic Carbon (POC) which is transported downward at each level (the fraction not transported is remineralized).
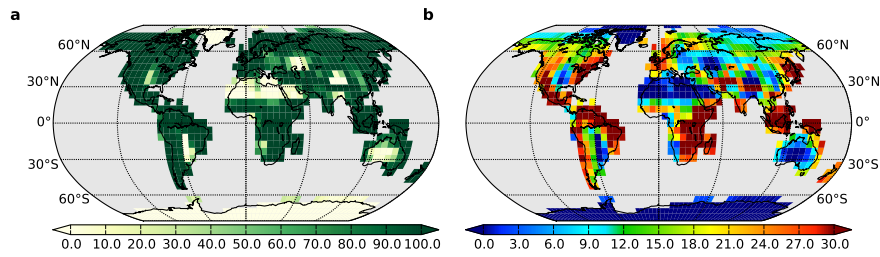
Fig. 3. Distribution of (a) total vegetation cover (%) and (b) total carbon content (kgC/m^2).
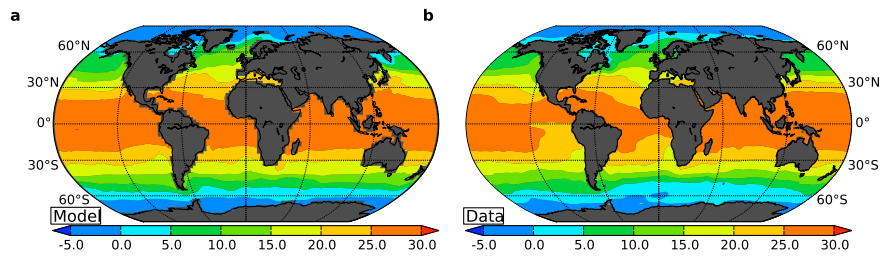

Fig. 4. Temperature distribution ($^{\circ}$C) at the ocean surface. Data from the World ocean Atlas 2009 (Locarnini et al., 2010).
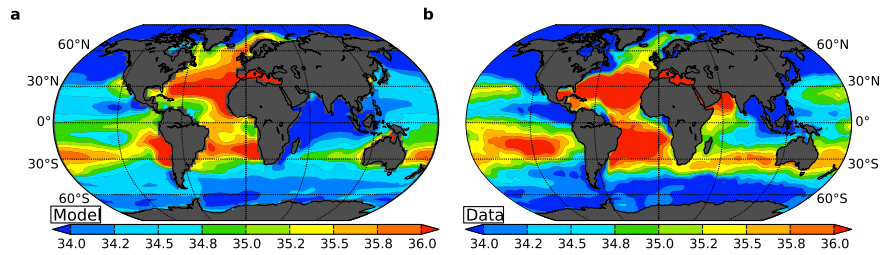

Fig. 5. Salinity distribution (psu) at the ocean surface. Data from the World ocean Atlas 2009 (Antonov et al., 2010).
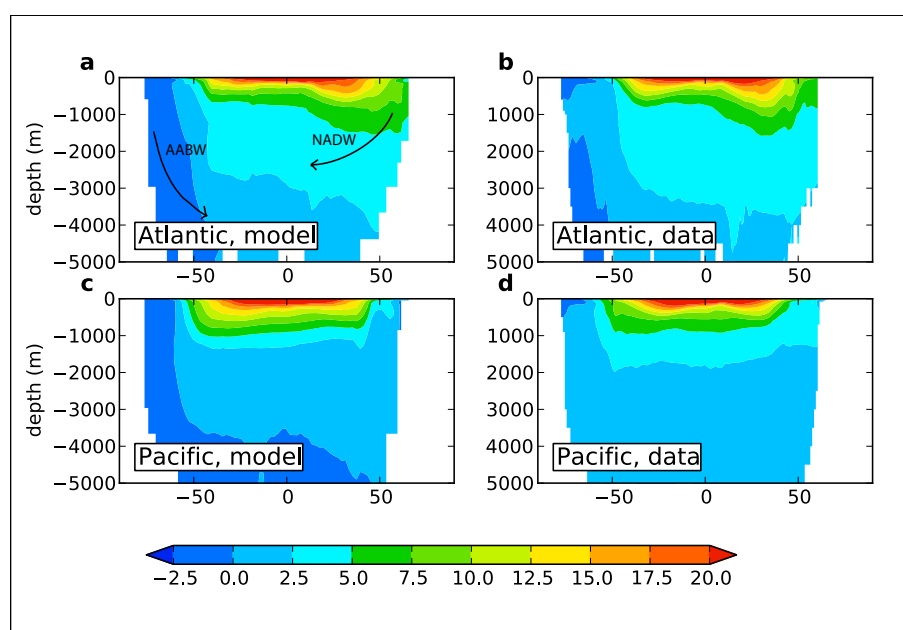
Fig. 6. Zonal average of the temperature distribution (°C) in the ocean. Data from the World ocean Atlas 2009 (Locarnini et al., 2010). North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW) are indicated on panel a.
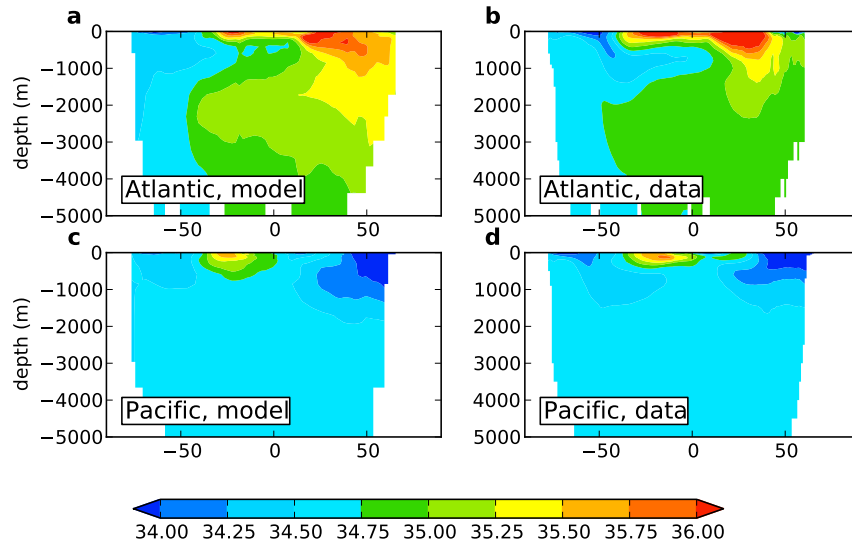
Fig. 7. Zonal average of the salinity distribution (psu) in the ocean. Data from the World ocean Atlas 2009 (Antonov et al., 2010).
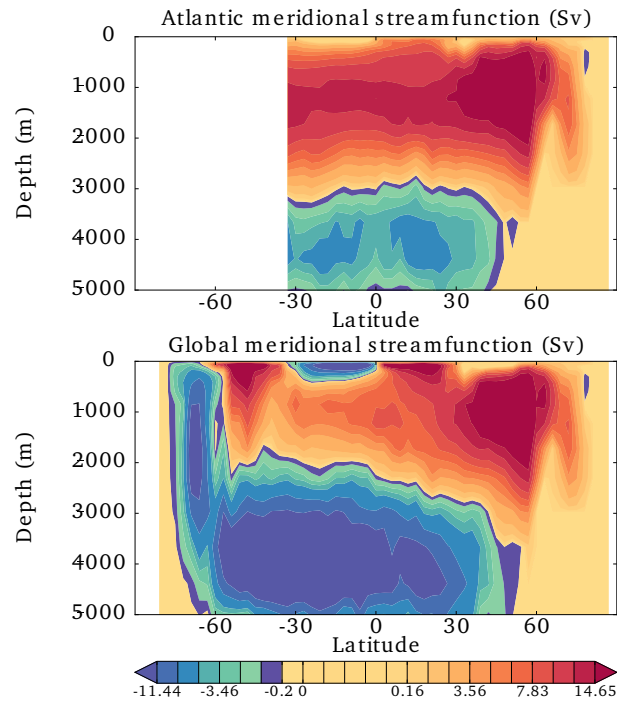
Fig. 8. Atlantic and global meridional streamfunction (Sv).
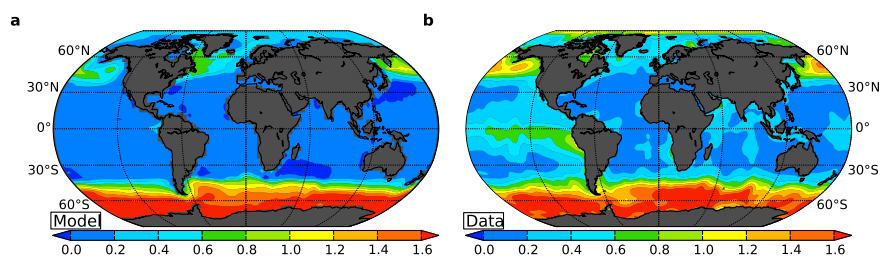
Fig. 9. Phosphate distribution at the ocean surface (μmol/kg). Data from the World ocean Atlas 2009 (Garcia et al., 2010b).
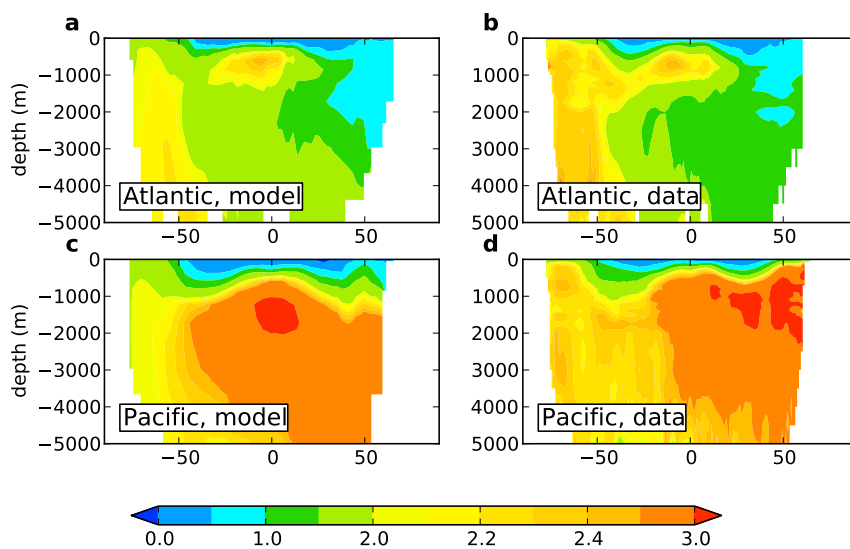
Fig. 10. Zonal average of the phosphate distribution in the ocean (μmol/kg). Data from the World ocean Atlas 2009 (Garcia et al., 2010b).
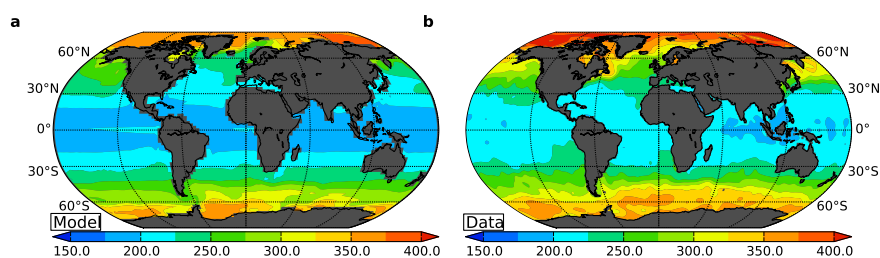
Fig. 11. Oxygen distribution at the ocean surface (μmol/kg). Data from the World ocean Atlas 2009 (Garcia et al., 2010a).
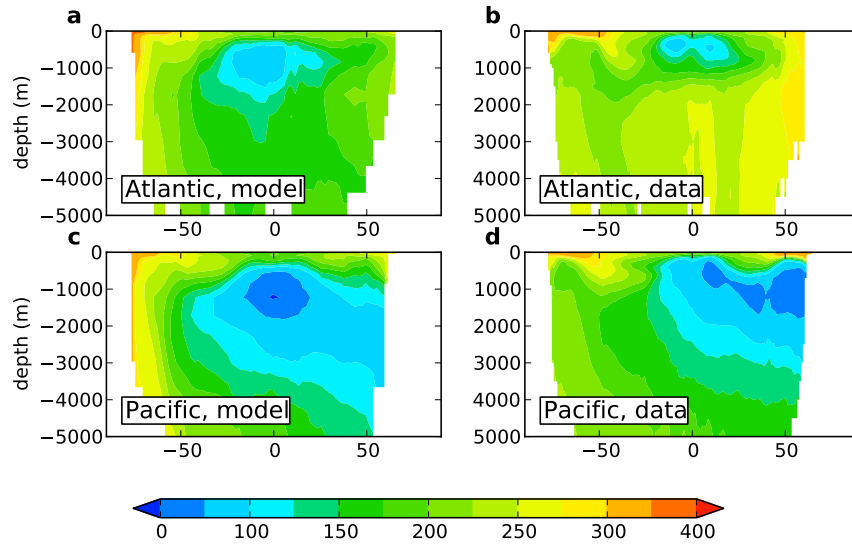
Fig. 12. Zonal average of the oxygen distribution in the ocean (μmol/kg). Data from the World ocean Atlas 2009 (Garcia et al., 2010a).
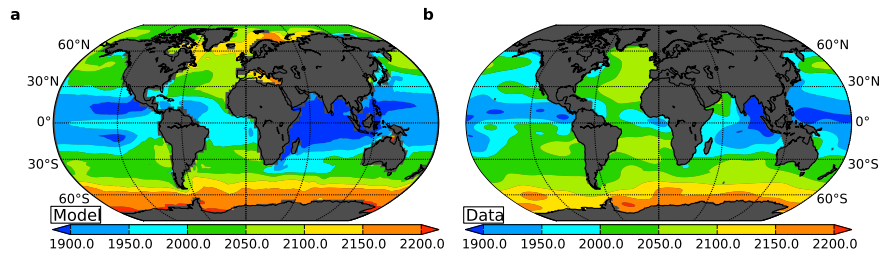
Fig. 13. Dissolved Inorganic Carbon (DIC) distribution at the ocean surface (μmol/kg). Data from GLODAP (Key et al., 2004).
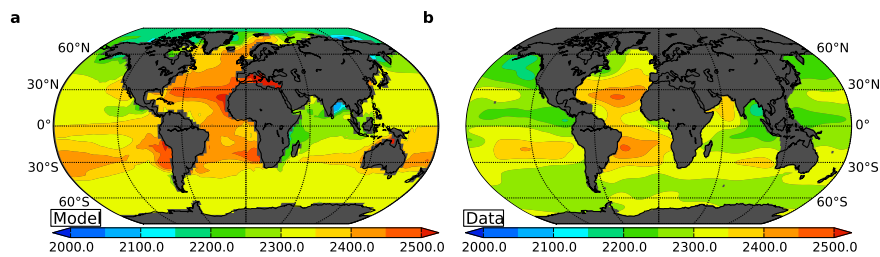
Fig. 14. Alkalinity (ALK) distribution at the ocean surface (μmol/kg). Data from GLODAP (Key et al., 2004).
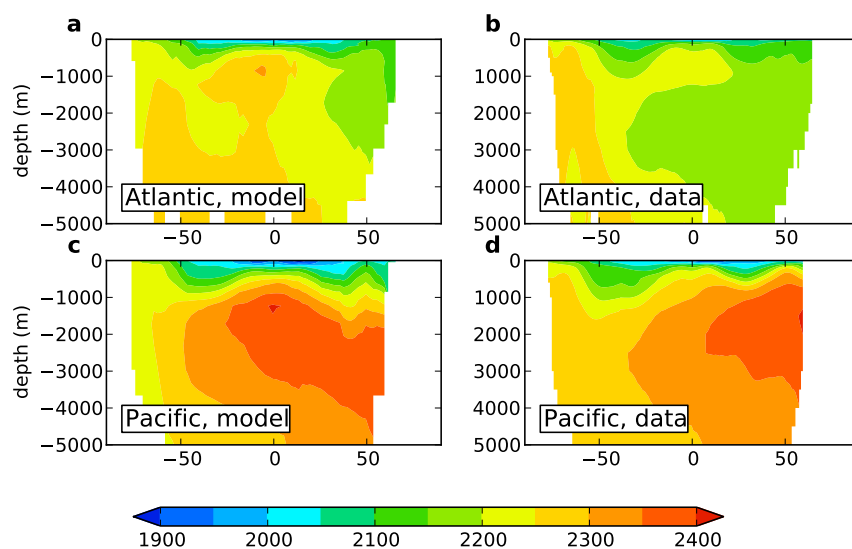
Fig. 15. Zonal average of the dissolved Inorganic Carbon (DIC) distribution in the ocean (μmol/kg). Data from GLODAP (Key et al., 2004).
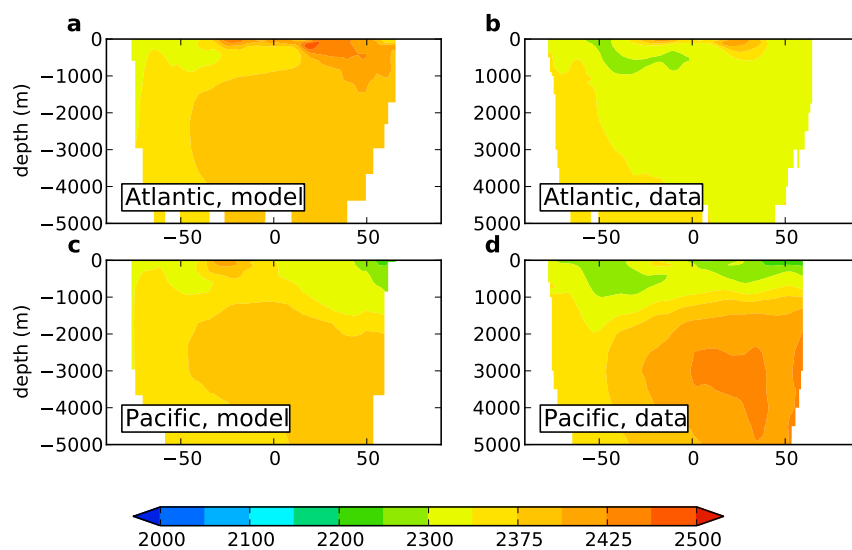
Fig. 16. Zonal average of the alkalinity (ALK) distribution in the ocean (μmol/kg). Data from GLODAP (Key et al., 2004).
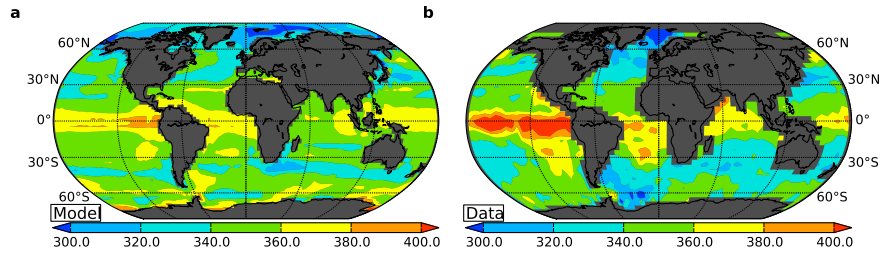
Fig. 17. pCO2 distribution at the ocean surface (μatm). Data from Takahashi et al. (2009).
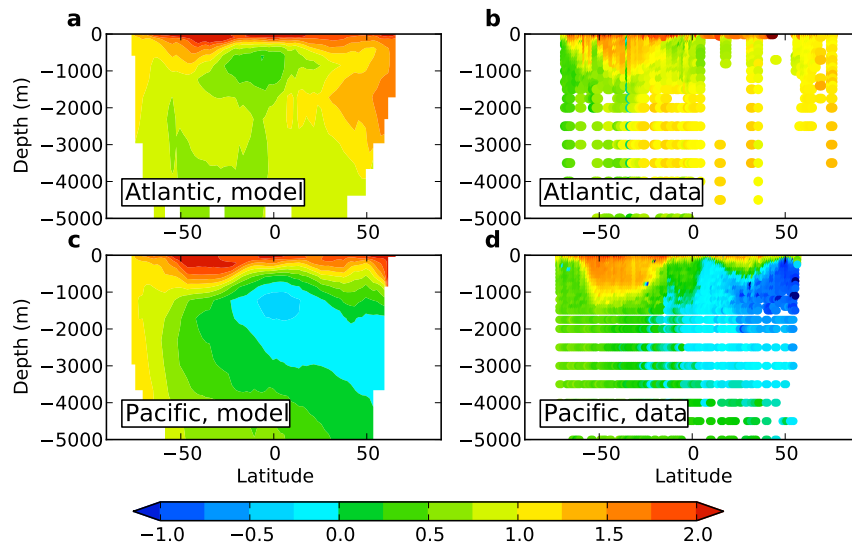

Fig. 18. Zonal average of the δ^{13}C distribution in the ocean (permil). Data from Schmittner et al. (2013).
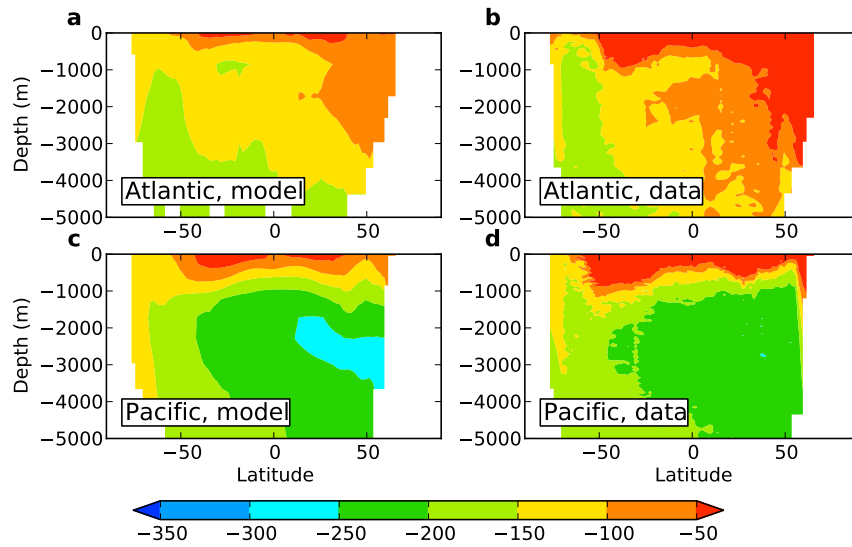
Fig. 19. Zonal average of the Δ^{14}C distribution in the ocean (permil). Data from GLODAP (Key et al., 2004).
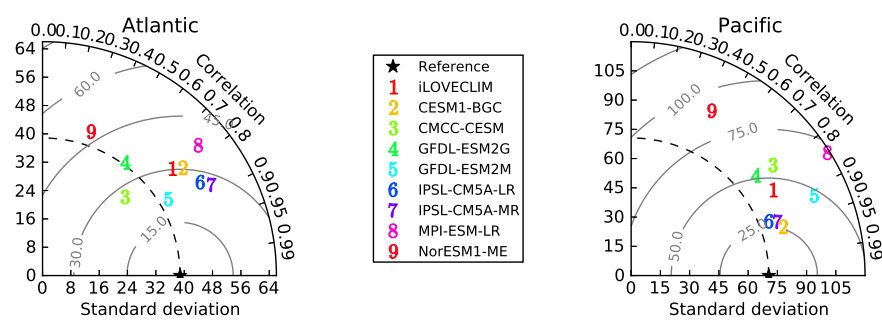

Fig. 20. Comparison of the latitude-depth pattern of zonally averaged Dissolved Inorganic Carbon in the Atlantic and Pacific basins (Taylor diagrams). Data from GLODAP (Key et al., 2004).
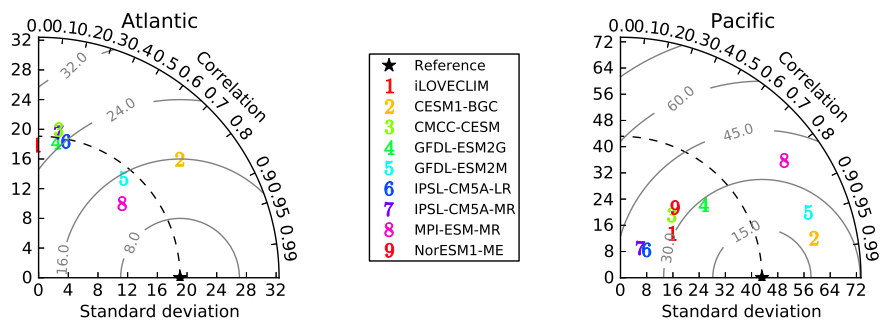
ALK

Fig. 21. Comparison of the latitude-depth pattern of zonally averaged alkalinity in the Atlantic and Pacific basins (Taylor diagrams). Data from GLODAP (Key et al., 2004).
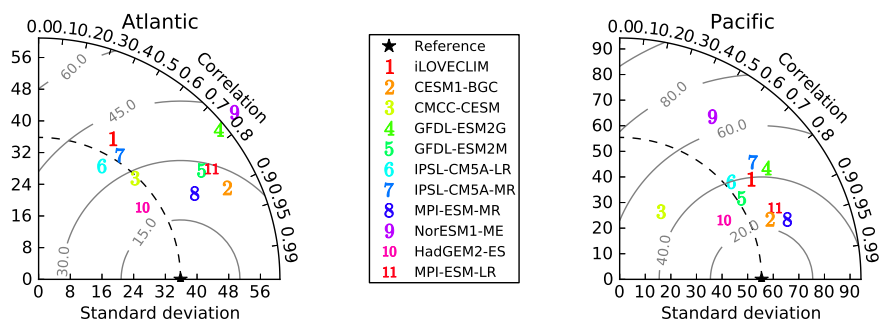
O2

Fig. 22. Comparison of the latitude-depth pattern of zonally averaged oxygen in the Atlantic and Pacific basins (Taylor diagrams). Data from the World ocean Atlas 2009 (Garcia et al., 2010a).