

Author reply to editor and reviewers' comments

Dear Luke Skinner,

We very much apologise it has taken us so long to compile a revised version of our manuscript. Partly this was due to the usual time constraints, but partly it was also due to the extensive changes we made to the manuscript in order to address the concerns of reviewer #2. In order to address these concerns we had to perform some additional experiments, generate new Figures and perform additional analyses. The latter has taken substantially more time than we anticipated.

In the present letter I include your comments to our manuscript, as well as the reviewers' comments and our original replies to these comments. These sections have the original comments in bold font and our original replies in normal font. In addition, I have added sections, marked by blue colour, where I describe how we have addressed the comments in the revised manuscript. Pages and line numbers indicated refer to the version of the manuscript, where changes are not highlighted. I have also included an additional file, where changes are marked and notes indicate which reviewer comment caused these changes.

Our main changes to the manuscript are as follows:

We exchanged experiment HOL_ALL from our original submission, which had a reduced anthropogenic emission forcing in comparison to the original Kaplan et al. scenario, with a new experiment HOL_ANT, which was driven by the original Kaplan et al. emission scenario. This does not change results in any substantial way but addresses one concern by reviewer #2. We also added a new experiment HOL_MPT, where we use a minimum peat accumulation estimate in order to address some of the uncertainty in peat carbon accumulation, as asked by reviewer #1.

In addition we introduced a new set of Figures, Figures 3, 7, and 11, where we disaggregate the net fluxes of C to the atmosphere into the geological, land, and ocean components. These new Figures then required substantial changes to the results section, which is mostly rewritten.

We have made substantial changes to the discussion section, where we now discuss uncertainties in peat accumulation, land carbon cycle, and sea level forcing in more detail than in our original submission.

Finally, many smaller changes were made to the manuscript in order to address reviewer concerns, as detailed below, including the removal of Figure 1, as suggested by Reviewer 1. In rewriting the text we also made a number of wording changes, which improve readability, especially in the experiment description section.

We thank you and the reviewers very much for your comments, and we hope you will agree that the manuscript is now substantially improved.

All the best,

Thomas Kleinen

Editor comment

Dear Thomas Kleinen,

Thank you for your responses to the reviewer comments on your manuscript. Both reviewers are clearly supportive of publication, though both also raise issues that they feel must be addressed prior to acceptance for publication.

I therefore recommend that you prepare a revised version of your manuscript, at your earliest convenience, that judiciously takes into consideration the points raised by both reviewers. I have indicated 'major revisions' only so as to be able to request a further review of your revised manuscript should this be necessary.

From my side, I would draw your attention in particular to the need to frame more clearly the motivation for the study (and the choice of interglacials), such that you are also able to spell out more clearly exactly what can be learned from your model results and how these results address the stated motivation of the study. I would urge you also to reconsider the extent to which your study strictly does show "...how the CO₂ evolution during the Holocene and two recent interglacials can be explained consistently using an identical model setup" (as stated in the abstract) and that "...trends in interglacial atmospheric CO₂ can be reproduced by a climate model with identical forcing" (as stated in the conclusions). More specifically I would propose that the abstract and conclusions might be softened somewhat to state more accurately (in my view) that the simulation of the trends in atmospheric CO₂ across all of these interglacials is significantly *improved* by the inclusion of two key 'slow' carbon cycle processes; coral reef growth and peat accumulation. I think it is less clear that CO₂ has truly been *reproduced* in each case, and always for the right reasons (e.g. given mismatches in simulated and observed d¹³C, or the lack- or uncertainty of data constraints to test simulated carbonate accumulation for example).

We have taken up your suggestion and softened both abstract and conclusions. They now state that model results are substantially improved by the inclusion of the slow carbon cycle processes. We also motivate our choice of interglacials (page 3, lines 24-30).

I look forward to receiving your revised manuscript. It would be helpful if you accompanied your revision with a succinct description of the changes included, or else a highlight the changes in the revised manuscript itself.

Yours sincerely,

Luke Skinner

Author reply to comments by anonymous reviewer #1

We very much thank the reviewer for taking the time to review our manuscript. We aim to incorporate all of the reviewer's comments in the final manuscript, this will lead to a substantial improvement over the original submission. For reader convenience we have included the reviewer's comments in full in this reply, marking them by bold font.

The goal of the study is to understand the atmospheric CO₂ and d¹³CO₂ evolution during three interglacials: the Holocene, the Eemian and MIS11 using the CLIMBER2 model. The study focuses on the role of shallow water carbonate sedimentation and peat accumulation. For that purpose CLIMBER2 is coupled to the land model LPJ and shallow water carbonate sedimentation is estimated from a simple formulation.

The roles of CaCO₃ sedimentation and changes in land carbon on atmospheric CO₂ and d¹³CO₂ have been previously studied for the Holocene (including by the authors in Kleinen et al. 2010). However, changes in atmospheric CO₂ and d¹³CO₂ during the Eemian and MIS11 have received little (if any) attention. It is an interesting paper, worth publishing in *Climate of the Past*. Please find a few comments below.

1) Since it has been more studied, estimates of CaCO₃ sedimentation and peat accumulation as well as pCO₂ and d¹³CO₂ measurements are more accurate for the Holocene. The Holocene simulation could work as a validation of the modelling approach used here. More information could thus be taken out of that simulation to inform on the other 2.

This had been our aim, but in the light of the reviews it has become clear that we need to further extend the discussion of the Holocene results. Therefore we will take this up and extend the discussion of the Holocene results, especially extending the discussion of marine C changes.

[We have extended the results and discussion sections, especially including discussion of the fluxes leading to the changes in CO₂, which should clarify this issue.](#)

The simulated changes in peat accumulation for the Holocene are in line with previous studies (e.g. Yu et al. 2010, Spahni et al. 2013). But I wonder what are the uncertainties associated with the peat accumulation estimates and with land carbon changes in general. The authors discuss the mismatch between the simulated d¹³CO₂ compared to the ice core measurement during the late Holocene. The mismatch almost reaches 0.2 permil at 0.5 ka B.P. Elsig et al. 2009 estimated the land carbon change occurring during the Holocene to match their d¹³CO₂ record. They suggest a land carbon uptake of 290GtC during the early Holocene (10-6 ka B.P.), followed by a 36GtC release. The simulated changes in CaCO₃ sedimentation for the Holocene are quite high. Much higher than Vecsei and Berger 2004, but roughly in line with other studies (e.g. Kleypas 1997, Ryan et al. 2001). So the mismatch between simulated and observed d¹³CO₂ during the late Holocene could be explained by an overestimated Holocene peat accumulation, or more broadly an overestimated land carbon uptake coupled with an overestimated CaCO₃ sedimentation (because pCO₂ follows the observation). The mismatch starts at about 4.5 ka B.P. and as also stated by the authors, I doubt it is due to anthropogenic land carbon changes. The authors briefly mention permafrost. Would permafrost thawing occur that late in the interglacial? It might be interesting to add a few sentences on the possible role of permafrost. The same could be true for the other time periods. For example, simulated d¹³CO₂ between ~126-122 ka B.P. is significantly lower than observations.

A discussion of uncertainties associated with land carbon changes (and peat, please see comment below) could be added in the Discussion section. Additionally, the abstract could reflect these uncertainties.

We agree that there are substantial uncertainties with regard to the data on peat accumulation. In our 2012 paper on the peat model we have also published minimum and maximum estimates of peatland areas and peat carbon accumulation. We will use the minimum estimate and derive a second calibration for the CaCO₃ model corresponding to the lower peat carbon uptake estimate. This will a) reduce the CaCO₃ sedimentation required to match the CO₂ record and b) bring down d¹³CO₂ somewhat, but judging from older model results we have available, a mismatch between d¹³CO₂ data and model results will remain. With regard to the dynamics of the permafrost carbon, it is difficult to provide quantitative estimates of its changes in the Holocene. Assuming that the permafrost extent and C storages are linked to the temperature dynamics, one could conclude that permafrost carbon increased during the late Holocene when summer

temperatures in the northern high latitudes decreased following the summer insolation decline. However, processes of thermokarst and water erosion which disturb the permafrost C storages may require much more than several thousand years for equilibration with climate change. The amount of ice in high-latitude permafrost soils formed during the last glacial cycle is large, and disturbances could possibly release glacial-aged carbon to the atmosphere.

We will extend the discussion section with regard to uncertainties in land C processes and also extend the discussion on the possible role of permafrost.

We have added the HOL_MPT experiment containing a minimum peatland area peat accumulation estimate in order to address part of the uncertainty with regard to peat accumulation. We have also added a discussion of uncertainties in peat accumulation estimates (page 17, line 27 – page 18, line 10) and a paragraph discussing uncertainties in land carbon changes (page 18, lines 11 – 19).

2) It has been suggested that Northern hemisphere summer insolation modulates peat accumulation (e.g. Yu et al. 2010). Apart from a slightly lower accumulation rate between 395 and 380 ka B.P., figures 4c, 7c and 10c display similar linear trends in peat accumulation rate for the 3 time periods (Holocene, Eemian and 1st part of MIS11), which is a bit surprising giving the fact that sea level variations (and thus most likely ice sheet evolution and NH insolation) are different for the 3 periods. What is the sensitivity of CLIMBER2-LPJ peat accumulation to NH summer insolation? Plotting NH summer insolation timeseries in figures 4, 7 and 10 could be useful.

Since they are a main part of the study, it would be nice to add some explanation on peat carbon changes in sections 3.2 and 3.3. In addition, maps of peatland extent and carbon density such as the ones shown in Figures 3 and 6 of Kleinen et al. 2012 would be useful.

We doubt whether maps of peatland extent and carbon density would really improve the paper, but will consider adding them.

Overall, peatland extent is mainly determined by topography in our model, with some variation determined by the land water balance P-E. Therefore, peatland extent is very similar in all interglacials. This may be a shortcoming of our peatland model, although this is hard to judge since reconstructions of past peatland extents are poor for the early Holocene and non-existent for previous interglacials.

The modulation of peat accumulation by NH summer insolation is, unfortunately, not clear with respects to the mechanisms. What Yu et al. (2010) show is a correlation between peatland *initiation* and insolation forcing. This translates to a change in total peat accumulation through the change in peatland area, but not necessarily to a change in peat accumulation at any particular site. The increased peatland initiation could either result from increased moisture through increased precipitation, or from increased peat accumulation. For the latter it is unclear, what the exact mechanism might be. The carbon balance in a peatland is determined by productivity and respiration, i.e., NPP-Rh. NPP is dependent on both radiation and temperature, whereas Rh is only dependent on temperature. Which one of the two dominates under changed insolation is difficult to foresee. We will check in our model what the exact sensitivity of peat accumulation to insolation is, but overall the direct sensitivity of peat accumulation to insolation is relatively low in CLIMBER2-LPJ.

In the revised version of the paper will extend the discussion of the peat accumulation rates, also including the sensitivity to climate and insolation changes.

We have also added a discussion of uncertainties in peat accumulation estimates (page 17, line 27 – page 18, line 10), but decided against additional figures since the number of figures is already rather large. Also we decided against adding insolation Figures, since the sensitivity against insolation is negligible in our model, and – as discussed above – the main influence of insolation seems to be with regard to peatland initiation.

3) Why is pCO₂ decreasing between 126 and 122 ka B.P. In Eem-Orb?

Carbon is taken up by both land and ocean. On land we see an uptake of carbon by the soil carbon pools. We also see enhanced weathering due to warmer temperatures at 126 ka in comparison to the Holocene. The stronger weathering leads to an increase in alkalinity, which drives oceanic CO₂ uptake.

We will extend the discussion of marine C changes in the revised manuscript, discussing the differences between the interglacials further.

We have added Figures disaggregating the net flux of carbon into components. These show that the geological flux is rather negative at the time, i.e. stronger weathering leads to a drawdown of CO₂. This is also discussed in the revised manuscript (page 14, lines 16-18 and page 19, lines 27-30).

Minor:

- Is Figure 1 necessary?

At the time of writing we thought it would help to clarify the model parameterisation. It is not necessary, though, so we will consider removing it from the final manuscript.

We removed Fig. 1 from the revised manuscript.

- Figure 5: The reference for the sea level should be added in the legend? i.e. why -3m at 0 ka B.P.?

The sea level forcing we used in our experiments is the result of a forward model simulation of the last eight glacial cycles performed with CLIMBER-SICOPOLIS. Since the model does not know in advance what the final ice sheet mass will be, the sea level at the end of the experiment may differ from zero. In fact the present-day Greenland ice sheet mass is slightly overestimated by the ice sheet model. We did not correct for this when plotting the results, but we will do so for the submission of the revised paper.

We will also extend the discussion of the sea level forcing used (please, see also our reply to reviewer #2).

We have corrected the sea level shown in all Figures and added a discussion of sea level uncertainties (page 18, lines 27-32, page 19, lines 24-26, page 20, lines 5-16). We have also clarified the issue of NH / Antarctic ice sheet changes (page 7, line 29 – page 8, line 1).

- Figure 9: Simulated $\delta^{13}\text{CO}_2$ could be shown.

We will be happy to show it in the revised version.

We show this in the revised paper as Figure 10b.

Author reply to comments by anonymous reviewer #2

We very much thank the reviewer for taking the time to review our manuscript. We aim to incorporate all of the reviewer's comments in the final manuscript, this will lead to a substantial improvement over the original submission. For reader convenience we have included the reviewer's comments in full in this reply, marking them by bold font.

The manuscript describes how the carbon cycle within the EMIC CLIMBER is improved by two slow processes ((a) shallow water CaCO₃ accumulation (coral reef growth) and (b) peat accumulation) and how the improved model is performing for parts of three interglacials (Holocene, Eemian, MIS 11).

The content of the paper is certainly of interest for readers of the journal. However, I believe there are some more steps in the analysis and in the presentation of the paper necessary before it should be accepted for publication in *Climate of the Past*.

My main concerns are the following:

1. One of the objectives to analyse and to compare interglacial carbon cycles was the hypothesis of Ruddiman, who proposed that the rise in CO₂ after 8 kyr BP in the Holocene is due to early anthropogenic contributions (and potential feedbacks). This hypothesis is clearly mentioned in the paper, but most recent idea in that direction are not taken up (e.g. Ruddiman (2013, *The Anthropocene*, *Annual Review of Earth and Planetary Sciences*, DOI: 10.1146/annurev-earth-050212-123944) already claimed that a large peat burial in the Holocene would offset a large anthropogenic CO₂ rise). Furthermore, the authors have chosen to simulate only the later parts of the interglacials, while the first some thousand years in all three interglacials are omitted. This might be motivated by the potential influence of the long-term feedbacks from the previous deglaciation, but then also reduces the chances of really investigation the Ruddiman hypothesis and to compare the interglacials. One might also learn from this decision of the authors to focus on the final part of the interglacials, that in transient simulation the deglaciations need to be taken also into account, when understanding interglacial carbon cycle dynamics as widely as possible. This shortcoming of the study (caused by the chosen setup) might need to be discussed (and maybe motivated) more widely as done so far. Please also note, that others (e.g. Joos et al., 2004; Menviel and Joos 2012) include the whole deglaciation in order to understand Holocene carbon cycle dynamics.

We have the impression that the reviewer may have misunderstood our intentions. It was neither our intent to fully investigate and discuss the "Early anthropogenic hypothesis" by Ruddiman (2003) including later modifications (e.g., Ruddiman 2013), nor to fully explain the last glacial-interglacial cycles. Instead our intent was much more modest: we aim at understanding *trends* in the carbon cycle during three recent interglacials. We will clarify this in the revised submission.

With regard to the original early anthropogenic hypothesis, it is addressed in detail in many publications (e.g. Claussen et al., 2005). Later modifications, e.g., the approach to account for peat carbon accumulation (Ruddiman 2013), which partly compensates the land carbon emissions from anthropogenic land use, still cannot fully explain the observed record of an increase in CO₂ and simultaneously relatively stable δ¹³C because both peat and landuse carbon have similar ¹³C signatures. This implies that any compensation of δ¹³C changes from land use changes through peat uptake would require that the entire carbon emitted is taken up by peatlands. Here, we disentangle an oceanic CO₂ source which does not affect δ¹³C from terrestrial sinks (peat) and sources (land use in the Holocene) with significant δ¹³C fractionation.

In addition, some components of the system that are crucial to address the original early anthropogenic hypothesis in full depth are missing in our model. For example, methane emissions from agriculture are not something we can determine – and that would be very difficult to quantify in any meaningful way since we lack data on historical rice agricultural practices.

[We have added a sentence on the Ruddiman hypothesis \(page 17, lines 13-16\), clarifying the aims of our paper.](#)

With regard to the setup of initial conditions, we indeed have a limitation of our equilibrium approach since the carbon cycle is never in equilibrium, neither at the early Holocene nor during the Last Glacial Maximum. Performing transient runs through several glacial cycles would be the most appropriate way to address interglacials, but this is very challenging, both computationally and scientifically. While we have made some progress in simulating the full glacial CO₂ cycle with the CLIMBER-2 model (Brovkin et al., 2012), the

processes that govern the interglacial carbon cycle dynamics are different from those that play a dominant role in glacial periods. During glacial periods, atmospheric CO₂ is mainly driven by changes in ocean volume, SSTs, circulation, and marine productivity, i.e., oceanic processes play a much more important role in the carbon cycle than terrestrial ones. During interglacials, land carbon also plays a significant role, as climate and oceanic circulation are relatively stable and memory effects from the previous glacial period/deglaciation are operating through relatively slow changes in the marine carbonate chemistry. Our approach is to start simulations several thousand years after stabilization of CO₂ in the atmosphere at the beginning of interglacials to reduce the memory effects, but we cannot completely exclude them.

[We have improved the discussion of our experimental setup \(page 9, lines 4-25\) and clarified the reasons for not performing entire glacial cycle experiments \(page 8, line 29 – page 9, line 3\).](#)

2. One of the most interesting aspects of interglacial differences in the carbon cycle is the 0.2‰ offset in atmospheric $\delta^{13}\text{C}$ CO₂ observed from ice cores between Holocene and Eemian (Schneider et al., 2013), while CO₂ itself was comparable between both interglacials. In this data-based study of Schneider it was already suggested, that slow, long-term processes (weathering or volcanism) in the carbon cycle might be responsible for these effects. However, again, the authors have chosen an experimental setup by which this open research question can not be tackled, since they prescribe $\delta^{13}\text{C}$ CO₂ at the beginning of their experiments from data and only simulate its dynamics over the rest of the interglacials. Since it is evident from the Schneider et al. (2013) data, that the sources and sinks for $\delta^{13}\text{C}$ CO₂ changed slowly over time, these results might only be of limited values, and might follow the $\delta^{13}\text{C}$ CO₂ (for those scenarios which meet the data) for the wrong reasons. Again, this is even more than my comment #1 above an argument for transient simulations which cover longer time periods.

Indeed, our approach is limited because we cannot yet model the full glacial cycle. Since the difference between Eemian and Holocene is apparent through the entire interglacial, the reason for this difference must lie somewhere in the glacial period, which we cannot yet model sufficiently well. To simulate drifts in atmospheric $\delta^{13}\text{C}$ from one to another interglacial, we would need (1) to simulate the carbon cycle dynamics through several glacial cycles, and (2) to account for mechanisms which could lead to an imbalance in $\delta^{13}\text{C}$. Such an imbalance could, for example, result from unbalanced sinks or sources of organic material, such as burial of organic material in the marine sediments, or mineralization of carbon stored in permafrost soils during interglacials. The sediment model we use in this study accounts only for carbonate, but not for organic sedimentation. Since neither deep-sea sedimentary organic burial nor permafrost burial are accounted for in our model, we cannot test the “organic burial” hypothesis and have to use observed $\delta^{13}\text{C}$ data as initial conditions for the carbon cycle. Our goal is then to simulate trends in $\delta^{13}\text{C}$ CO₂, and not to explain the difference in the initial conditions.

[Since we cannot run the model through entire glacial cycles \(page 8, line 29 – page 9, line 3\), we cannot address this comment.](#)

3. I can not remember, that the choice of the investigated interglacials (Holocene, Eemian, MIS 11) was ever motivated. Why have the interglacials between Eemian and MIS 11 (MIS 7, MIS 9) not be chosen? There are various studies published, which compared different aspects of interglacial climate (aligning orbital configuration or greenhouse gas changes or temperature records of different interglacials) in search for the best analogue for the Holocene and to investigate the Ruddiman hypothesis (e.g. Ruddiman 2007, Reviews in Geophysics, doi:10.1029/2006RG000207; Yin and Berger 2010 (NGS, DOI:10.1038/NGeo771) 2012 (CD, DOI 10.1007/s00382-011-1013-5) 2015 (QSR, http://dx.doi.org/10.1016/j.quascirev.2015.04.008)). From my reading of the literature MIS 19 seems to be the best analogue of the Holocene.

While it is in principle possible to model any particular interglacial, doing so becomes less and less fruitful as one goes back further in time, due to the lack of data of sufficiently high resolution and precision. Therefore we chose MIS 1 and 5 as a much better test for a model, since sufficient data are available for a meaningful test of model results. We furthermore chose MIS 11 because of its unusual length (e.g., Tzedakis et al., 2012). We will discuss the choice of the analysed interglacials in the revised manuscript.

[We motivate our choice of interglacials on page 3, lines 24-30.](#)

4. The analysis lack some important details on what the marine carbon cycle is doing. So far, one can understand how in the different scenarios carbon is accumulated in terrestrial vegetation, soil or shallow water. However, the changes in biomass+soil (for scenarios investigating the impact of the new peat carbon formation) do not add up to the changes that the anomalies in atmospheric CO₂

produces, implying that the marine carbon cycle is also affected. For example, page 1957, lines 4-10, it is said that the decrease in atmospheric CO₂ of 25 ppmv is explained by the uptake of 320 PgC by peatland growth. However, 25 ppmv in CO₂ correspond only to a change in the atmospheric carbon pool of about 50 PgC, so where are the other (320-50=)270 PgC coming from? Furthermore, shallow water CaCO₃ accumulation also changes ocean alkalinity, which then changes in the marine carbonate system and thus the ability of the ocean to absorb CO₂ from the atmosphere. What is needed here, is either the addition of several new subplots or an overview results table on various additional (mainly marine) carbon pools and fluxes: ocean C content, C content in deep-ocean sediments, shallow-water C content, ocean alkalinity, weathering flux (does weathering change over time and is a function of climate or CO₂ and is it different for different interglacials?). Furthermore, to compare results with earlier studies (e.g. Elsig et al., 2009) the reader would be interested why marine carbon pools changed as they did. Was it because of SST changes or because of carbonate compensation or because a reduced atmospheric CO₂ (due to land carbon uptake) led to outgassing?

On the time scales of interest (i.e., several thousands to tens of thousands of years), it will be unavoidable for the ocean carbon cycle to feed back onto atmospheric perturbations arising from CO₂ exchanges with the terrestrial biosphere and permafrost. In this respect, although we have not looked into the finer details, the 25 ppm (or about 50 PgC) decrease together with the 320 PgC uptake by peatland growth mentioned by the reviewer fits quite well with the usual ocean buffering and carbonate compensation framework. If peatlands take up 320 PgC from the atmosphere, 85%, or about 270 PgC will be replenished from the oceans on time scales of several hundreds to a few thousands of years as a result of ocean buffering (which would already fit the balance fluxes in our case). On longer time-scales of several thousands to a few tens of thousands of years, increased carbonate accumulation in the deep-sea due to the decreased ocean DIC would decrease global ocean alkalinity, thus contributing in turn to reduce the remaining 50 PgC deficit in the atmosphere (compared to the pre-peatland-uptake situation) by an extra one third to one half. However, these latter time scales are possibly already somewhat too long to play a significant role in our case.

Weathering is dependent on climate (via runoff, as stated on p. 1949, ll. 21-23). Therefore it changes with time and is different between the interglacials. During the early Eemian, temperatures, as well as precipitation and runoff, are higher than during the Holocene, leading to stronger weathering.

We plan to extend the discussion of marine C cycle changes in the revised submission of our manuscript (with extra figures and attribution analysis of C cycle changes as appropriate).

We have extended the paper by adding three figures disaggregating the net C flux to the atmosphere into geological, land and ocean components. These Figures are discussed extensively. However, further analyses of reasons for changes in a particular carbon flux in a particular experiment would require additional paired experiments tailored to the exact question. The SST hypothesis mentioned above, for example would require an additional experiment keeping SSTs fixed. These special experiments would clearly go far beyond the scope of the present paper.

5. For the anthropogenic carbon emissions in the Holocene results from Kaplan et al (2011) are taken. However, in order to obtain simulation results which agree with CO₂ data the authors downscaled the Kaplan-based anthropogenic carbon emissions by 25%. I argue that this is an arbitrary non-scientific approach to fit the simulation results to the data. The authors should test different anthropogenic carbon emissions — as they were published — in their model and then discuss how their results meet the data. Please note, that the Kaplan et al. (2011) study contains two different anthropogenic carbon emissions, others are cited within Kaplan et al. (2011) and in Ruddiman (2013). See also Stocker et al (2011) BG, doi:10.5194/bg-8-69-2011.

With regard to the anthropogenic carbon emission scenarios, uncertainties are certainly very large. While we admit that our approach of rescaling the Kaplan et al. scenario is to some extent arbitrary, we do, however, disagree with it being unscientific. First of all, a 25% difference certainly falls within the uncertainty range of the Kaplan et al. (2011) scenario. Secondly, a C emission scenario similar to our rescaled version of the scenario of Kaplan et al. (2011) has been derived from Kaplan's land use change data using a different carbon cycle model (B. Stocker, personal communication). Unfortunately this latter scenario has not yet been published and we therefore cannot use it in this study. Nonetheless we will extend the discussion of the forcing data and include model runs with Kaplan's original scenario, as well as other scenarios, in the revised submission, as recommended by the reviewer.

We have exchanged the criticised experiment HOL_ALL with an experiment HOL_ANT, where we do not reduce Kaplan's emissions, but rather use their original scenario. The consequences of this change are far

smaller than we expected. Running additional experiments with additional scenarios would not improve the paper, since all other scenarios discussed in the literature have smaller emissions and would fall between our HOL_NAT and HOL_ANT experiments. We have therefore not added additional scenarios.

6. The records of sea level change, that are important for the shallow-water CaCO₃ accumulation needs a wider description and discussion. So far, the sea level change (plotted in Figs 5a, 8a 11a) is obtained from CLIMBER-SICOPOLIS coupling. To my knowledge, this setup only considers changes in northern hemisphere land ice, but none from Antarctica. This needs at least to be mentioned or even better discussed. The plotted sea level records which force the coral reef growth should be compared with other sea level records in order to understand if any mismatch here might influence the simulated coral reef growth. In detail: (a) the Holocene sea level does not reach zero, but the over change over time seems to be reasonable; (b) Eemian sea level only falls, while Rohling et al (2008) NGS, doi:10.1038/ngeo.2007.28, finds rising sea level until about 122-123 kyr BP, then falling, clearly in disagreement with Fig 8a; (c) The pronounced sea level variation of CLIMBER (Fig 11a) with rising sea level around 420 ka BP by 20 m and falling around 400 ka BP by 15 m (which shows clearly a large imprint on simulated CO₂ in scenario MISS11_NAT (Fig 9), is this discussed as such in the text?) needs to be compared with others. For MIS-11 please see Rohling et al (2010) in EPSL, doi: 10.1016/j.epsl.2009.12.054, who find a rise and fall in MIS-11 sea level by about 40 m between 420 and 390 ka BP, thus about twice as much as used here. Also note, that deconvolution of benthic $\delta^{18}O$ into temperature and sea level by models (e.g. de Boer et al (2013) CD, DOI:10.1007/s00382-012-1562-2) is different in MIS 11 showing a decreasing sea level from 400 ka BP onward without any plateau around 395-380 ka BP. The paper of de Boer et al (2013) also analyses the contribution of Antarctic ice sheets to sea level, but from my reading it indeed seems to be the case that the Antarctic contribution to sea level change during interglacials is minor, so this is NOT the reason for the disagreement between both studies.

Sea level change contributions from Antarctica are actually included in our model sea level forcing. It is assumed that they are 10% of the NH ice sheet changes, which is a decent approximation for glacial-interglacial changes, but which might underestimate Antarctic contributions to strong sea level high stands during interglacials.

The sea level forcing we used in our experiments comes from a forward model simulation of the last eight glacial cycles performed with CLIMBER-SICOPOLIS. Since the model does not include any a priori information about the final ice sheet mass, the sea level at the end of the experiment may differ from zero. In fact the present-day Greenland ice sheet mass is slightly overestimated by the ice sheet model. We did not correct for this mismatch when plotting the results, but we will do so for the submission of the revised paper.

The Holocene is the only interglacial where sea level reconstruction are reliable, and here our model is in very good agreement with observations.

Although the CLIMBER-SICOPOLIS results for the Eemian are clearly different from Rohling et al (2008), they are very similar to IPCC AR5, Chapter 5, Figs. 5.15 a and b. We therefore believe that our results are reasonable for the Eemian.

For MIS 11 we estimate that uncertainties on the reconstructed sea-level stands are probably +/- 20 meters at the very best. Rohling et al. (2010) and also Grant et al. (2014) indeed find a sea level substantially lower than in our model at 390 ka BP, but they also find sea levels 5-10m below present during the entire MIS 11, while other studies (Raymo et al., 2012; Bowen, 2010) show sea levels 5-10m above present. De Boer et al. (2013) indeed find a decrease after 400 ka, Rohling et al. (2010) and Grant et al. (2014) document a plateau around 395-380 ka BP, and Elderfield et al. (2012) a rise in sea level during this period of time.

Our model sea-level therefore fits well into the available reconstructions. We will discuss these issues in more detail in the revised manuscript.

We have corrected the sea level shown in all Figures and added a discussion of sea level uncertainties (page 18, lines 27-32, page 19, lines 24-26, page 20, lines 5-16). We have also clarified the issue of NH / Antarctic ice sheet changes (page 7, line 29 – page 8, line 1).

7. After this revision the whole discussion section probably needs a complete rewriting.

We agree.

We have completely rewritten the results section and strongly modified relevant parts of the discussion section.

Minors:

1. The title should be changed according to what is contained in the paper, e.g. "The importance of peat accumulation and coral reef growth for the carbon cycle dynamics during interglacials in MIS1, 5, 11".

The present title might indeed raise reader expectations that the paper would not fulfil. Unfortunately the title suggested by the reviewer does not quite fit our paper either, since we do include a full carbon cycle in our model. We will reconsider the title, though, and aim to make it fit better to the paper.

We have modified the title to "Interglacial carbon cycle dynamics during the Holocene, the Eemian and MIS11". We believe the new title fits the content of the paper well.

2. It is difficult to compare the dynamics during the different interglacials from the way the results are plotted right now. At best, the changes in CO₂ and $\delta^{13}\text{C}$ are given for all 3 interglacials on plots, that have the same scales in x and y direction, see for example Fig 11 of Yin and Berger 2015 (QSR).

We will try to add a figure of all interglacials on the same axes.

We tried to add such a Figure, but found it impossible to do well. We also think that the interglacials are too different from each other to make such a figure meaningful.

3. Although no atmospheric $\delta^{13}\text{C}$ data from ice cores yet exist for MIS 11 it would of course be of interest to see the educated guess (simulation results) of $\delta^{13}\text{C}$ from this study, which might illustrate, what dynamics in that variable might be expected.

We will include it in the revised submission.

We show this in the revised paper as Figure 10b.

4. What is called "shallow-water CaCO₃ sedimentation" throughout the text is for my understanding "shallow-water CaCO₃ accumulation", please change.

We will clarify the text.

We have corrected this throughout the text.

5. page 1946, line 23: "While the Holocene CO₂ trend has generated considerable interest previously (Ruddiman, 2003), the context of previous interglacials has been neglected." This is not correct. The whole idea of the Ruddiman hypothesis is about the trend in CO₂ (and CH₄) in the Holocene in comparison to other interglacials. It might be correct that so far no process-based carbon cycle models addressed other interglacials. Please rephrase.

We will clarify the text.

We have modified the text accordingly (page 2, line 5).

6. page 1949, line 5: "DGVM" was already explained on page 1948.

Thank you for pointing this out.

7. page 1950, line 14: Please state briefly name and reference of the DGVM embedded within CLIMBER, probably VECODE.

Thank you for pointing this out, we will clarify the text.

We have clarified this. (Page 4, line 13)

8. page 1950, line 27: ...“corals as the main” SHALLOW WATER “carbonate producers”

We will clarify the text.

We have clarified this (Page 5, line 19)

9. page 1951, line 9: Please give a reference for the SST growth limit of corals.

We have modified the description of the coral accumulation model, making clearer that these were taken from the original Kleypas model (Page 5, lines 21-29).

10. Please include a figure, in which the vertical coral accumulation rate G is plotted as function of light. No values of the parameters G_{max} and l_k are yet given. Please extend on parameter values and motivation (reference) for your choice.

Parameters (incl. SST growth limits – reviewer’s point 9 above) were taken from Kleypas (1997). We will revise the text to make this clearer. Since a figure of G over l_z is already included in original Kleypas (1997), we prefer not to print this again but instead refer to the original paper.

We have modified the description of the coral accumulation model, making clearer that these were taken from the original Kleypas model (Page 5, lines 21-29).

11. page 1952, line 16: “last glacial maximum” should be written as “Last Glacial Maximum (LGM)”, that would then introduce “LGM” which is used later-on.

Thank you for pointing this out.

We have clarified this.

12. page 1953, line 3: It is not clear if “this publication” is related to “Yu et al (2010)” or to this manuscript (Kleinen et al 2015).

It was the Yu et al. ‘2010) paper that was meant. Thank you for pointing out this source of potential misunderstanding, we will correct the text accordingly.

We have clarified this. Page 7, line 16

13. page 1953, line 16: There is no reference “Ganopolski et al (2011)” in the reference list, maybe you mean “Ganopolsi and Calov (2011)”, please check and correct.

We will correct the citation.

We have corrected this. Page 7, line 29.

14. Ice core CO₂ data: The authors might refer to the most recent compilation of ice core CO₂ data on the most recent ice core age model as published in (and available in the supplement to) Bereiter et al (2015) in GRL.

Unfortunately the original submission was written before the compilation by Bereiter et al. was available. We will refer to it in the revised version of the paper.

We have used the Bereiter et al compilation throughout the paper, see page 8, lines 17-23 and Figure captions.

15. Ice core $\delta^{13}C_{CO_2}$ data: I suggest to show the Monte-Carlo-based spline through all available $\delta^{13}C_{CO_2}$ data as published in Schmitt et al (2012) in Science, DOI:10.1126/science.1217161 (here: the Elsig data as taken so far in this manuscript are included) and in Schneider et al (2013) Climate of the

Past; doi: 10.5194/cp-9-2507-2013. The Schmitt spline is available as download at Science, and the Schneider spline certainly via email from the Bern ice core group.

When writing the original submission, we decided to use the raw data in order to also show the uncertainties in the measurements. We will reconsider that choice and also show the MC spline.

[We have used MC estimates for both the Holocene and the Eemian \(the latter, btw., is available from Pangaea\). \(Page 13, lines 18-32; page 14, l. 11-14\)](#)

16. page 1956, line 24: “terrestrial biomass”, this means vegetation? If so, say so.

We will clarify the text.

[We have clarified this throughout the text.](#)

17. page 1958, line 20: Please include SHALLOW WATER before “CaCO₃ accumulation rate”.

We will clarify the text.

[We have clarified this.](#)

18. page 1959, lines 1-5: Modelled CO₂ and $\delta^{13}\text{CO}_2$ are within the range of the data (including errors). Please expand on what the variations in simulation and data are, not just that you meet the data, and briefly mention where there are disagreements, I again suggest to use the spline for $\delta^{13}\text{CO}_2$ data.

We will extend the discussion of model results and data.

[We have extended the discussion of our CO₂ and \$\delta^{13}\text{CO}_2\$ results \(page 14, line 11-14\).](#)

19. Discussion: As explanation (a) of the misfit to the Holocene $\delta^{13}\text{CO}_2$ data it is suggested that Elsig underestimates the true uncertainty. By using the spline in $\delta^{13}\text{CO}_2$ such a potential shortcoming should be overcome. Furthermore, another explanation for the misfit might be, that the marine C cycle change (which are not yet described, see my major point #4) are wrong.

When writing the original submission we had underestimated the significance of the MC spline fit. We will reconsider that choice for the revised submission. We will also extend the discussion to marine C cycle changes, although these are less relevant for $\delta^{13}\text{CO}_2$ in our model. Nonetheless we will check for this when we analyse marine C changes, as written in our reply to the reviewer's major point #4.

[We have removed this point from the discussion since the use of the MC spline indeed clarifies this. Marine C cycle changes, however, generally have minor impact on \$\delta^{13}\text{C}\$ in our model.](#)

20. Figures: In the figures which show ice core data, the ice cores from which the data are, should be mentioned in the caption (at best with reference) and the age model, on which the data are plotted.

We will clarify this in the figure caption.

[We have modified the captions for Figures 2, 6, and 10 accordingly, though we found mentioning the age model excessive, so the latter is only mentioned in section 2.5 \(page 8, line 22\).](#)

21. Figure 4: No results for HOL_PEAT are shown, or are they similar to HOL_NAT? If they are indeed similar, I have probably not fully understood the modelling setup. My understanding is, that the internal simulated atmospheric CO₂ concentration is used by the CLIMBER model to calculate also any temperature changes via the greenhouse effect. This would imply, that any change in CO₂ would change temperature and therefore also peat accumulation. I therefore expect that results for HOL_PEAT and HOL_NAT differ. Please extend the model description in order to clarify this issue. But

maybe I missed some details, e.g. a different coupling scheme between climate and carbon cycle.

Results for HOL_PEAT and HOL_NAT are indeed different since climate and CO2 are different. We decided not to show them to avoid overloading the Figure. We will reconsider this choice for the revised submission.

[We have modified Figure 4 to now include all Holocene experiments.](#)

Formatiert: Kopfzeile

~~Carbon~~ Interglacial carbon cycle dynamics during recent interglacialsthe Holocene, the Eemian and MIS 11

Kommentar [TK1]: Reviewer 2, minor 1

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Feldfunktion geändert

Abstract

Trends in the atmospheric concentration of CO₂ during three recent interglacials, the Holocene, the Eemian and Marine Isotope Stage (MIS) 11, are investigated using an Earth system Model of Intermediate Complexity, which we extended with process-based modules to ~~dynamically determine~~consider two slow carbon cycle processes — peat accumulation and shallow-water CaCO₃ sedimentation (coral reef formation). For all three interglacials, model simulations considering peat accumulation and shallow water CaCO₃ sedimentation substantially improve the agreement between model results and ice core CO₂ reconstructions in comparison to a carbon cycle setup neglecting these processes. This enables us to model the trends in atmospheric CO₂, with modelled trends similar to the ice core data, forcing the model only with orbital and sea level changes. During the Holocene, anthropogenic CO₂ emissions are required to match the observed rise in atmospheric CO₂ after 3 ka BP, but are not relevant before this time. ~~Therefore our~~Our model experiments show ~~for a considerable~~improvement in the first time how modelled CO₂ trends by the inclusion of the slow carbon cycle processes, allowing us to explain the CO₂ evolution during the Holocene and two recent interglacials ~~can be explained~~ consistently using an identical model setup.

Kommentar [TK2]: Editor

1 Introduction

The atmospheric concentration of carbon dioxide (CO₂) increased from 260 to 280 ppm CO₂ during the Holocene between 8 ka BP and preindustrial. This trend in CO₂ has to be seen in

Formatiert: Kopfzeile

1 the context of previous interglacials, since all processes affecting the atmospheric
2 concentration, with the exception of possible human influences, should have been active
3 during all interglacials. While the Holocene CO₂ trend has generated considerable interest
4 previously (Ruddiman, 2003), the context of previous interglacials has been neglected. ~~in~~
5 ~~process-based carbon cycle model studies~~. The present study aims ~~to fill~~ ~~at filling~~ this gap.

Kommentar [TK3]: Reviewer 2, minor
5

6 Investigations of the Holocene trend in CO₂ can be classified into two basic approaches: an
7 inverse modelling approach, and a forward or process-based modelling approach. The inverse
8 modelling approach takes the ice core record of CO₂ and δ¹³CO₂ as a starting point and aims
9 ~~to deduce~~ ~~at deriving~~ the sources and sinks of CO₂ from this record, while the forward
10 modelling approach starts from the carbon cycle processes and aims ~~to determine~~ ~~at~~
11 ~~determining~~ a CO₂ trajectory from combinations of these.

12 Following the inverse modelling approach, based on records of CO₂ and its stable carbon
13 isotopic ratio δ¹³CO₂ from ice cores, Indermühle et al. (1999) deconvolved the mass balance
14 equations for CO₂ and δ¹³CO₂ to solve for the unknown terrestrial and oceanic sources and
15 sinks of CO₂. They explained the changes in atmospheric CO₂ by major contributions from
16 decreases in land carbon (C) storage and changes in sea surface temperature (SST), while
17 changes in the cycling of CaCO₃ played only a minor role. This approach was subsequently
18 refined by Elsig et al. (2009) who presented ~~new atmospheric δ¹³CO₂ records of δ¹³CO₂~~ with
19 higher resolution and better precision. They ~~explained~~ ~~attributed~~ the change in atmospheric
20 CO₂ between 8 ka BP and ~~the~~ preindustrial ~~by~~ ~~to~~ carbonate compensation induced by earlier
21 land-biosphere uptake, as well as coral reef formation, with some contribution ~~by~~ ~~from~~ carbon
22 ~~releaser~~ ~~released~~ from the land biosphere.

23 Using the forward modelling approach, Ridgwell et al. (2003) used estimates of deep ocean
24 carbonate ion concentrations to constrain the carbon cycle. They found that the observed trend
25 in atmospheric CO₂ during the last 8000 years can best be explained by the buildup of coral
26 reefs and other forms of shallow water carbonate deposition. Joos et al. (2004), employing the
27 Bern carbon cycle climate model to simulate the interval from the ~~last glacial maximum~~
28 ~~to~~ Last Glacial Maximum (LGM) ~~to the~~ preindustrial, found that a combination of processes
29 contributed to the Holocene rise in CO₂, with carbonate compensation in response to
30 terrestrial vegetation regrowth, SST changes and coral reef buildup playing a role. On the
31 other hand, Brovkin et al. (2002), as well as Menviel and Joos (2012), found almost no effect
32 of SST changes on CO₂ during the Holocene.

Kommentar [TK4]: Reviewer 2, minor
11

1 Kleinen et al. (2010), using the CLIMBER2-LPJ model, showed that the trend in atmospheric
2 CO₂ over the Holocene is controlled by the balance of two slow processes: ~~Carbon~~carbon
3 uptake by boreal peatlands, which is (slightly over-) compensated by outgassing of CO₂ due
4 to ~~sedimentation~~the accumulation of CaCO₃ in shallow oceanic areas. Finally, Menviel and
5 Joos (2012) investigated the Holocene CO₂ rise by applying the Bern3D ocean carbon cycle
6 model, ~~prescribing with~~ prescribed scenarios of shallow-water carbonate
7 ~~sedimentation~~accumulation and land C uptake. In their experiments, shallow-water carbonate
8 ~~sedimentation, carbonate compensation of land uptake~~accumulation, land carbon uptake and
9 release, and the consecutive carbonate compensation response, as well as the response of the
10 ocean-sediment system to marine changes during the termination contribute roughly equally
11 to the CO₂ rise.

12 For earlier interglacials, investigations are rare. Schurgers et al. (2006) investigated the
13 changes in atmospheric CO₂ during ~~both~~ the Holocene and during the Eemian using the ~~GCM~~
14 ECHAM3-LSG General Circulation Model (GCM), including the dynamic global vegetation
15 model (DGVM) LPJ and the marine biogeochemistry model HAMOCC3. They found
16 increases in atmospheric CO₂ for both the Eemian and the Holocene, mainly driven by
17 decreases in the terrestrial C storage, ~~but they do not. They were, however, unable to~~
18 the overall magnitude of the CO₂ trend during the Holocene, and their positive trend in
19 ~~Eemian atmospheric~~ CO₂ during the Eemian is ~~distinct~~different from that in the ice core data,
20 which ~~shows~~actually show no trend.

21 ~~Here~~In the present publication we ~~address two major shortcomings of advance on~~ the study by
22 Kleinen et al. (2010): in two respects: (1) both the accumulation of peatland carbon and the
23 burial of CaCO₃ were ~~prescribed~~scenarios and not modelled interactively, and (2) the study
24 only considered the Holocene, while neglecting to show that the same mechanisms can also
25 explain the evolution of CO₂ during previous interglacials. Our ~~current~~ now includes a
26 dynamic peatland model, as well as a dynamic model of carbonate accumulation by coral reef
27 ~~growth~~reefs, which finally enables us to consistently investigate the evolution of atmospheric
28 CO₂ ~~in~~during the Holocene and during two interglacials ~~previous~~that preceded it: the Eemian
29 and MIS 11. The Holocene and the Eemian are particularly interesting because validation data
30 of reasonable time resolution and reliability are available for these interglacials. We did not
31 investigate interglacials prior to the Holocene. In this paper, we therefore aim MIS 5 since no
32 δ¹³CO₂ data are available for validation, with the exception of MIS 11 since its unusual length

1 ~~makes it a particularly interesting case and ice core CO₂ data of reasonable time resolution are~~
2 ~~still available]. We investigate to show how what extent~~ the evolution of CO₂ ~~in~~during these
3 three ~~recent~~different interglacials, ~~the Holocene, the Eemian, and MIS 11,~~ can be explained by
4 the interplay of two slow carbon cycle processes, peat accumulation and CaCO₃ accumulation
5 in shallow waters.

Formatiert: Kopfzeile

Kommentar [TK5]: Editor, Reviewer 2,
major 3

7 2 Model and experiments

8 2.1 The model

9 To investigate these questions we are using CLIMBER2-LPJ, which consists of the Earth
10 system Model of Intermediate Complexity (EMIC) CLIMBER2, coupled to the ~~dynamic~~
11 ~~global vegetation model (DGVM) -~~LPJ. This combination of models allows experiments on
12 timescales of an interglacial due to the low computational cost of CLIMBER2, while
13 accounting for the heterogeneity of land surface processes on the much finer grid of LPJ.

14 CLIMBER2 (Petoukhov et al., 2000, Ganopolski et al., 2001) consists of a 2.5-dimensional
15 statistical-dynamical atmosphere with a latitudinal resolution of 10° and a longitudinal
16 resolution of roughly 51°, an ocean model resolving three zonally averaged ocean basins with
17 a latitudinal resolution of 2.5°, a sea ice model, and ~~at~~the dynamic terrestrial vegetation model
18 VECODE (Brovkin et al., 2002). In the present model experiments, the latter model is used
19 only for determining biogeophysical responses to climate change, ~~(i.e. as a land surface~~
20 ~~scheme for the climate model),~~ while biogeochemical effects, i. e., the corresponding carbon
21 fluxes, are determined by LPJ. VECODE and LPJ produce similar vegetation changes and
22 discrepancies therefore are very small.

Kommentar [TK6]: Reviewer 2, minor
7

23 ~~In addition~~ CLIMBER2 also contains an oceanic biogeochemistry model (Ganopolski et al.,
24 1998, Brovkin et al., 2002, 2007) and a sediment model that describes the diffusive pore-
25 water dynamics, assuming oxic ~~-~~only respiration and 4.5-order CaCO₃ dissolution kinetics
26 (Archer, ~~1996~~1991; Brovkin et al., 2007). Volcanic emissions of CO₂ are assumed to be
27 constant at 0.07 GtC a⁻¹ (Gerlach, 2011). Weathering fluxes scale to runoff from the land
28 surface grid cells, with separate carbonate and silicate lithological classes. The long-term
29 carbon cycle that includes the processes of deep-sea and shallow-water carbonate
30 accumulation, weathering and volcanic outgassing, is brought to equilibrium for the pre-
31 industrial climate as in Brovkin et al. (2012).

We have coupled the DGVM LPJ (Sitch et al., 2003, Gerten et al., 2004) to CLIMBER-2 in order to investigate land surface processes at a resolution significantly higher than that of CLIMBER2. We also extended the model by implementing carbon isotope fractionation according to Scholze et al. (2003). LPJ is run on a $0.5^\circ \times 0.5^\circ$ grid and is called at the end of every model year simulated by CLIMBER2. Anomalies from the climatology of the temperature, precipitation and cloudiness fields are passed to LPJ, where they are added to background climate patterns based on the CRU-TS climate data set (New et al., 2000). In order to retain some temporal variability in these climate fields, the anomalies are not added to the climatology of the CRU-TS data set, but rather to the climate data for one year randomly drawn from the range 1901--1930. The change in the LPJ carbon pools is then passed back to CLIMBER2 as the carbon flux F_{AL} between atmosphere and land surface and is employed to determine the atmospheric CO_2 concentration for the next model year.

~~Biogeochemical feedbacks between atmosphere and land surface are thus determined by the combination of CLIMBER2 and LPJ, while biogeophysical effects are solely determined by the CLIMBER2 land surface model, which includes its own dynamical vegetation model. The latter model produces vegetation changes very similar to LPJ. Therefore discrepancies are very small.~~

2.2 Accumulation of Calcium carbonate in shallow waters

The accumulation of $CaCO_3$ in shallow waters leads to an increase in the atmospheric CO_2 concentration. The production of $CaCO_3$ proceeds following the carbonate precipitation

equation ~~$Ca^{2+} + 2HCO_3^- \rightarrow CaCO_3 + CO_2 + H_2O$~~ $Ca^{2+} + 2HCO_3^- \rightarrow CaCO_3 + CO_2 + H_2O$.

Under present conditions in seawater about 0.6 mol of CO_2 will be released for every mol of $CaCO_3$ produced (Frankignoulle et al., 1994). This is implicitly handled by the carbonate speciation and air-sea gas exchange routines.

~~As part of the~~The marine carbon cycle in CLIMBER2 ~~contains a model of early diagenesis of carbonate in the deep sea sediments (Archer, 1996; Brovkin et al., 2007) and has been extended by~~ a model of carbonate accumulation in shallow waters, which was derived from ReefHab (Kleypas 1997). The original ReefHab predicts reef habitat area and accumulation of $CaCO_3$ in these environments as a function of temperature, salinity, nutrients, and light. The model considers corals as the main shallow-water carbonate producers, but it is also applicable to calcareous algae, which have calcification rates very similar to corals.

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1 For the implementation in CLIMBER2, we determined the potential reef area A by diagnosing
2 the sea floor area above the maximum depth of reef growth for each ocean grid cell,
3 depending on the global sea level, from the ETOPO2 data set (US Dept. of Commerce, 2006).
4 In addition, we determined the topographic relief function TF , as described by Kleypas
5 (1997). The vertical coral accumulation rate we then determine ~~as $G = G_{max} \tanh(I_z/I_k)$~~
6 ~~according to Kleypas (1997) as $G = G_{max} \tanh(I_z/I_k)$~~ , with G_{max} the maximum
7 accumulation rate, I_z the Photosynthetically Active Radiation (PAR) at depth z , and I_k the
8 saturating light intensity necessary for photosynthesis. We calculate G for all grid cells where
9 SST > 18.1°C and < 31.5°C, the growth limits for corals- (Kleypas, 1997).

Kommentar [TK8]: Reviewer 2, minor
9 and 10

10 In the original Kleypas (1997) model, sea level is only used to calculate the area available for
11 shallow water sedimentation, but the rate of sea level change is not considered in calculating
12 the rate of CaCO₃ sedimentation. However, the rate of CaCO₃ accumulation by coral reefs
13 will be strongly perturbed during periods of sea level drop or very fast sea level rise. A
14 moderate rate of sea level rise, on the other hand, can maximise coral reef buildup. We
15 therefore implemented a dependence of the CaCO₃ sedimentation rate on the rate of sea level
16 change based on Munhoven and François (1996). Munhoven and François (1996) consider a
17 trapezoidal growth-limiting function ~~θ as shown in Fig. 1,~~ which restricts the coral reef
18 growth in case sea-level rises too fast or falls. According to Buddemeier and Smith (1988) the
19 best overall estimate for the sustained maximum rate of reef growth is 10 mm a⁻¹. For
20 simplicity we therefore adopt 0 and 10 mm a⁻¹ as the limiting sea-level rates. To avoid too
21 abrupt ~~a changechanges~~, accumulation rates are reduced from 100 to 0% of the normal rate
22 from 10 to 15 mm a⁻¹; similarly we let them increase from 0 to 100% from -2.5 mm a⁻¹ (i.e., a
23 2.5 mm a⁻¹ decrease) to +2.5 mm a⁻¹. We thus allow for a small accumulation ~~even when as~~
24 ~~long as~~ sea-level falls ~~only slowly~~. Carbonate accumulation rates will not drop to zero
25 immediately since corals may live even at depths of 50 m and more, and their habitat
26 therefore does not vanish immediately.

Kommentar [TK9]: Reviewer 1, minor
1

27 The total CaCO₃ ~~productionaccumulation~~ in each grid cell where ocean temperature is within
28 the acceptable range therefore is ~~$P = G \times \theta \times A \times TF$~~ ; $P = G \times \theta \times A \times TF$, which we sum
29 up for all grid cells to determine the total shallow-water CaCO₃ ~~productionaccumulation~~.
30 Total ~~productionaccumulation~~ is scaled to conform to the Milliman (1993) estimate of
31 shallow water CaCO₃ ~~sedimentationaccumulation~~ for the late Holocene. Milliman (1993)

Kommentar [TK10]: Reviewer 2,
minor 4, also other places

1 estimates a ~~sedimentation rate in shallow waters~~ water carbonate accumulation rate of about
2 ~~1.5 bt a⁻¹ (billion tons, Milliman's units)~~ tonnes of CaCO₃ per year, which converts to 15
3 Tmol a⁻¹ ~~using the CaCO₃ molar weight of 100 g mol⁻¹~~. The area factors A and TF are more or
4 less constant over the sea level range of our experiments. Therefore variations in CaCO₃
5 formation are primarily due to ~~changes~~ variations in the rate of sea level change. In
6 experiments where the dynamic calculation of CaCO₃ sedimentation is disabled, a ~~small~~
7 constant shallow water CaCO₃ sedimentation flux of 2 Tmol a⁻¹ is prescribed to balance the
8 oceanic alkalinity budget.

9 2.3 Carbon accumulation in peatlands

10 According to Yu et al. (2010), global peatlands store about 615 Pg of carbon in the form of
11 peat soils. ~~The bulk of the carbon is contained in:~~ northern high latitude peatlands, ~~which~~
12 ~~contain~~ account for about 550 PgC, ~~while~~ tropical peatlands ~~have accumulated~~ for about 50
13 PgC and southern peatlands for about 15 PgC. This carbon was largely accumulated since the
14 ~~last glacial maximum~~ LGM.

15 In order to account for this accumulation of carbon, we have extended ~~the~~ CLIMBER2-LPJ
16 ~~model~~ by ~~developing a~~ the dynamic model of peatland extent and peat carbon accumulation, ~~as~~
17 ~~described in of~~ Kleinen et al. (2012). ~~This~~ The model determines peatland extent from
18 topography and climatic conditions. ~~Within~~ In the obtained peatland areas ~~obtained~~ it
19 ~~considers~~ accumulates carbon due to the ~~anoxic~~ slow decomposition of C under the anaerobic
20 conditions in ~~the soil to accumulate carbon in the modelled~~ peatlands. For the last 8 ka, this
21 model calculates an accumulation of 330 PgC in high northern latitude areas, which is
22 roughly in line with the Yu et al. (2010) estimate of 550 PgC for the time period from the
23 LGM to the present (Kleinen et al., 2012). The main factor influencing the uncertainty of this
24 model result is the peatland area estimate. Kleinen et al. (2012) considered minimum and
25 maximum area estimates from which they derived an uncertainty range of 240-490 PgC for
26 the peat accumulation between 8 and 0 ka BP, with a most likely value of 330 PgC.

Kommentar [TK11]: Reviewer 1,
major 1 and 2

27 Tropical peatlands could, unfortunately, not be considered in the present experiments, due to
28 the lack of reliable calibration data for tropical peatlands. Preliminary model experiments
29 ~~for~~ indicate that the ~~Holocene show a constant~~ carbon stock in tropical peatlands ~~has not~~
30 varied over the Holocene, though, ~~and we~~ We therefore assume that we introduce no major
31 errors by neglecting them. Furthermore, they represent less than 10% of the total, according to

1 the figures from Yu et al. (2010). ~~Experiments in this publication.~~ It should be noted that
2 experiments where peat accumulation is considered, display a decreased total carbon stock for
3 soil carbon in mineral soils in comparison to the experiments where peat accumulation is not
4 considered. In ~~these~~ the former type of experiments the area covered by mineral soils is
5 smaller since part of ~~the~~ grid cell may be set aside for peatlands. The offset in total carbon
6 stocks between the experiments with and without consideration of peat carbon accumulation
7 therefore does not reflect a different carbon density in any particular location, but rather the
8 reduced area of mineral soils.

9 2.4 Forcing data

10 The model is forced by orbital changes following Berger (1978) in all experiments. For the
11 experiments that include shallow water CaCO₃ accumulation, we also force the model by
12 providing sea level data. We obtained the sea level, as well as the rate of sea level change,
13 from a previous experiment performed with CLIMBER2 coupled to the ice sheet model
14 SICOPOLIS, run over the last 8 glacial-interglacial cycles (Ganopolski ~~et al.,~~ and Calov,
15 2011). The sea level change in these experiments is mainly derived from changes in the
16 northern hemisphere (NH) ice sheets, though changes in Antarctic ice sheet size are
17 considered by assuming that these are 10% of the NH changes. The global ice sheet volume
18 obtained compares favourably well with the reconstruction of sea level by Waelbroeck et al.
19 (2002).

20 One model experiment for the Holocene is also forced with ~~data on~~ anthropogenic carbon
21 ~~emission~~ emission data. We obtained a scenario of carbon emissions from land use changes
22 from Kaplan et al. (2011), who reconstructed global changes in land use over the last 8000
23 years and provided a scenario of corresponding carbon emissions. In addition, we use data on
24 carbon emissions from fossil fuel use and cement production from 1765 onwards. The
25 scenario is called KK10. It covers the time from 8 ka BP to 1900 AD, and we extend it to
26 1950 with land use from the RCP scenario database (Meinshausen et al., 2011).

27 ~~The Kaplan et al. (2011) scenario on CO₂ emissions from land use changes assumes~~
28 ~~cumulative emissions of ~409 PgC by 1950 (0 a BP), which we found to lead to excessively~~
29 ~~high CO₂ concentrations for the present, when combined with historical fossil fuel CO₂~~
30 ~~emissions. We therefore scaled their emission scenario by a constant factor of 0.75 to reduce~~
31 ~~the total cumulative release to 307 PgC by 1950, keeping the timing of their CO₂ emissions.~~
32 After 1765 (or 185 a BP) we After 1765 AD (or 185 a BP) we also add historical emissions

Formatiert: Kopfzeile

Kommentar [TK12]: Reviewer 2,
minor 12

Kommentar [TK13]: Reviewer 2,
major 6

Kommentar [TK14]: Reviewer 2,
major 5

1 from fossil fuel use from the RCP database (Meinshausen et al., 2011). The adopted
2 cumulative emissions are shown in Fig. 21. For simplicity, CO₂ emissions from land use
3 changes are directly added to the atmospheric CO₂, i.e., we do not change the land carbon
4 stocks when emitting CO₂ from land use changes. This simplification will lead to a slight
5 overestimate of the carbon uptake by vegetation through CO₂ fertilisation, ~~though we~~. We
6 nevertheless judge ~~it~~the impact of this simplification to be minor. Both land use and fossil
7 fuel emissions are assumed to have a $\delta^{13}\text{CO}_2$ of -25%.

8 2.5 Ice core data

9 We compare the atmospheric CO₂ concentrations from our experiments to CO₂ concentration
10 reconstructions from ice cores. ~~For the Holocene, we use the CO₂ reconstruction by Monnin~~
11 ~~et al. (2004), obtained by analysing ice cores from Dome Concordia (EDC) and Dronning~~
12 ~~Maud Land. From their reconstruction we use the CO₂ concentration from EDC and the~~
13 ~~corresponding one sigma error bars. For the most recent times, we extend their time series by~~
14 ~~using data from Law Dome published by Etheridge et al. (1996), who provide CO₂~~
15 ~~concentration only. For $\delta^{13}\text{CO}_2$, we compare to the data obtained from EDC by Elsig et al.~~
16 ~~(2009), including their error estimate. We use the recent compilation by Bereiter et al. (2015),~~
17 ~~which consists of data from Law Dome (Rubino et al., 2013; MacFarling Meure et al., 2006)~~
18 ~~and EPICA Dome C (EDC) (Monnin et al. 2001; 2004) for the Holocene, data from EDC~~
19 ~~(Schneider et al., 2013) for the Eemian, and data from Vostok (Petit et al., 1999) and EDC~~
20 ~~(Siegenthaler et al., 2005) for MIS 11. The data are on the AICC2012 time scale (Bazin et al.,~~
21 ~~2013), with the exception of the data from Law Dome.~~

22 ~~For the Eemian, we compare with data by Schneider et al. (2013) for both CO₂ and $\delta^{13}\text{CO}_2$.~~
23 ~~This data was also obtained from EDC, and error estimates from sample replication are~~
24 ~~provided for most of the data points. For MIS 11, we use the data from the EDC (Siegenthaler~~
25 ~~et al., 2005) and Vostok (Petit et al., 1999; Raynaud et al., 2005) ice cores on the EDC3 gas~~
26 ~~age time scale, as published by Lüthi et al. (2008). For this data no detailed error estimate is~~
27 ~~provided, though Petit et al. estimate an error range of +/- 2-3 ppmv.~~

28 ~~For $\delta^{13}\text{CO}_2$, we compare the Holocene results to the compilation by Schmitt et al. (2012),~~
29 ~~based on data obtained from EDC by Elsig et al. (2009) and Laurantou et al. (2010), and we~~
30 ~~compare the Eemian results to data from Schneider et al. (2013). For both time frames, we~~
31 ~~compare to both the original data and to the Monte Carlo average, which should remove most~~
32 ~~of the analytical uncertainties.~~

2.6 Model experiments

Due to its long memory the carbon cycle is not in equilibrium at any particular point in time during an interglacial. The best approach to investigate the carbon cycle during an interglacial would therefore be to perform a model simulation of several glacial cycles. This would ensure that the carbon cycle is equilibrated to the time-varying boundary conditions as much as possible. Unfortunately this approach is not yet feasible, in particular because of computational constraints. We therefore aim to initialise the model to conditions early in the interglacial but after the large transient changes associated with the deglaciation are over. For the Holocene this implies starting the model simulation at 8 ka BP, when most of the ice sheets have melted and the initial regrowth of vegetation is finished. For the Eemian we begin the model experiment at 126 ka BP, after the large transient peak in CO₂ has decayed, and for MIS 11 we start the model at 420 ka BP. From these starting points onward, we drive the model with orbital and other forcings as appropriate until the end of the experiment at 0 ka, 116, and 380 ka BP for the Holocene, the Eemian and MIS 11, respectively.

Since the carbon cycle cannot be regarded as being in equilibrium on multi-millennial timescales, we initialized the model for our experiments with a similar procedure as in Kleinen et al. (2010). Firstly, the model was run with equilibrium conditions appropriate for the beginning of the respective interglacial, including constant CO₂ as diagnosed from ice cores for ~~thethat~~ time. Atmospheric δ¹³CO₂ was also initialized to the ice core value. In a second step, ocean alkalinity was increased to get a carbonate sedimentation flux of 16 Tmol a⁻¹ in the deep ocean and 2 Tmol a⁻¹ on the shelves in order to simulate the maximum in CaCO₃ preservation in the deep sea before the onset of the interglacial. The model was then run with prescribed CO₂ for 5000 years. This setup of initial conditions ensures that the ocean biogeochemistry is in equilibrium with the climate at the onset of the interglacial, while it is in transition from the glacial to interglacial state thereafter. Initial times and CO₂ concentrations are summarized in Table 1. After the climate model state for the beginning of the model experiment has been obtained, this climate state is used for a separate offline spin up of the LPJ DGVM to determine an appropriate vegetation distribution and land carbon storage for the beginning of the experiment. The length of this spin up is 2000 years.

Using these initial conditions, we then perform our experiments for the Holocene, the Eemian and MIS 11. For the Holocene, we perform ~~four~~five experiments to investigate the role of the

1 various forcings in the interglacial carbon cycle: (1) ~~a model~~ HOL_ORB, an experiment
2 ~~containing where~~ neither peat accumulation, nor CaCO₃ sedimentation, nor anthropogenic
3 land use emissions. ~~This experiment is purely driven by, but only~~ orbital forcing. ~~We denote it~~
4 HOL_ORB, is considered; (2) ~~A~~ HOL_PEAT, an experiment ~~containing where~~ peat
5 accumulation, but neither CaCO₃ sedimentation nor anthropogenic land use emissions,
6 ~~denoted HOL_PEAT~~, are considered. (3) ~~A~~ HOL_NAT, an experiment ~~using where~~ all of the
7 natural forcing mechanisms, i.e., peat accumulation and CaCO₃ sedimentation, ~~denoted~~
8 HOL_NAT, are considered; (4) ~~The same setup~~ HOL_MPT, an experiment where all the
9 ~~natural forcing mechanisms are considered (as in HOL NAT), but where the minimum~~
10 ~~peatland area estimate from Kleinen et al. (2012) is used;~~ (5) HOL_ANT, which uses again
11 ~~the same forcings~~ as HOL_NAT, but also including anthropogenic carbon emissions, ~~denoted~~
12 HOL_ALL. ~~Experiments for~~ HOL_MPT, HOL_ANT, and HOL_ALL ~~follow the setup~~
13 HOL_NAT with appropriate initial conditions, assuming that anthropogenic land use did not
14 play a role then. In addition, we performed an experiment for each interglacial, where we
15 disabled the slow forcing factors as in set up two experiments analogous to HOL_ORB and
16 HOL_NAT, with adapted initial conditions. The characteristics of all experiments are
17 summarised in Table 1.

18 All experiments are driven by orbital changes (Berger, 1978). ~~The~~ In the experiments that
19 consider variable shallow-water CaCO₃ accumulation rates (HOL_NAT, HOL_ALL, MPT,
20 HOL_ANT, EEM_NAT, and MIS11_NAT) ~~also require~~ sea level changes, ~~as described in~~
21 Sect. 2.4, and in experiment HOL_ALL anthropogenic CO₂ emissions from land use
22 changes and fossil fuel burning are ~~provided as an additional forcing~~ prescribed, as described
23 in Sect. 2.4.

24 3 Results

25 3.1 Holocene

26 The ~~model experiment~~ HOL_ORB, atmospheric carbon dioxide concentrations resulting from
27 ~~the Holocene experiments are shown in Fig. 2a, as well as ice core data for comparison. In~~
28 model experiment HOL_ORB (without peat accumulation and CaCO₃ sedimentation in
29 shallow waters, ~~would correspond to the carbon cycle implemented in most earth system~~
30 models (ESM), i.e., a carbon cycle not taking into account slow processes of the C cycle. As
31

1 ~~shown in Fig. 3a, this model setup leads to a small decrease in CO₂ (-5 ppm) atmospheric~~
2 ~~CO₂ decreases by ~5 ppm~~ over the first 2000 years, followed by constant CO₂ for the
3 remainder of the experiment. ~~The modelled terrestrial biomass carbon decreases by about 30~~
4 ~~PgC during this time, as shown in Fig. 4a, while the soil carbon increases by a similar~~
5 ~~amount. Overall the conventional carbon cycle setup HOL_ORB would only lead to minor~~
6 ~~changes in atmospheric CO₂, especially missing the increase in atmospheric CO₂ by 20 ppm~~
7 ~~shown in the ice core record for 6 ka BP to 0 ka.~~

8 ~~The results from model (Fig. 2a, blue line). In experiment HOL_PEAT, including (which~~
9 ~~includes carbon accumulation in boreal peatlands but excludingexcludes CaCO₃ accumulation~~
10 ~~in shallow waters, is shown in green in Fig. 3a. It exhibits an) atmospheric CO₂~~
11 ~~decreasedecreases~~ by 25 ppm at 0 ka BP relative to 8 ka BP, which is explained by the uptake
12 of 320 PgC by peatland growth. Yu et al. (2010) estimate a total accumulation of 550 PgC in
13 northern peatlands from the LGM to the present, which indicates that the peat accumulation is
14 reasonable in our model, considering the time frame of our experiment.

15 ~~The results from our8 to 0 ka BP (Fig. 2a, magenta line). In experiment HOL_NAT, including~~
16 ~~(which includes carbon storage in boreal peatlands and shallow water CaCO₃ accumulation;~~
17 ~~are shown as a magenta line in Fig. 3a. Here, the trajectory of) atmospheric CO₂ closely~~
18 follows the ice core measurements ~~rather closely~~ until about 3 ka BP, (Fig. 2a, black line).
19 Between 8 ka and 6 ka BP, the model overestimates CO₂ by up to 5 ppm, while it
20 underestimates atmospheric CO₂ after 4 ka BP, ~~with the.~~ The discrepancy ~~risinggrows~~ as the
21 model gets closer to the present. Atmospheric CO₂ stays constant at ~~268267~~ ppm after 4 ka
22 BP in this experiment.

23 ~~Finally, the results from HOL_ALL, i.e., a model setup similar to HOL_ In experiment~~
24 ~~HOL_MPT (as HOL_NAT, but with the minimum peatland area) atmospheric CO₂ increases~~
25 ~~slightly stronger than in HOL_NAT during the early Holocene and keeps increasing until 2.5~~
26 ~~ka BP, after which it stays constant at 273 ppm CO₂ (Fig. 2a, cyan line). Finally, in~~
27 ~~HOL_ANT (as HOL_NAT but with anthropogenic emissions of CO₂ from land use changes~~
28 ~~and fossil fuel use considered, are shown in black in Fig. emissions)3a. Here the atmospheric~~
29 CO₂ is ~~very~~ similar to CO₂ in HOL_NAT until about 4 ka BP, ~~and similar to HOL_MPT until~~
30 ~~2.5 ka BP.~~ after which HOL_ALLANT displays a continued increase in CO₂, in line with ice
31 core CO₂: (Fig 2a, green line). The CO₂ trajectory stays relatively close to the measurements

1 over the entire time frame of ~~the experiment, with a maximum deviation of about 8 ppm CO₂~~
2 ~~at 1.5 ka BP.~~ HOL_ANT.

3 Biomass carbon, shown in Fig. 4a, stays nearly constant at 550 PgC over the entire simulation
4 period of experiment HOL_NAT, in contrast to the decrease observed for HOL_ORB. For the
5 first 5 ka, biomass carbon in HOL_ALL is very similar to HOL_NAT, but after 2.5 ka BP it
6 increases driven by the increase in atmospheric CO₂, and reaches more than 600 PgC at the
7 end of the experiment. Soil carbon stocks, shown in Fig. 4b, initially are 110 PgC lower in
8 HOL_NAT and HOL_ALL than in HOL_ORB. This difference is due to the fact that some
9 areas, especially in the high latitudes rich in soil C, are set aside as peatlands and therefore not
10 available for mineral soil carbon storage. In experiment HOL_NAT the soil carbon stock
11 increases from an initial 1325 PgC to about 1400 PgC at 0 ka. The evolution in HOL_ALL is
12 very similar for the first 5 ka, but after 3 ka BP soil carbon increases more than in HOL_NAT
13 due to higher CO₂, and reaches a maximum of 1425 PgC at the end of the experiment.

14 Figure 3b shows the carbon 13 isotope of CO₂, $\delta^{13}\text{CO}_2$ from experiment HOL_ALL (black) in
15 comparison to ice core measurements from EDC (Elsig et al., 2009) (red). Modelled $\delta^{13}\text{CO}_2$
16 mostly stays within the range of the error bars before 4.5 ka BP, and only after 3 ka BP is the
17 model $\delta^{13}\text{CO}_2$ consistently above the range of the error bars. Overall, the model setup
18 HOL_ALL therefore captures changes in atmospheric CO₂ as measured from Antarctic ice
19 cores reasonably well, though there is a divergence in $\delta^{13}\text{CO}_2$ after 3 ka BP.

20 Figure 4e shows the cumulative carbon uptake by peatlands in experiments HOL_NAT and
21 HOL_ALL. Carbon storage in peatlands increases nearly linearly over the entire time of the
22 experiment (in fact, carbon uptake only saturates after several tens of ka), up to a total of 330
23 PgC accumulated at the end of experiment HOL_ALL, while HOL_~~ANT~~. The disaggregated net
24 carbon fluxes leading to these trajectories in CO₂ are shown in Fig. 3. For clarity we have
25 smoothed the plots using a Gaussian filter (length 1000 a, stronger weighting in the center
26 following a Gaussian distribution). Fig. 3a shows the geological and anthropogenic C flux to
27 the atmosphere, i.e., the sum of volcanic outgassing, weathering and anthropogenic fluxes,
28 while Fig. 3b shows the net land – atmosphere carbon flux and Fig. 3c shows the net ocean –
29 atmosphere C flux. Since the volcanic input is constant in time, the changes shown in Fig. 3a
30 mainly reflect changes in weathering, with the exception of experiment HOL_ANT, where the
31 bulk of the changes is due to anthropogenic emissions. In experiment HOL_ORB, shown in
32 blue, weathering takes up slightly more carbon in the early than in the late Holocene. This

1 carbon uptake by weathering is compensated by C emissions from the ocean, as shown in Fig.
2 3c, while the land is carbon neutral. The slight decrease in atmospheric CO₂ displayed during
3 the first 2000 years of experiment HOL_ORB therefore is the result of the slightly stronger
4 weathering during the early Holocene. Although the land is carbon neutral overall, a shift in
5 carbon allocation becomes apparent in Fig. 4, where the evolutions of the land C pools are
6 shown. In experiment HOL_ORB, vegetation carbon decreases by 30 PgC over the time of
7 the experiment (Fig. 4a), while soil carbon increases by a similar amount (Fig. 4b).

8 The decrease in CO₂ in experiment HOL_PEAT (not shown) accumulated 320 PgC. The
9 difference (magenta) is caused by an uptake of carbon by the land surface. In Fig. 3b, a more or
10 less constant carbon uptake flux of about -0.03 PgC a⁻¹ is shown, caused by the accumulation
11 of 310 PgC of peat (Fig. 4c). Vegetation loses about 80 PgC (Fig. 4a), while soils lose about
12 20 PgC (Fig. 4b), due to the lower concentration of atmospheric CO₂, lessening the impact of
13 the peat accumulation. Note that the overall soil C pool is decreased in comparison to
14 HOL_ORB for all experiments considering peat carbon accumulation since the area available
15 for carbon storage in mineral soils is decreased due to the consideration of peatlands. In
16 addition, C is released from the ocean (Fig. 3c), partially compensating the carbon uptake.

17 Carbon fluxes in experiments HOL_NAT and HOL_MPT are very similar. They also display
18 a slight decrease in the atmospheric weathering flux (Fig. 3a), though less pronounced than in
19 experiments HOL_ORB and HOL_PEAT due to the higher concentration in atmospheric CO₂
20 later in the Holocene. The uptake of carbon by the land, shown in Fig. 3b, is higher than in
21 experiment HOL_PEAT, due to higher CO₂ concentrations, and displays a maximum at the
22 beginning of the experiment. Soil carbon (Fig. 4b) increases over the entire experiment, while
23 vegetation carbon (Fig. 4a) stays constant. Peat accumulates (Fig. 4c) in both experiments,
24 though the total accumulation is different: 340 PgC in HOL_NAT, but only 250 PgC in
25 HOL_MPT. The increase in atmospheric CO₂ during the early Holocene is driven by the
26 release of carbon from the ocean in these experiments, as shown in Fig. 3c. This increase in C
27 release from the ocean relative to the previous experiments is caused by the release of CO₂
28 during the formation of the CaCO₃ that accumulates in shallow waters, especially in coral
29 reefs, as shown in Fig. 5b. The coral CaCO₃ accumulation flux is 28 Tmol a⁻¹ during the early
30 Holocene due to the large change in sea level (Fig. 5a). It decreases to 18 Tmol a⁻¹ at 4 ka BP
31 and stays constant thereafter due to the constant sea level.

1 Carbon fluxes in experiment HOL ANT cannot be shown fully in Fig. 3 since their
2 magnitude becomes substantially larger than the natural carbon fluxes described for the
3 previous experiments towards the late Holocene. We have therefore added inset figures
4 showing fluxes from HOL NAT and HOL ANT for the last 3.5 ka of the experiments. The
5 plots in the inset Figures have been smoothed using a shorter filter length of 50 years since
6 the long filtering used previously hides the substantial changes induced by industrial CO₂
7 emissions after 1765 AD. The geological and anthropogenic C flux to the atmosphere peaks at
8 2 PgC a⁻¹ in 0 BP, due to the large anthropogenic C emissions. During the early Holocene the
9 anthropogenic flux is substantially smaller, though. Here, it lessens the impact of weathering
10 and slightly increases the geological and anthropogenic C flux to the atmosphere in
11 comparison to HOL NAT, as shown in Fig. 3a. The land – atmosphere flux is very similar to
12 experiment HOL NAT until 3 ka BP, the land generally takes up carbon due to peat
13 accumulation. After 3 ka BP, the land – atmosphere flux becomes more negative than in the
14 other experiments, the land C uptake increases due to CO₂ fertilisation effect of CO₂ on
15 photosynthesis, and reaches a minimum of -0.6 PgC a⁻¹ at the end of the experiment (Fig. 3b,
16 inset Figure). This is reflected in the land C pools (Fig. 4), which keep increasing throughout
17 experiment HOL ANT. The ocean – atmosphere flux in experiment HOL ANT is similar to
18 those in experiments HOL NAT and HOL MPT, though slightly smaller, until 3 ka BP,
19 when it starts deviating from the other experiments (Fig. 3c). The flux becomes negative after
20 75 BP (1875 AD) (Fig. 3c, inset Figure), i.e., the ocean switches from being a sink for carbon
21 to a source.

22 The increase in atmospheric CO₂ during the early Holocene in experiments HOL NAT,
23 HOL MPT, and HOL ANT is due to the strongly positive ocean – atmosphere carbon flux,
24 caused by the accumulation of CaCO₃ in shallow waters. Sea level initially rises fast (see Fig.
25 5a), reaching stable levels around 5 ka BP. The shallow water CaCO₃ accumulation rate,
26 shown in Fig. 5b, varies with the rate of sea level change. The rate of sea level change is
27 highest early during the Holocene, about 2 mm a⁻¹, leading to a CaCO₃
28 sedimentation accumulation of about 27 Tmol a⁻¹. Sea level stabilises later in the Holocene,
29 leading to reducing shallow-water CaCO₃ sedimentation of accumulation to about 15 Tmol
30 a⁻¹ in all three experiments. The formation of CaCO₃ in experiments HOL NAT,
31 HOL MPT, and HOL ANT leads to a reduction in mean ocean alkalinity, as shown in Fig.
32 5c. Over the course of the Holocene, the mean ocean alkalinity is reduced by about 10% in
33 the experiments where shallow water CaCO₃ accumulation is considered.

The evolution of atmospheric $\delta^{13}\text{CO}_2$ in experiments HOL NAT, HOL MPT, and HOL ANT is shown in Fig. 2b. The plots are smoothed using a Gaussian filter for clarity. To enable comparison of the trends in $\delta^{13}\text{CO}_2$ to ice core data, we also added a small constant offset to these results (-0.0712‰ for HOL NAT and HOL ANT, and -0.04‰ for HOL MPT). This offset became necessary because the model displays a small drift in $\delta^{13}\text{CO}_2$ when coupling the interactive CO_2 from CLIMBER and the LPJ land carbon cycle after the spinup of LPJ. Modelled $\delta^{13}\text{CO}_2$ increases from -6.4‰ to -6.2‰ for experiments HOL NAT and HOL MPT, while HOL ANT displays an increase from -6.4‰ to -6.3‰ at 3 ka BP and decreases again thereafter. The Schmitt et al. (2012) Monte Carlo average of $\delta^{13}\text{CO}_2$ increases from -6.4‰ to -6.33‰ at 6 ka BP and slowly decreases again thereafter. The model trajectories from HOL NAT and HOL MPT stay within the 1σ Monte Carlo uncertainty range until 4.8 ka BP and leave the 2σ uncertainty range after 3.9 ka BP. For experiment HOL ANT, the trajectory of $\delta^{13}\text{CO}_2$ stays within the 2σ uncertainty range for the entire experiment. Considering the measurement data directly, as shown by the error bars in Fig. 2b, the trajectories from experiments HOL NAT and HOL MPT leave the range of the error bars at 2.5 ka BP, while HOL ANT remains within the range of the error bars for most of the measurements until the end of the experiment.

Kommentar [TK20]: Reviewer 2, , minor 15, minor 18, minor 19

~~3.2 Eemian~~

~~We consider the full natural setup of the model for the Eemian in experiment EEM_NAT, similar to experiment HOL_NAT. In Fig. 6 we show atmospheric CO_2 and $\delta^{13}\text{CO}_2$ as simulated by the model in comparison to the ice core data from Schneider et al. (2013). Modelled atmospheric CO_2 is generally within the range spanned by the error bars of the measurements, with few exceptions. Similarly, modelled $\delta^{13}\text{CO}_2$ is within the range of the error bars for most of the measurements.~~

~~**3.2 In contrast, experiment EEM_ORB, shown as a blue line in Fig. 6a, is not able to explain the CO_2 trajectory as reconstructed from the ice core. Here,**~~

~~The CO_2 concentrations from the Eemian experiments EEM_ORB and EEM_NAT are shown in Fig. 6a, together with the ice core data. Experiment EEM_ORB (blue line), is in poor agreement with the ice core data. CO_2 decreases from the initial value of 276 to about 267 ppm CO_2 at 121 ka BP, after which it increases again to 278 ppm at 116 ka BP. While the~~

1 discrepancy in CO₂ between experiment EEM_ORB and the ice core data is not excessive, the
2 fit of experiment EEM_NAT to the data is substantially better. The slow natural processes we
3 consider therefore seem to be required to explain the evolution of CO₂ during the Eemian
4 experiment EEM_NAT, in contrast, modelled atmospheric CO₂ changes relatively little for
5 the entire experiment and is generally within the range spanned by the error bars of the
6 measurements, with exceptions only around 119.5 ka BP, where 3 data points show lower
7 CO₂ concentrations than modelled. $\delta^{13}\text{CO}_2$ for EEM_NAT, smoothed by a Gaussian filter and
8 offset by -0.0647‰ as for the Holocene experiments, is shown in Fig. 6b. Modelled $\delta^{13}\text{CO}_2$
9 stays within the 1 σ uncertainty of the Monte Carlo estimate from Scheider et al. (2013) for the
10 entire time of the experiment.

Kommentar [TK21]: Reviewer 2,
minor 18

11 The terrestrial biomass (Fig. 7a) reaches a maximum of about 600 PgC early in experiment
12 EEM_NAT at 124 ka BP. It decreases thereafter and reaches a minimum value of ~490 PgC
13 at the end of the experiment at 116 ka BP. Biomass carbon in experiment EEM_ORB follows
14 a very similar trajectory. Soil carbon in EEM_NAT (Fig. 7b) increases from an initial value of
15 1325 PgC to about 1400 PgC at 121 ka BP and decreases thereafter, reaching 1225 PgC at
16 116 ka BP. The evolution in EEM_ORB is similar, though offset by about 90 PgC, again due
17 to the larger area available for mineral soil carbon when no peatlands are considered. The
18 carbon storage in peatlands, shown in Fig. 7c for EEM_NAT, increases linearly during the
19 The disaggregated net carbon fluxes leading to these CO₂ trajectories are shown in Fig. 7.
20 Weathering is very strong during the early Eemian, leading to a geological C flux of -0.24
21 PgC a⁻¹ for 126 ka BP, shown in Fig. 7a. Weathering then decreases, allowing the geological
22 C flux to increase to -0.205 PgC a⁻¹ at 116 ka BP in both experiments. While the geological C
23 flux is similar in both experiments, the other C fluxes are substantially different. For
24 experiment EEM_ORB, the land is generally carbon-neutral during the early Eemian. The
25 land – atmosphere C flux is close to zero until 121.5 ka BP (see Fig. 7b). Later in the Eemian,
26 the land – atmosphere C flux increases to a maximum of 0.05 PgC a⁻¹ at 117 ka BP. For the
27 early Eemian this is due to counteracting contributions from the land carbon pools: vegetation
28 carbon decreases continually from 126 ka BP to 116 ka BP, with an initially slow rate of
29 decrease that increases after 123 ka BP (see Fig. 8a). Soil carbon, in contrast, increases from
30 1415 PgC at 126 ka BP to 1480 PgC at 120.5 ka BP, after which it decreases to 1350 PgC at
31 116 ka BP (Fig. 8b). The ocean – atmosphere carbon flux, on the other hand, is initially at
32 0.23 PgC a⁻¹ and decreases to a minimum of 0.15 PgC a⁻¹ at 117 ka BP. Therefore the initial
33 strong carbon uptake through weathering is not completely compensated by marine C fluxes,

Kommentar [TK22]: Reviewer 1,
major 3; Reviewer 2, major 4

1 leading to the modelled reduction in atmospheric CO₂ between 126 and 121.5 ka BP. After
2 this time, the land loses carbon, allowing a compensation of the (reduced) weathering flux and
3 leading to an increase in atmospheric CO₂.

4 In experiment EEM_NAT, on the other hand, the land – atmosphere C flux is negative until
5 119.5 ka BP, increasing from -0.095 PgC a⁻¹ at 126 ka BP to 0.04 PgC a⁻¹ at 116 ka (Fig. 7b).
6 While the vegetation and soil carbon pools behave in a generally similar way to experiment
7 EEM_ORB (Fig. 8a, b), the model accumulates 445 PgC of peat carbon (Fig. 8c), resulting in
8 the generally negative land – atmosphere flux. The ocean – atmosphere flux initially is
9 substantially higher than in EEM_ORB, with an initial flux of 0.33 PgC a⁻¹, that decreases
10 over time to 0.17 PgC a⁻¹. Eemian as well, until about 440 PgC are accumulated at the end of
11 the experiment.

12 The sea level forcing, shown in Fig. 8a9a, is stable early during the experiment and decreases
13 after 121 ka BP. Therefore shallow-water CaCO₃ accumulation (Fig. 8b) is accumulated at
14 a rate of ~20 Tmol a⁻¹ during the early Eemian, (Fig. 9b), lower than during the early
15 Holocene. It decreases to about zero at 119 ka and stays at this level thereafter. This increases
16 the ocean – atmosphere flux in comparison to experiment EEM_ORB, thus releasing carbon
17 to the atmosphere, which compensates the peat carbon uptake and the strong weathering flux
18 during the early Eemian.

19 3.3 MIS 11

20 For MIS 11, the agreement between the modelled atmospheric CO₂ concentrations in
21 MIS11_NAT and the ice core reconstruction is not as good as for the other two interglacials.
22 As shown in Fig. 910, modelled CO₂ in experiment MIS11_NAT increases initially from 271
23 ppm CO₂ to about 290 ppm at 412 ka BP. It declines thereafter to about 250 ppm CO₂ at 395
24 ka BP, after which CO₂ varies much less. Setup MIS11_ORB, on the other hand, shows a
25 slowly decreasing trend in CO₂, from the initial 271 ppm CO₂ to slightly less than 260 ppm at
26 380 ka BP, with only little variation about this trend.

27 The initial increase in CO₂ is slower in the ice core data than in MIS11_NAT. CO₂ increases
28 to about 285 ppm at 407 ka BP. Measured CO₂ decreases strongly after 398 ka BP, until 250
29 ppm CO₂ are reached at 390 ka BP. Therefore the model setup MIS11_NAT overestimates
30 the initial increase in CO₂, and the peak in CO₂ is reached about 5 ka earlier than in the ice
31 core data. Similarly, the decrease after the peak in CO₂ also occurs earlier in the model than in

1 the ice core data. Nonetheless, the overall CO₂ trajectory, with an initial increase in CO₂
2 between 420 ka and 405 ka BP, followed by a decrease by about 25-30 ppm and a
3 stabilisation of CO₂ after 395 ka BP is captured by MIS11_NAT, though the timing is not
4 exactly the same as in the ice core data. MIS11_ORB, on the other hand, does not at all
5 follow the ice core CO₂ data. ~~For the interested reader, we show the $\delta^{13}\text{CO}_2$ for experiment~~
6 ~~MIS11 NAT in Fig. 10b, though no ice core data is available for comparison. $\delta^{13}\text{CO}_2$ starts~~
7 ~~out at -6.5‰, with a slowly decreasing trend until 405 ka BP. Afterwards it decreases quickly~~
8 ~~to -6.55‰ at 398 ka BP. It then slightly increases again until 390 ka BP, after which it~~
9 ~~decreases further until -6.57‰ are reached at 380 ka BP.~~

10 ~~The land carbon pools display substantially more variability in MIS11_NAT than in~~
11 ~~MIS11_ORB, shown in Fig. 10a and b. Biomass carbon (Fig. 10a) increases strongly in~~
12 ~~MIS11_NAT. The geological C flux, shown in Fig. 11a, is very similar in both experiments;~~
13 ~~changes in MIS11 NAT only present a slightly larger amplitude than MIS11 ORB due to the~~
14 ~~higher CO₂ concentrations. It decreases from -0.208 PgC a⁻¹ at 420 ka BP to -0.233 PgC a⁻¹ at~~
15 ~~410 ka BP in MIS11 NAT, reflecting increases in weathering. Subsequently it increases to~~
16 ~~the initial value at 395 ka BP, followed by a small decrease and further increase to a final~~
17 ~~value of -0.204 PgC a⁻¹. The land – atmosphere flux in experiment MIS11 ORB fluctuates~~
18 ~~around zero (Fig. 11b), with increases in vegetation carbon (Fig. 12a) compensated by~~
19 ~~decreases in soil carbon (Fig. 12b) and vice-versa. In MIS11 NAT, the net land – atmosphere~~
20 ~~flux is more negative due to the accumulation of peat carbon. It is -0.09 PgC a⁻¹ initially and~~
21 ~~increases to zero at 400 ka BP, with a subsequent decrease to -0.035 PgC a⁻¹ at 395 ka BP and~~
22 ~~an increase to -0.01 PgC a⁻¹ at 380 ka BP. This is due to the changes in the land carbon pools:~~
23 ~~vegetation carbon (Fig. 11a) increases strongly, until a maximum value of about 630 PgC is~~
24 ~~reached at 412 ka BP. Carbon storage decreases afterwards, until a minimum of 480 PgC is~~
25 ~~reached at 395 ka BP, with only small changes afterwards. Similarly, soil carbon increases~~
26 ~~early in MIS11_NAT from an initial value of 1350 to about 1425 PgC at 414 ka BP. It then~~
27 ~~stays constant until 403 ka BP, when it starts ~~decreasing to~~ strongly ~~decrease~~.~~ After 395 ka BP
28 soil carbon stays constant at 1345 PgC. ~~In contrast, the variations in biomass and soil carbon~~
29 ~~are much less pronounced in experiment MIS11_ORB. Biomass carbon increases from 540 to~~
30 ~~560 PgC early in MIS 11, then decreases again to 515 PgC at 395 ka BP, and changes little~~
31 ~~afterwards. Soil carbon, on the other hand, varies between 1490 and 1445 PgC during the~~
32 ~~entire time frame of the experiment. Peat accumulation in MIS11_NAT (Fig. 10e) once~~

again 11c) increases nearly linearly between 420 ka BP and 398 ka BP. After 398 ka BP the rate of increase decreases slightly due to the lower atmospheric CO₂ concentration.

The net ocean – atmosphere flux, shown in Fig. 11c is nearly constant at 0.21 PgC a⁻¹ in experiment MIS11_ORB. In experiment MIS11_NAT, on the other hand, the flux is initially at 0.313 PgC a⁻¹, then decreases to about 0.207 PgC a⁻¹ at 400 ka BP, to increase to 0.273 PgC a⁻¹ over the next 10 ka. These changes in the ocean – atmosphere C flux are mainly driven by changes in the CaCO₃ accumulation in shallow waters. During the first 13 ka of MIS 11 sea level increases from -20 m to near zero (Fig. 11a12a). It starts decreasing then decreases again at 407 ka BP, but stabilises at -15 m after 395 ka BP. This sea level trajectory is reflected in the CaCO₃ accumulation flux, shown in Fig. 11b12b: the initial fast rise in sea level leads to an accumulation rate of up to 29 Tmol a⁻¹, with a correspondingly high CO₂ release to the atmosphere, which declines between 413 and 400 ka BP, when the accumulation rate is zero due to the decrease in sea level. With the slowing rate of sea level decrease, sedimentation increases again after 396 ka BP and reaches values of about 15 Tmol a⁻¹ again at 390 ka BP.

4 Discussion

From our results for the Holocene carbon cycle, it becomes quite clear that all of the forcings and processes considered taken together deliver the best match to the ice core CO₂ data. The model setup HOL_ORB, i.e., a carbon cycle setup without anthropogenic CO₂ emissions ~~enor~~ slow natural processes, leads to a more or less constant CO₂ trajectory, where the carbon uptake by weathering is compensated by a carbon release from the ocean, while the land is generally carbon-neutral. The consideration of peat accumulation by itself in HOL_PEAT leads to a decrease in atmospheric carbon dioxide- due to a large carbon uptake by the land, which is partially compensated by additional carbon release from the ocean, in comparison to HOL_ORB. The additional consideration of CO₂ emissions from CaCO₃ shallow water sedimentation in HOL_NAT then leads to an increase in atmospheric CO₂, not just compensating the C uptake by peatlands, but also releasing additional CO₂ to the atmosphere. ~~From the difference between experiments HOL_NAT and HOL_ALL it becomes clear~~

According to the “early anthropogenic hypothesis” (Ruddiman 2003; 2013), anthropogenic land emissions related to land use from early agriculture strongly influenced climate already in the early Holocene. The present study does not aim at either validating or falsifying this

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Kommentar [TK26]: Reviewer 2,
major 1

1 hypothesis. However, it becomes clear from the difference between experiments HOL NAT
2 and HOL ANT that anthropogenic CO₂ emissions from land use changes only make a
3 significant difference to atmospheric CO₂ after about 3 ka BP, which would shift to an even
4 later time with smaller peat carbon accumulation (as in HOL MPT). Anthropogenic
5 emissions therefore cannot explain the 10 ppm rise in CO₂ observed in ice cores between 8
6 and 4 ka BP. For the earlier Holocene, CO₂ emissions from shallow water CaCO₃
7 sedimentation are required instead. The continued rise in CO₂ after 2.5 ka BP, on the other
8 hand, can only be explained if anthropogenic emissions are accounted for as well.

9 While our assessment contains several uncertainties. The modelled peat accumulation
10 rates compare well to site data (Kleinen et al., 2012), and the overall peatland carbon
11 accumulation fits well to observed peat carbon stocks (Yu et al., 2010), which are relatively
12 well constrained. What is less well constrained is the peatland carbon accumulation history,
13 and our modelled peatland carbon accumulation trajectory may not reflect the actual
14 accumulation history. What is also not very well constrained is the areal extent of peatlands
15 (Kleinen et al., 2012), as well as its temporal development. MacDonald et al. (2006) and Yu
16 et al. (2010) show high rates of peatland initiation between 11 and 9 ka BP, possibly caused
17 by the high northern high-latitude insolation during this time. Northern peatland initiation
18 dates decrease in number after this time, but remain significant. In our model peatland area
19 changes little after 8 ka BP, one of the reasons for the nearly linear accumulation of carbon in
20 peatlands. We also neglect the small remains of the Laurentide ice sheet remaining at 8 ka BP,
21 since we have shown that its influence is small (Kleinen et al., 2012). The change in peatland
22 area over the time of the model simulation may therefore be underestimated by our model,
23 which would modify the trajectory of peat carbon accumulation to a trajectory where less
24 carbon is accumulated earlier in the Holocene and more C is accumulated later.

Kommentar [TK27]: Reviewer 1,
major 1 and 2

25 Modelled vegetation changes in our model compare well against tree cover reconstructions
26 from Eurasia (Kleinen et al., 2011). They are also similar to vegetation changes obtained with
27 other models, for example CLIMBA and Bern3D, as published by Brovkin et al. (2016).
28 However, it is not possible to validate the modelled changes in terrestrial carbon storage since
29 no direct proxy exists for carbon stored in terrestrial ecosystems. The CO₂ fertilisation effect
30 displayed by CLIMBER2-LPJ as well as other DGVMs, which leads to increases in biomass
31 with increasing CO₂, seems well-understood at the leaf level (De Kauwe et al., 2014), but

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Kommentar [TK28]: Reviewer 1,
major 1

1 ~~may be overestimated in models because constraining mechanisms such as nutrient limitation~~
2 ~~are not taken into account (Reich et al., 2014).~~

3 ~~Our~~ CaCO₃ accumulation model seems to capture the late Holocene sedimentation, ~~within~~
4 good agreement ~~to~~with Milliman (1993). ~~Nonetheless~~ the increase in accumulation rate due
5 to the rate of sea level rise during the earlier Holocene is relatively uncertain. This is due to
6 uncertainties in the parameterisation, as well as uncertainties in the rate of sea level rise.
7 While both are plausible, there is considerable uncertainty with respect to magnitude and
8 timing of the CO₂ emissions from CaCO₃ formation. Previous assessments ~~nevertheless~~ agree,
9 ~~though~~, that coral growth was stronger in the early Holocene (Ryan et al., 2001; Vecsei and
10 Berger, 2004). ~~The change in sea level we use as a model forcing agrees well with sea level~~
11 ~~reconstructions for the Holocene, though sea level stabilises up to 2 ka earlier in our model~~
12 ~~than it does in reconstructions (Masson-Delmotte et al., 2013, Fig. 5.17 f). A later stabilisation~~
13 ~~of the sea level, as in palaeoclimatic archives, would lead to a prolongation of the relatively~~
14 ~~larger emissions from CaCO₃ formation we model for the early Holocene. Therefore we may~~
15 ~~underestimate the emissions from shallow water CaCO₃ accumulation.~~

Kommentar [TK29]: Reviewer 2,
major 6

16 ~~Finally, the modelled trajectory of $\delta^{13}\text{CO}_2$ for the Holocene has relatively high values~~
17 ~~between 4 ka BP and stays well within the present, as shown in Fig. 4b. These values are~~
18 ~~outside the range of the error bars estimated by Elsig et al. (2009). This result can be~~
19 ~~explained in three different ways: (a) Elsig et al. might have underestimated the true σ~~
20 ~~uncertainty, (b) we may have underestimated the $\delta^{13}\text{CO}_2$ band of the Monte Carlo based~~
21 ~~uncertainty assessment by Schmitt et al. (2012), if one considers experiment HOL ANT,~~
22 ~~while the other experiments leave this range. While the general dynamics of ^{13}C seem to be~~
23 ~~captured well by the LPJ model (Scholze et al., 2003), there is some uncertainty with regard~~
24 ~~to the ^{13}C changes induced by the accumulation of peat, and (c) we may require an unknown~~
25 ~~additional source of isotopically depleted carbon to explain the trajectory of $\delta^{13}\text{CO}_2$. This~~
26 ~~latter explanation has been favoured by proponents of large anthropogenic emissions from~~
27 ~~land-use changes, since CO₂ released from the biosphere would have such a depleted isotopic~~
28 ~~signature (Ruddiman et al., 2011). At 262 PgC cumulative emissions from land-use changes,~~
29 ~~the scenario adopted here already assumes larger fluxes than other recent estimates. Stocker et~~
30 ~~al. (2014), for example, estimate the cumulative emissions by 2004 at 243 PgC. Besides,~~
31 ~~judging from Fig. 4b, the modelled atmospheric $\delta^{13}\text{CO}_2$ is higher than the measurements after~~
32 ~~about 4.5 ka BP, earlier than the bulk of the emissions in the scenario based on Kaplan et al.~~

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1 ~~(2011). Emissions from anthropogenic land use changes therefore do not appear to be a likely~~
2 ~~cause of the mismatch in $\delta^{13}\text{C}$, but we cannot rule out other isotopically depleted sources of~~
3 ~~C, such as methane emissions or the release of carbon from thawing permafrost soils. With~~
4 ~~regard to (b), we. We~~ assume that the carbon uptake by peat accumulation has a similar
5 signature in $\delta^{13}\text{C}$ as the growth of C3 grass. Since photosynthesis in mosses generally follows
6 the C3 pathway, this assumption appears reasonable, and values for $\delta^{13}\text{C}$ in mosses reported
7 in the literature ~~(e.g. Waite and Sack, 2011)~~ are in a similar range as values for other C3
8 vegetation. ~~With regard to (a), finally, there are no reasons to believe (e.g. Waite and Sack,~~
9 ~~2011). However, the cycling of ^{13}C in peatlands seems to be less well understood than in other~~
10 ~~terrestrial systems, which makes the modelled $\delta^{13}\text{C}$ changes induced by peat accumulation~~
11 ~~less certain. In addition we cannot rule out other isotopically depleted sources of C, such as~~
12 ~~methane emissions or the release of carbon from thawing permafrost soils, that measurement~~
13 ~~errors are underestimated by Elsig et al. (2009), forcing us to reject (a) as well. This leaves~~
14 ~~unknown sources of isotopically depleted C as the most likely explanation for the discrepancy~~
15 ~~in $\delta^{13}\text{C}$ we have not accounted for in our model.~~

Kommentar [TK30]: Reviewer 1,
minor 18, minor 19

16 With regard to the evolution of atmospheric CO_2 during the Eemian, the fit between ice core
17 data and model results is clearly better for experiment EEM_NAT than for EEM_ORB. While
18 the model produces an initial decrease followed by an increase for EEM_ORB, EEM_NAT
19 shows a ~~near~~nearly constant CO_2 concentration for the entire period of time we modelled,
20 very close to the measurements ~~by Schneider from Bereiter et al. (20132015)~~. Similarly,
21 modelled $\delta^{13}\text{CO}_2$ is within the ~~error bars~~ 1σ uncertainty range of the ~~ice core~~
22 ~~measurements~~Schneider et al. (2013) Monte Carlo average for most of the time, though the
23 ~~model displays less change in $\delta^{13}\text{CO}_2$ than the MC average.~~ Here the largest uncertainty in our
24 setup again stems from the sea level history, leading to uncertainty with respect to magnitude
25 and timing of CO_2 emissions that result from CaCO_3 sedimentation. However, the sea level
26 forcing we use is similar to reconstructed global mean sea level as shown by Masson-
27 Delmotte et al. (2013, Fig. 5.15 a and b) and should therefore be a reasonable approximation.

Kommentar [TK31]: Reviewer 1,
minor 18

28 In our setup, and with the sea level forcing data we use, the CO_2 emissions from CaCO_3
29 sedimentation counterbalance the weathering-induced decrease in CO_2 shown in setup
30 EEM_ORB for the early Eemian, while carbon uptake by peatlands compensates for the
31 increase in CO_2 modelled in EEM_ORB during the second half of the Eemian.

Kommentar [TK32]: Reviewer 2,
major 6

1 For MIS 11 our model experiment MIS11_NAT displays a qualitatively similar evolution of
2 atmospheric CO₂ as the ice core data, with an initial increase, followed by a decrease during
3 the middle of the interglacial until the CO₂ concentration stabilises for the later part of the
4 interglacial. This leads to a clearly better fit to the ice core measurements than setup
5 MIS11_ORB, which shows a continuous slow decrease ~~in of atmospheric~~ CO₂. Nonetheless
6 there still are discrepancies in the timing and the magnitude of the changes in CO₂ between
7 model and ice core data. This discrepancy is most likely again due to uncertainty in the sea
8 level history that we use to force the model. ~~If the increase in sea level before 410 ka BP were~~
9 ~~slightly less pronounced and the decrease in sea level after 405 ka BP slightly delayed, our~~
10 ~~model results would fit the ice core data even better.~~

11 ~~Carbon uptake by peatlands does not change substantially, neither during any of the~~
12 ~~interglacials, nor between interglacials. In all cases we obtain a more or less linear rise in~~
13 ~~peatland carbon storage.~~ Sea level reconstructions for MIS11 show considerable discrepancies
14 ~~between each other, making an evaluation of the quality of our forcing data very difficult.~~
15 ~~Between the four reconstructions of MIS11 sea level that we considered (Waelbroeck et al.,~~
16 ~~2002; Rohling et al, 2010; Elderfield et al., 2012; Grant et al., 2014), there is a general~~
17 ~~qualitative agreement in timing: Sea level rises between 420 and 405 ka BP, followed by a~~
18 ~~decrease. While the sea level history from Waelbroeck et al. (2002) shows a sustained~~
19 ~~decrease, this decrease ends at about 390 ka BP in the other three reconstructions, followed by~~
20 ~~either a plateau or a slight increase. The four reconstructions furthermore disagree with regard~~
21 ~~to the magnitude of changes, with the sea level highstand at ~405 ka BP ranging from -10m to~~
22 ~~+30m relative to the present. Our forcing trajectory falls well within the range of these~~
23 ~~reconstructions. However, the magnitude and timing of changes in sea level has a large~~
24 ~~impact on the CO₂ emissions from shallow water CaCO₃ accumulation, making the latter~~
25 ~~relatively uncertain. If, for example, the increase in sea level before 410 ka BP were slightly~~
26 ~~less pronounced in our forcing data and the decrease in sea level after 405 ka BP slightly~~
27 ~~delayed, our model results would better fit the ice core data.~~

28 Our study has several other limitations. We imposed anthropogenic emissions from land use
29 changes as a simple flux to the atmosphere without changing the land carbon stocks. This
30 simplification modifies the uptake of carbon by the biosphere and should already be contained
31 in the Kaplan et al. (2011) CO₂ emission estimate, but an inconsistency remains nonetheless.
32 We also neglected the long-term memory of the carbonate compensation response to the

1 release of carbon from the deep ocean and the early interglacial carbon uptake by the
2 terrestrial biosphere during deglaciation. While CLIMBER2-LPJ contains all relevant
3 processes, we did not model this period transiently and our results therefore do not
4 have contain the long-term memory signal ~~in our results.~~ Menviel and Joos (2012) found that
5 these memory effects could be of the order of a few ppm for the Holocene. ~~Furthermore~~
6 ~~we~~ we furthermore assumed that the long-term carbon cycle was in equilibrium in the pre-
7 industrial climate, but this assumption is a simplification as the balance amongbetween
8 carbonate burial, weathering, and volcanic outgassing could be out of equilibrium for other
9 climates. As follows from control simulations without forcings (not shown), these effects can
10 be of the order of few ppm as well. Last, but not least, several other mechanisms that are
11 currently under discussion such as changes in permafrost carbon pools (Schneider von
12 Deimling et al., 2012) or methane hydrate storages (Archer et al., 2009) are not accounted for,
13 as modelling of these processes is still in an early stage and because of the lack of reliable
14 constraints on the amplitude of interglacial changes in these potentially large carbon pools.

15 5 Conclusions

16 We show -- to our best knowledge for the first time -- how the trends in interglacial
17 atmospheric CO₂, as reconstructed from ice cores, can be reproduced relatively well by a
18 climate model with identical forcing parameterisation for three recent interglacials. ~~For~~
19 ~~these~~ These trends in atmospheric CO₂ ~~it is important to account not just for~~ cannot be
20 reproduced well if only the marine and terrestrial carbon cycle components, as implemented
21 in most ~~earth~~ Earth system models (Ciais et al., 2013) ~~), are considered.~~ Instead, it is necessary
22 ~~to also consider~~ the modelled CO₂ change is considerably improved if two slow processes of
23 CO₂ change currently neglected in the most comprehensive carbon cycle models, namely the
24 carbon accumulation in peatlands and the CO₂ release from CaCO₃ ~~formation and~~
25 ~~burial~~ accumulation in shallow waters. ~~This, are accounted for as well. The~~ latter process leads
26 to an increase in atmospheric CO₂ during periods of constant or slowly rising sea level, while
27 the former process leads to a decrease in atmospheric CO₂.

28 For the Holocene, we can explain the rise in atmospheric CO₂ between and 3 ka BP purely by
29 natural forcings, while later in the Holocene, starting at about 3 ka BP, anthropogenic
30 emissions from land use changes and fossil fuel use play an important role. The increase in
31 atmospheric CO₂ during the early Holocene therefore is the result of enhanced shallow-water
32 ~~sedimentation~~ accumulation of CaCO₃ due to rising sea level. For the Eemian, our carbon

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1 ~~cycle~~ model also ~~leads to a satisfactory simulation of~~ yields an atmospheric CO₂, ~~which~~
2 ~~history that~~ is ~~very~~in close ~~to~~agreement with the ice core data. Here the consideration of the
3 slow carbon cycle processes also ~~led~~leads to an improvement over the conventional model.
4 ~~approach that neglects these.~~ For MIS 11, finally, the conventional model setup does not
5 ~~simulate~~produce the changes in CO₂ observed throughout MIS 11, while the model with
6 consideration of the slow forcings can explain the magnitude of changes in atmospheric CO₂,
7 though the timing of changes is slightly different from the ice core data. This discrepancy is
8 possibly due to the sea level forcing history that we use to drive the shallow water CaCO₃
9 accumulation in our model, and which remains uncertain.

Kommentar [TK35]: Editor

10 Despite the uncertainties discussed above, we can draw some robust conclusions with regard
11 to the timing of CO₂ changes. Early during interglacials, when sea level still rises, shallow
12 water accumulation of CaCO₃ and the related CO₂ release is larger than in periods of
13 stagnating or receding sea level. The carbon uptake by peatlands, on the other hand, is a more
14 or less constant forcing factor. This uptake balances the CO₂ emission from CaCO₃
15 precipitation during periods of constant sea level. A rising sea level therefore leads to
16 atmospheric CO₂ increases, while a decline in sea level strongly reduces shallow-water
17 CaCO₃ ~~sedimentation~~accumulation, leading to a reduction in atmospheric CO₂.

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4

1 Table 1: Setup of experiments performed for the Interglacials, including the forcing factors
 2 varied.

Name	Interglacial	Initial CO ₂ [ppm]	Initial $\delta^{13}\text{CO}_2$ [‰]	Initial time [ka BP]	Peat accumulation	Coral CaCO ₃ sedimentation	Anthropogenic land use emissions
HOL_ORB	Holocene	260	-6.4	8	No	No	No
HOL_PEAT	Holocene	260	-6.4	8	Yes	No	No
HOL_NAT	Holocene	260	-6.4	8	Yes	Yes	No
<u>HOL_MPT</u>	<u>Holocene</u>	<u>260</u>	<u>-6.4</u>	<u>8</u>	<u>Minimum</u>	<u>Yes</u>	<u>No</u>
HOL_ ALLAN <u>T</u>	Holocene	260	-6.4	8	Yes	Yes	Yes
EEM_ORB	Eemian	276	-6.7	126	No	No	No
EEM_NAT	Eemian	276	-6.7	126	Yes	Yes	No
MIS11_ORB	MIS11	271	6.5	420	No	No	No
MIS11_NAT	MIS11	271	6.5	420	Yes	Yes	No

3
4

Figure captions

Figure 1: Coral growth modification function. CaCO_3 sedimentation is limited in cases of negative and very fast sea level rise.

Figure 2: Cumulative anthropogenic carbon emissions from land use (Kaplan et al., 2011) (black) and land use and fossil fuel (Meinshausen et al., 2011) (red).

Figure 32: Holocene CO_2 concentration (a) and $\delta^{13}\text{C}$ of CO_2 (b) from EPICA Dome C (red) and Siple Dome, model with all forcings HOL_ALL (black), model without anthropogenic forcing HOL_NAT (magenta), model without anthropogenic, peat and coral forcing HOL_ORB (blue), model without coral and anthropogenic forcing HOL_PEAT (green) experiments and ice core data. $\delta^{13}\text{C}$ CO_2 model results have been smoothed for clarity and offset, as described in the text. Ice core CO_2 data is from Law Dome and EDC as compiled by Bereiter et al. (2015). $\delta^{13}\text{C}$ CO_2 data is from EDC (Elsig et al., 2009), with Monte Carlo average (MC) and uncertainty estimate (MC σ) (Schmitt et al., 2012).

Kommentar [TK36]: Reviewer 2, minor 20

Figure 3: Disaggregation of the net carbon fluxes in the Holocene experiments: geological and anthropogenic flux to atmosphere (a), land – atmosphere flux (b), and ocean – atmosphere flux (c). Plots have been smoothed using a 1000 year Gaussian filter for clarity. Inset Figures show fluxes from HOL_ANT and HOL_NAT with axis scaling appropriate for HOL_ANT and smoothed using a 50 year Gaussian filter.

Figure 4: Land carbon pools in Holocene experiments HOL_ALL, HOL_NAT and HOL_ORB: total biomass/vegetation carbon (a), total non-peat soil carbon (b), and cumulative C uptake by peatlands (c).

Kommentar [TK37]: Reviewer 2, minor 21

Figure 5: Holocene experiment HOL_ALL: sea/Sea level forcing (a) and shallow water CaCO_3 formation/accumulation flux (b). (b) Also contains background CaCO_3 formation from HOL_ORB (blue), and mean ocean alkalinity (c) in the Holocene experiments. Plots are have been smoothed for clarity.

Figure 6: Eemian CO_2 concentration (a) for experiments EEM_NAT (black) and EEM_ORB (blue) and $\delta^{13}\text{C}$ of CO_2 for EEM_NAT (b). Red error bars are CO_2 and $\delta^{13}\text{C}$ from EPICA Dome C.

Figure 7: Land carbon pools in Eemian experiment EEM_NAT (black) and EEM_ORB (blue): total biomass carbon (a), total non-peat soil carbon (b), and cumulative C uptake by peatlands (c).

Figure 8: Eemian experiment EEM_NAT: sea level forcing (a) and shallow water CaCO_3 formation (b). (b) Also contains background CaCO_3 formation from EEM_ORB (blue line). Plots are smoothed for clarity.

Figure 9: MIS11 CO_2 concentration for experiments MIS11_NAT (black) and MIS11_ORB (blue), as well as CO_2 reconstruction from ice core (red) model experiments and ice core data. $\delta^{13}\text{C}$ CO_2 model results have been smoothed for clarity and offset, as described in the text. Ice core CO_2 data is from EDC as compiled by Bereiter et al. (2015). $\delta^{13}\text{C}$ CO_2 data are from EDC (Schneider et al., 2013), with Monte Carlo average (MC) and uncertainty estimate (MC σ).

Kommentar [TK38]: Reviewer 2, minor 20

Figure 107: Disaggregation of the net carbon fluxes in the Eemian experiments: geological flux to atmosphere (a), land – atmosphere flux (b), and ocean – atmosphere flux (c). Plots have been smoothed for clarity.

Formatiert: Kopfzeile

1 Figure 8: Land carbon pools in MIS11 experiment MIS11_NAT (black) and MIS11_ORB (blue); the
2 Eemian experiments: total biomass/vegetation carbon (a), total non-peat soil carbon (b), and
3 cumulative C uptake by peatlands (c).

4 Figure 9: Sea level forcing (a), shallow water CaCO₃ accumulation flux (b), and mean ocean alkalinity
5 (c) in the Eemian experiments. Plots have been smoothed for clarity.

6 Figure 10: MIS 11 CO₂ concentration (a) from model experiments and ice core data, and δ^{13} of CO₂
7 from model experiment MIS11_NAT. δ^{13} CO₂ model results have been smoothed for clarity.
8 Ice core CO₂ data are from EDC and Vostok as compiled by Bereiter et al. (2015). δ^{13} CO₂ from ice
9 cores is not available.

Kommentar [TK39]: Reviewer 2,
minor 20

10 Figure 11: MIS11 experiment MIS11_NAT: seaDisaggregation of the net carbon fluxes in the MIS 11
11 experiments: geological flux to atmosphere (a), land – atmosphere flux (b), and ocean – atmosphere
12 flux (c). Plots have been smoothed for clarity.

13 Figure 12: Land carbon pools in the MIS 11 experiments: total vegetation carbon (a), total non-peat
14 soil carbon (b), and cumulative C uptake by peatlands (c).

15 Figure 13: Sea level forcing (a) and, shallow water CaCO₃ formation (b) accumulation flux (b) Also
16 contains background CaCO₃ formation from MIS11_ORB (blue line). Plots are), and mean ocean
17 alkalinity (c) in the MIS 11 experiments. Plots have been smoothed for clarity.