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 d13CO2, 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 regards 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 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 d13CO2 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 the 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 investigating 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 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 δ13CO2 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 cannot be tackled, since they prescribe δ13CO2 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 δ13CO2 changed slowly over time, these results might only be of limited values, and might follow the δ13CO2 (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 d13C 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 d13C. 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 d13C data as initial conditions for the carbon cycle. Our goal is then to simulate trends in d13CO2, 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 for 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 ppmv (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
6. The records of sea level change, that are important for the shallow-water CaCO3 accumulation need 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 δ 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 δ 13CO2 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 δ 13CO2 data from ice cores yet exist for MIS 11 it would of course be of interest to see the educated guess (simulation results) of δ 13CO2 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 test 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 Gmax and Ik 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 Iz 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

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 δ13 CO2 data: I suggest to show the Monte-Carlo-based spline through all available δ13CO2 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)


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 δ13 CO2 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 δ13CO2 data.

We will extend the discussion of model results and data.

We have extended the discussion of our CO2 and d13CO2 results (page 14, line 11-14).

19. Discussion: As explanation (a) of the misfit to the Holocene δ13 CO2 data it is suggested that Elsig underestimates the true uncertainty. By using the spline in δ13 CO2 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 d13CO2 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 d13C 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 extent 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 HOLPEAT 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.
Carbon Interglacial carbon cycle dynamics during recent interglacials: the Holocene, the Eemian and MIS 11

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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 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 model experiments show 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 consistently using an identical model setup.

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

Investigations of the Holocene trend in CO₂ can be classified into two basic approaches: an inverse modelling approach, and a forward or process-based modelling approach. The inverse modelling approach takes the ice core record of CO₂ and δ^{13}CO₂ as a starting point and aims to derive the sources and sinks of CO₂ from this record, while the forward modelling approach starts from the carbon cycle processes and aims to determine a CO₂ trajectory from combinations of these.

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

Using the forward modelling approach, Ridgwell et al. (2003) used estimates of deep ocean carbonate ion concentrations to constrain the carbon cycle. They found that the observed trend in atmospheric CO₂ during the last 8000 years can best be explained by the buildup of coral reefs and other forms of shallow water carbonate deposition. Joos et al. (2004), employing the Bern carbon cycle climate model to simulate the interval from the last glacial maximum to the preindustrial, found that a combination of processes contributed to the Holocene rise in CO₂, with carbonate compensation in response to terrestrial vegetation regrowth, SST changes and coral reef buildup playing a role. On the other hand, Brovkin et al. (2002), as well as Menviel and Joos (2012), found almost no effect of SST changes on CO₂ during the Holocene.
Kleinen et al. (2010), using the CLIMBER2-LPJ model, showed that the trend in atmospheric CO$_2$ over the Holocene is controlled by the balance of two slow processes: Carbon uptake by boreal peatlands, which is (slightly over-)compensated by outgassing of CO$_2$ due to sedimentation and the accumulation of CaCO$_3$ in shallow oceanic areas. Finally, Menviel and Joos (2012) investigated the Holocene CO$_2$ rise by applying the Bern3D ocean carbon cycle model, prescribing scenarios of shallow-water carbonate sedimentation and land C uptake. In their experiments, shallow-water carbonate sedimentation, carbonate compensation of land uptake, land carbon uptake and release, and the consecutive carbonate compensation response, as well as the response of the ocean-sediment system to marine changes during the termination contribute roughly equally to the CO$_2$ rise.

For earlier interglacials, investigations are rare. Schurgers et al. (2006) investigated the changes in atmospheric CO$_2$ during both the Holocene and the Eemian using the ECHAM3-LSG General Circulation Model (GCM), including the dynamic global vegetation model (DGVM) LPJ and the marine biogeochemistry model HAMOCC3. They found increases in atmospheric CO$_2$ for both the Eemian and the Holocene, mainly driven by decreases in the terrestrial C storage, but they do not. They were, however, unable to explain the overall magnitude of the CO$_2$ trend during the Holocene, and their positive trend in atmospheric CO$_2$ during the Eemian is distinct different from that in the ice core data, which shows actually show no trend.

Here, in the present publication, we address two major shortcomings of advance on the study by Kleinen et al. (2010): (1) both the accumulation of peatland carbon and the burial of CaCO$_3$ were prescribed scenarios and not modelled interactively, and (2) the study only considered the Holocene, while neglecting to show that the same mechanisms can also explain the evolution of CO$_2$ during previous interglacials. Our current model now includes a dynamic peatland model, as well as a dynamic model of carbonate accumulation by coral reef growth, which finally enables us to consistently investigate the evolution of atmospheric CO$_2$ during the Holocene and during two interglacials that preceded it: the Eemian and MIS 11. The Holocene and the Eemian are particularly interesting because validation data of reasonable time resolution and reliability are available for these interglacials. We did not investigate interglacials prior to the Holocene. In this paper, we therefore aim MIS 5 since no $\delta^{13}CO_2$ data are available for validation, with the exception of MIS 11 since its unusual length
makes it a particularly interesting case and ice core CO₂ data of reasonable time resolution are still available. We investigate to show how what extent the evolution of CO₂ in these three recent different interglacials, the Holocene, the Eemian, and MIS 11, can be explained by the interplay of two slow carbon cycle processes, peat accumulation and CaCO₃ accumulation in shallow waters.

2 Model and experiments

2.1 The model

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

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

In addition CLIMBER2 also contains an oceanic biogeochemistry model (Ganopolski et al., 1998, Brovkin et al., 2002, 2007) and a sediment model that describes the diffusive pore-water dynamics, assuming oxic-only respiration and 4.5-order CaCO₃ dissolution kinetics (Arch, 1996; Brovkin et al., 2007). Volcanic emissions of CO₂ are assumed to be constant at 0.07 GtC a⁻¹ (Gerlach, 2011). Weathering fluxes scale to runoff from the land surface grid cells, with separate carbonate and silicate lithological classes. The long-term carbon cycle that includes the processes of deep-sea and shallow-water carbonate accumulation, weathering and volcanic outgassing, is brought to equilibrium for the pre-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° x 0.5° 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:

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

Under present conditions in seawater about 0.6 mol of CO$_2$ will be released for every mol of CaCO$_3$ produced (Frankignoule et al., 1994). This is implicitly handled by the carbonate speciation and air-sea gas exchange routines.

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

In the original Kleypas (1997) model, sea level is only used to calculate the area available for shallow water sedimentation, but the rate of sea level change is not considered in calculating the rate of CaCO$_3$ sedimentation. However, the rate of CaCO$_3$ accumulation by coral reefs will be strongly perturbed during periods of sea level drop or very fast sea level rise. A moderate rate of sea level rise, on the other hand, can maximise coral reef buildup. We therefore implemented a dependence of the CaCO$_3$ sedimentation rate on the rate of sea level change based on Munhoven and François (1996). Munhoven and François (1996) consider a trapezoidal growth-limiting function $\Theta$ as shown in Fig. 1, which restricts the coral reef growth in case sea-level rises too fast or falls. According to Buddemeier and Smith (1988) the best overall estimate for the sustained maximum rate of reef growth is 10 mm a$^{-1}$. For simplicity we therefore adopt 0 and 10 mm a$^{-1}$ as the limiting sea-level rates. To avoid too abrupt changes, accumulation rates are reduced from 100 to 0% of the normal rate 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 2.5 mm a$^{-1}$ decrease) to +2.5 mm a$^{-1}$. We thus allow for a small accumulation even whereas long as sea-level falls only slowly. Carbonate accumulation rates will not drop to zero immediately since corals may live even at depths of 50 m and more, and their habitat therefore does not vanish immediately.

The total CaCO$_3$ production accumulation in each grid cell where ocean temperature is within the acceptable range therefore is $P = G \times \Theta \times A \times TF$, which we sum up for all grid cells to determine the total shallow-water CaCO$_3$ production accumulation. Total production accumulation is scaled to conform to the Milliman (1993) estimate of shallow water CaCO$_3$ sedimentation accumulation for the late Holocene. Milliman (1993)
estimates a sedimentation rate in shallow-water carbonate accumulation rate of about 1.5 $10^4$ billion tons, Milliman’s units, $10^4$ tonnes of CaCO$_3$ per year, which converts to 15 Tmol a$^{-1}$ using the CaCO$_3$ molar weight of 100 g mol$^{-1}$. The area factors A and TF are more or less constant over the sea level range of our experiments. Therefore variations in CaCO$_3$ formation are primarily due to changes in the rate of sea level change. In experiments where the dynamic calculation of CaCO$_3$ sedimentation is disabled, a small constant shallow water CaCO$_3$ sedimentation flux of 2 Tmol a$^{-1}$ is prescribed to balance the oceanic alkalinity budget.

### 2.3 Carbon accumulation in peatlands

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

In order to account for this accumulation of carbon, we have extended the CLIMBER2-LPJ model by developing the dynamic model of peatland extent and peat carbon accumulation, as described in Kleinen et al. (2012). This model determines peatland extent from topography and climatic conditions. Within the obtained peatland areas, it considers accumulates carbon due to the slow decomposition of C under the anaerobic conditions in the soil to accumulate carbon in the modelled peatlands. For the last 8 ka, this model calculates an accumulation of 330 PgC in high northern latitude areas, which is roughly in line with the Yu et al. (2010) estimate of 550 PgC for the time period from the LGM to the present (Kleinen et al., 2012). The main factor influencing the uncertainty of this model result is the peatland area estimate. Kleinen et al. (2012) considered minimum and maximum area estimates from which they derived an uncertainty range of 240-490 PgC for the peat accumulation between 8 and 0 ka BP, with a most likely value of 330 PgC. Tropical peatlands could, unfortunately, not be considered in the present experiments, due to the lack of reliable calibration data for tropical peatlands. Preliminary model experiments indicate that the Holocene shows a constant carbon stock in tropical peatlands has not varied over the Holocene, though we therefore assume that we introduce no major errors by neglecting them. Furthermore, they represent less than 10% of the total, according to
the figures from [Yu et al. (2010)]. Experiments in this publication It should be noted that experiments where peat accumulation is considered, display a decreased total carbon stock for soil carbon in mineral soils in comparison to the experiments where peat accumulation is not considered. In the former type of experiments the area covered by mineral soils is smaller since part of the grid cell may be set aside for peatlands. The offset in total carbon stocks between the experiments with and without consideration of peat carbon accumulation therefore does not reflect a different carbon density in any particular location, but rather the reduced area of mineral soils.

2.4 Forcing data

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

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

The Kaplan et al. (2011) scenario on CO₂ emissions from land use changes assumes cumulative emissions of ~409 PgC by 1950 (0 a BP), which we found to lead to excessively high CO₂ concentrations for the present, when combined with historical fossil fuel CO₂ emissions. We therefore scaled their emission scenario by a constant factor of 0.75 to reduce the total cumulative release to 307 PgC by 1950, keeping the timing of their CO₂ emissions. After 1765 (or 185 a BP) we add historical emissions.
from fossil fuel use from the RCP database (Meinshausen et al., 2011). The adopted
cumulative emissions are shown in Fig. 21. For simplicity, CO$_2$ emissions from land use
changes are directly added to the atmospheric CO$_2$, i.e., we do not change the land carbon
stocks when emitting CO$_2$ from land use changes. This simplification will lead to a slight
overestimate of the carbon uptake by vegetation through CO$_2$ fertilisation, though we
nevertheless judge the impact of this simplification to be minor. Both land use and fossil
fuel emissions are assumed to have a $\delta^{13}$CO$_2$ of -25‰.

2.5 Ice core data

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

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

For $\delta^{13}$CO$_2$, we compare the Holocene results to the compilation by Schmitt et al. (2012),
based on data obtained from EDC by Elsig et al. (2009) and Lourantou et al. (2010), and we
compare the Eemian results to data from Schneider et al. (2013). For both time frames, we
compare to both the original data and to the Monte Carlo average, which should remove most
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 him 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 CO$_2$ 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 CO$_2$ as diagnosed from ice cores for that time. Atmospheric $\delta^{13}$CO$_2$ 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 $^{-1}$ in the deep ocean and 2 Tmol $^{-1}$ on the shelves in order to simulate the maximum in CaCO$_3$ preservation in the deep sea before the onset of the interglacial. The model was then run with prescribed CO$_2$ 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 CO$_2$ 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 five experiments to investigate the role of the
various forcings in the interglacial carbon cycle: (1) a model HOL ORB, an experiment containing where neither peat accumulation, nor CaCO$_3$ sedimentation, nor anthropogenic land use emissions. This experiment is purely driven by but only orbital forcing. We denote it HOL ORB. is considered; (2) an HOL PEAT, a
experiment containing where peat accumulation, but neither CaCO$_3$ sedimentation nor anthropogenic land use emissions,
denoted HOL PEAT, are considered; (3) an HOL NAT, an experiment using where all of the natural forcing mechanisms, i.e., peat accumulation and CaCO$_3$ sedimentation, denoted HOL NAT, are considered; (4) The same setup HOL MPT, an experiment where all the natural forcing mechanisms are considered (as in HOL NAT), but where the minimum peatland area estimate from Kleinen et al. (2012) is used; (5) HOL ANT, which uses again the same forcings as HOL NAT, but also including anthropogenic carbon emissions, denoted HOL ALL. Experiments of in HOL ANT, which uses again the same forcings as HOL NAT, but also including anthropogenic carbon emissions, denoted HOL ALL. Experiments for each of the Eemian and MIS$_{11}$ follow the setup HOL NAT with appropriate initial conditions, assuming that anthropogenic land use did not play a role then. In addition, we performed an experiment for each interglacial, where we disabled the slow forcing factors as in set up two experiments analogous to HOL ORB, and HOL NAT, with adapted initial conditions. The characteristics of all experiments are summarised in Table 1.

All experiments are driven by orbital changes (Berger, 1978). The In the experiments that consider variable shallow-water CaCO$_3$ accumulation rates (HOL NAT, HOL ALL MPT, HOL ANT, EEM NAT, and MIS$_{11}$ NAT) also require sea level changes, as described in Sect. 2.4, and in experiment HOL ALL ANT anthropogenic CO$_2$ emissions from land use changes and fossil fuel burning are provided as an additional forcing prescribed, as described in Sect. 2.4.

3 Results

3.1 Holocene

The model experiment HOL ORB, atmospheric carbon dioxide concentrations resulting from the Holocene experiments are shown in Fig. 2a, as well as ice core data for comparison. In model experiment HOL ORB (without peat accumulation and CaCO$_3$ sedimentation in shallow waters, would correspond to the carbon cycle implemented in most earth system models (ESM), i.e., a carbon cycle not taking into account slow processes of the C cycle. As
shown in Fig. 3a, this model setup leads to a small decrease in CO₂ (~5 ppm) atmospheric CO₂ decreases by ~5 ppm over the first 2000 years, followed by constant CO₂ for the remainder of the experiment. The modelled terrestrial biomass carbon decreases by about 30 PgC during this time, as shown in Fig. 4a, while the soil carbon increases by a similar amount. Overall the conventional carbon cycle setup HOL_ORB would only lead to minor changes in atmospheric CO₂, especially missing the increase in atmospheric CO₂ by 20 ppm shown in the ice core record for 6 ka BP to 0 ka.

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

The results from our experiment HOL_NAT, including (which includes carbon storage in boreal peatlands and shallow water CaCO₃ accumulation, are shown as a magenta line in Fig. 3a. Here, the trajectory of atmospheric CO₂ closely follows the ice core measurements until about 3 ka BP (Fig. 2a, black line). Between 8 ka and 6 ka BP, the model overestimates CO₂ by up to 5 ppm, while it underestimates atmospheric CO₂ after 4 ka BP, with the discrepancy growing as the model gets closer to the present. Atmospheric CO₂ stays constant at 268 ppm after 4 ka BP in this experiment.

Finally, the results from HOL_ALL, i.e., a model setup similar to HOL_ in experiment HOL_MPT (as HOL_NAT, but with the minimum peatland area) atmospheric CO₂ increases slightly stronger than in HOL_NAT during the early Holocene and keeps increasing until 2.5 ka BP, after which it stays constant at 273 ppm CO₂ (Fig. 2a, cyan line). Finally, in HOL_ANT (as HOL_NAT but with anthropogenic emissions of CO₂ from land use changes and fossil fuel use considered, are shown in black in Fig. emissions). Here the atmospheric CO₂ is very similar to CO₂ in HOL_NAT until about 4 ka BP, and similar to HOL_MPT until 2.5 ka BP, after which HOL_ALL_ANT displays a continued increase in CO₂, in line with ice core CO₂ (Fig 2a, green line). The CO₂ trajectory stays relatively close to the measurements
over the entire time frame of the experiment, with a maximum deviation of about 8 ppm CO$_2$ at 1.5 ka BP, HOL_ANT.

Biomass carbon, shown in Fig. 4a, stays nearly constant at 550 PgC over the entire simulation period of experiment HOL_NAT, in contrast to the decrease observed for HOL_ORB. For the first 5 ka, biomass carbon in HOL_ALL is very similar to HOL_NAT, but after 2.5 ka BP it increases driven by the increase in atmospheric CO$_2$, and reaches more than 600 PgC at the end of the experiment. Soil carbon stocks, shown in Fig. 4b, initially are 110 PgC lower in HOL_NAT and HOL_ALL than in HOL_ORB. This difference is due to the fact that some areas, especially in the high latitudes rich in soil C, are set aside as peatlands and therefore not available for mineral soil carbon storage. In experiment HOL_NAT the soil carbon stock increases from an initial 1325 PgC to about 1400 PgC at 0 ka. The evolution in HOL_ALL is very similar for the first 5 ka, but after 3 ka BP soil carbon increases more than in HOL_NAT due to higher CO$_2$, and reaches a maximum of 1425 PgC at the end of the experiment.

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

Figure 4c shows the cumulative carbon uptake by peatlands in experiments HOL_NAT and HOL_ALL. Carbon storage in peatlands increases nearly linearly over the entire time of the experiment (in fact, carbon uptake only saturates after several tens of ka), up to a total of 330 PgC accumulated at the end of experiment HOL_ALL, while HOL_ANT carbon fluxes leading to these trajectories in CO$_2$ are shown in Fig. 3. For clarity we have smoothed the plots using a Gaussian filter (length 1000 a, stronger weighting in the center following a Gaussian distribution). Fig. 3a shows the geological and anthropogenic C flux to the atmosphere, i.e., the sum of volcanic outgassing, weathering and anthropogenic fluxes, while Fig. 3b shows the net land – atmosphere carbon flux and Fig. 3c shows the net ocean – atmosphere C flux. Since the volcanic input is constant in time, the changes shown in Fig. 3a mainly reflect changes in weathering, with the exception of experiment HOL_ANT, where the bulk of the changes is due to anthropogenic emissions. In experiment HOL_ORB, shown in blue, weathering takes up slightly more carbon in the early than in the late Holocene. This
carbon uptake by weathering is compensated by C emissions from the ocean, as shown in Fig. 3c, while the land is carbon neutral. The slight decrease in atmospheric CO$_2$ displayed during the first 2000 years of experiment HOL ORB therefore is the result of the slightly stronger weathering during the early Holocene. Although the land is carbon neutral overall, a shift in carbon allocation becomes apparent in Fig. 4, where the evolutions of the land C pools are shown. In experiment HOL ORB, vegetation carbon decreases by 30 PgC over the time of the experiment (Fig. 4a), while soil carbon increases by a similar amount (Fig. 4b).

The decrease in CO$_2$ in experiment HOL PEAT (not shown) accumulated 320 PgC. The difference (magenta) is caused by an uptake of carbon by the land surface. In Fig. 3b, a more or less constant carbon uptake flux of about -0.03 PgC a$^{-1}$ is shown, caused by the accumulation of 310 PgC of peat (Fig. 4c). Vegetation loses about 80 PgC (Fig. 4a), while soils lose about 20 PgC (Fig. 4b), due to the lower concentration of atmospheric CO$_2$, lessening the impact of the peat accumulation. Note that the overall soil C pool is decreased in comparison to HOL ORB for all experiments considering peat carbon accumulation since the area available for carbon storage in mineral soils is decreased due to the consideration of peatlands. In addition, C is released from the ocean (Fig. 3c), partially compensating the carbon uptake.

Carbon fluxes in experiments HOL NAT and HOL MPT are very similar. They also display a slight decrease in the atmospheric weathering flux (Fig. 3a), though less pronounced than in experiments HOL ORB and HOL PEAT due to the higher concentration in atmospheric CO$_2$ later in the Holocene. The uptake of carbon by the land, shown in Fig. 3b, is higher than in experiment HOL PEAT, due to higher CO$_2$ concentrations, and displays a maximum at the beginning of the experiment. Soil carbon (Fig. 4b) increases over the entire experiment, while vegetation carbon (Fig. 4a) stays constant. Peat accumulates (Fig. 4c) in both experiments, though the total accumulation is different: 340 PgC in HOL NAT, but only 250 PgC in HOL MPT. The increase in atmospheric CO$_2$ during the early Holocene is driven by the release of carbon from the ocean in these experiments, as shown in Fig. 3c. This increase in CO$_2$ release from the ocean relative to the previous experiments is caused by the release of CO$_2$ during the formation of the CaCO$_3$ that accumulates in shallow waters, especially in coral reefs, as shown in Fig. 5b. The coral CaCO$_3$ accumulation flux is 28 Tmol a$^{-1}$ during the early Holocene due to the large change in sea level (Fig. 5a). It decreases to 18 Tmol a$^{-1}$ at 4 ka BP and stays constant thereafter due to the constant sea level.
Carbon fluxes in experiment HOL_ANT cannot be shown fully in Fig. 3 since their magnitude becomes substantially larger than the natural carbon fluxes described for the previous experiments towards the late Holocene. We have therefore added inset figures showing fluxes from HOL_NAT and HOL_ANT for the last 3.5 ka of the experiments. The plots in the inset Figures have been smoothed using a shorter filter length of 50 years since the long filtering used previously hides the substantial changes induced by industrial CO\textsubscript{2} emissions after 1765 AD. The geological and anthropogenic C flux to the atmosphere peaks at 2 PgC a\textsuperscript{-1} in 0 BP, due to the large anthropogenic C emissions. During the early Holocene the anthropogenic flux is substantially smaller, though. Here, it lessens the impact of weathering and slightly increases the geological and anthropogenic C flux to the atmosphere in comparison to HOL_NAT, as shown in Fig. 3a. The land – atmosphere flux is very similar to experiment HOL_NAT until 3 ka BP, the land generally takes up carbon due to peat accumulation. After 3 ka BP, the land – atmosphere flux becomes more negative than in the other experiments, the land C uptake increases due to CO\textsubscript{2} fertilisation effect of CO\textsubscript{2} on photosynthesis and reaches a minimum of -0.6 PgC a\textsuperscript{-1} at the end of the experiment (Fig. 3b, inset Figure). This is reflected in the land C pools (Fig. 4), which keep increasing throughout experiment HOL_ANT. The ocean – atmosphere flux in experiment HOL_ANT is similar to those in experiments HOL_NAT and HOL_MPT, though slightly smaller, until 3 ka BP, when it starts deviating from the other experiments (Fig. 3c). The flux becomes negative after 75 BP (1875 AD) (Fig. 3c, inset Figure), i.e., the ocean switches from being a sink for carbon to a source.

The increase in atmospheric CO\textsubscript{2} during the early Holocene in experiments HOL_NAT, HOL_MPT, and HOL_ANT is due to the strongly positive ocean – atmosphere carbon flux, caused by the accumulation of CaCO\textsubscript{3} in shallow waters. Sea level initially rises fast (see Fig. 5a), reaching stable levels around 5 ka BP. The shallow water CaCO\textsubscript{3} accumulation rate, shown in Fig. 5b, varies with the rate of sea level change. The rate of sea level change is highest early during the Holocene, about 2 mm a\textsuperscript{-1}, leading to a CaCO\textsubscript{3} accumulation of about 27 Tmol a\textsuperscript{-1}. Sea level stabilises later in the Holocene, leading to reducing shallow-water CaCO\textsubscript{3} sedimentation of about 15 Tmol a\textsuperscript{-1} in all three experiments. The formation of CaCO\textsubscript{3} in experiments HOL_NAT, HOL_MPT, and HOL_ANT leads to a reduction in mean ocean alkalinity, as shown in Fig. 5c. Over the course of the Holocene, the mean ocean alkalinity is reduced by about 10% in the experiments where shallow water CaCO\textsubscript{3} accumulation is considered.
The evolution of atmospheric $\delta^{13}$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 $\delta^{13}$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 $\delta^{13}$CO$_2$ when coupling the interactive CO$_2$ from CLIMBER and the LPJ land carbon cycle after the spinup of LPJ. Modelled $\delta^{13}$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 $\delta^{13}$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 $\delta^{13}$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.

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 $\delta^{13}$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 $\delta^{13}$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
discrepancy in CO₂ between experiment EEM_ORB and the ice core data is not excessive, the fit of experiment EEM_NAT to the data is substantially better. The slow natural processes we consider therefore seem to be required to explain the evolution of CO₂ during the Eemian. In experiment EEM_NAT, in contrast, modelled atmospheric CO₂ changes relatively little for the entire experiment and is generally within the range spanned by the error bars of the measurements, with exceptions only around 119.5 ka BP, where 3 data points show lower CO₂ concentrations than modelled. δ¹³CO₂ for EEM_NAT, smoothed by a Gaussian filter and offset by -0.0647‰ as for the Holocene experiments, is shown in Fig. 6b. Modelled δ¹³CO₂ stays within the 1σ uncertainty of the Monte Carlo estimate from Scheider et al. (2013) for the entire time of the experiment.

The terrestrial biomass (Fig. 7a) reaches a maximum of about 600 PgC early in experiment EEM_NAT at 124 ka BP. It decreases thereafter and reaches a minimum value of ~490 PgC at the end of the experiment at 116 ka BP. Biomass carbon in experiment EEM_ORB follows a very similar trajectory. Soil carbon in EEM_NAT (Fig. 7b) increases from an initial value of 1325 PgC to about 1400 PgC at 121 ka BP and decreases thereafter, reaching 1225 PgC at 116 ka BP. The evolution in EEM_ORB is similar, though offset by about 90 PgC, again due to the larger area available for mineral soil carbon when no peatlands are considered. The carbon storage in peatlands, shown in Fig. 7c for EEM_NAT, increases linearly during the entire Eemian.

The disaggregated net carbon fluxes leading to these CO₂ trajectories are shown in Fig. 7. Weathering is very strong during the early Eemian, leading to a geological C flux of -0.24 PgC a⁻¹ for 126 ka BP, shown in Fig. 7a. Weathering then decreases, allowing the geological C flux to increase to -0.205 PgC a⁻¹ at 116 ka BP in both experiments. While the geological C flux is similar in both experiments, the other C fluxes are substantially different. For experiment EEM_ORB, the land is generally carbon-neutral during the early Eemian. The land – atmosphere C flux is close to zero until 121.5 ka BP (see Fig. 7b). Later in the Eemian, the land – atmosphere C flux increases to a maximum of 0.05 PgC a⁻¹ at 117 ka BP. For the early Eemian this is due to counteracting contributions from the land carbon pools: vegetation carbon decreases continually from 126 ka BP to 116 ka BP, with an initially slow rate of decrease that increases after 123 ka BP (see Fig. 8a). Soil carbon, in contrast, increases from 1415 PgC at 126 ka BP to 1480 PgC at 120.5 ka BP, after which it decreases to 1350 PgC at 116 ka BP (Fig. 8b). The ocean – atmosphere carbon flux, on the other hand, is initially at 0.23 PgC a⁻¹ and decreases to a minimum of 0.15 PgC a⁻¹ at 117 ka BP. Therefore the initial strong carbon uptake through weathering is not completely compensated by marine C fluxes.
leading to the modelled reduction in atmospheric CO\(_2\) between 126 and 121.5 ka BP. After this time, the land loses carbon, allowing a compensation of the (reduced) weathering flux and leading to an increase in atmospheric CO\(_2\).

In experiment EEM_NAT, on the other hand, the land – atmosphere C flux is negative until 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). While the vegetation and soil carbon pools behave in a generally similar way to experiment EEM_ORB (Fig. 8a, b), the model accumulates 445 PgC of peat carbon (Fig. 8c), resulting in the generally negative land – atmosphere flux. The ocean – atmosphere flux initially is substantially higher than in EEM_ORB, with an initial flux of 0.33 PgC a\(^{-1}\), that decreases over time to 0.17 PgC a\(^{-1}\). Eemian as well, until about 440 PgC are accumulated at the end of the experiment.

The sea level forcing, shown in Fig. 8a, is stable early during the experiment and decreases after 121 ka BP. Therefore shallow-water CaCO\(_3\) accumulation (Fig. 8b) is at a rate of \(\approx\)20 Tmol a\(^{-1}\) during the early Eemian (Fig. 9b), lower than during the early Holocene. It decreases to about zero at 119 ka and stays at this level thereafter. This increases the ocean – atmosphere flux in comparison to experiment EEM_ORB, thus releasing carbon to the atmosphere, which compensates the peat carbon uptake and the strong weathering flux during the early Eemian.

### 3.3 MIS 11

For MIS 11, the agreement between the modelled atmospheric CO\(_2\) concentrations in MIS11_NAT and the ice core reconstruction is not as good as for the other two interglacials. As shown in Fig. 9, modelled CO\(_2\) in experiment MIS11_NAT increases initially from 271 ppm CO\(_2\) to about 290 ppm at 412 ka BP. It declines thereafter to about 250 ppm CO\(_2\) at 395 ka BP, after which CO\(_2\) varies much less. Setup MIS11_ORB, on the other hand, shows a slowly decreasing trend in CO\(_2\), from the initial 271 ppm CO\(_2\) to slightly less than 260 ppm at 380 ka BP, with only little variation about this trend.

The initial increase in CO\(_2\) is slower in the ice core data than in MIS11_NAT. CO\(_2\) increases to about 285 ppm at 407 ka BP. Measured CO\(_2\) decreases strongly after 398 ka BP, until 250 ppm CO\(_2\) are reached at 390 ka BP. Therefore the model setup MIS11_NAT overestimates the initial increase in CO\(_2\), and the peak in CO\(_2\) is reached about 5 ka earlier than in the ice core data. Similarly, the decrease after the peak in CO\(_2\) also occurs earlier in the model than in
the ice core data. Nonetheless, the overall CO₂ trajectory, with an initial increase in CO₂ between 420 ka and 405 ka BP, followed by a decrease by about 25-30 ppm and a stabilisation of CO₂ after 395 ka BP is captured by MIS11_NAT, though the timing is not exactly the same as in the ice core data. MIS11_ORB, on the other hand, does not at all follow the ice core CO₂ data. For the interested reader, we show the δ¹³CO₂ for experiment MIS11_NAT in Fig. 10b, though no ice core data is available for comparison. δ¹³CO₂ starts out at -6.5‰, with a slowly decreasing trend until 405 ka BP. Afterwards it decreases quickly to -6.55‰ at 398 ka BP. It then slightly increases again until 390 ka BP, after which it decreases further until -6.57‰ are reached at 380 ka BP.

The land carbon pools display substantially more variability in MIS11_NAT than in MIS11_ORB, shown in Fig. 10a and b. Biomass carbon (Fig. 10a) increases strongly in MIS11_NAT. The geological C flux, shown in Fig. 11a, is very similar in both experiments; changes in MIS11_NAT only present a slightly larger amplitude than MIS11_ORB due to the higher CO₂ concentrations. It decreases from -0.208 PgC a⁻¹ at 420 ka BP to -0.233 PgC a⁻¹ at 410 ka BP in MIS11_NAT, reflecting increases in weathering. Subsequently it increases to the initial value at 395 ka BP, followed by a small decrease and further increase to a final value of -0.204 PgC a⁻¹. The land – atmosphere flux in experiment MIS11_ORB fluctuates around zero (Fig. 11b), with increases in vegetation carbon (Fig. 12a) compensated by decreases in soil carbon (Fig. 12b) and vice-versa. In MIS11_NAT, the net land – atmosphere flux is more negative due to the accumulation of peat carbon. It is -0.09 PgC a⁻¹ initially and increases to zero at 400 ka BP, with a subsequent decrease to -0.035 PgC a⁻¹ at 395 ka BP and an increase to -0.01 PgC a⁻¹ at 380 ka BP. This is due to the changes in the land carbon pools: vegetation carbon (Fig. 11a) increases strongly, until a maximum value of about 630 PgC is reached at 412 ka BP. Carbon storage decreases afterwards, until a minimum of 480 PgC is reached at 395 ka BP, with only small changes afterwards. Similarly, soil carbon increases early in MIS11_NAT from an initial value of 1350 to about 1425 PgC at 414 ka BP. It then stays constant until 403 ka BP, when it starts decreasing strongly. After 395 ka BP soil carbon stays constant at 1345 PgC. In contrast, the variations in biomass and soil carbon are much less pronounced in experiment MIS11_ORB. Biomass carbon increases from 540 to 560 PgC early in MIS 11, then decreases again to 515 PgC at 395 ka BP, and changes little afterwards. Soil carbon, on the other hand, varies between 1490 and 1445 PgC during the entire time frame of the experiment. Peat accumulation in MIS11_NAT (Fig. 10c) once
again 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 CO$_2$ 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 CaCO$_3$ accumulation in shallow waters. During the first 13 ka of MIS 11 sea level increases from -20 m to near zero (Fig. 11a). It starts decreasing again at 407 ka BP, but stabilises at -15 m after 395 ka BP. This sea level trajectory is reflected in the CaCO$_3$ accumulation flux, shown in Fig. 11b: the initial fast rise in sea level leads to an accumulation rate of up to 29 Tmol a$^{-1}$, with a correspondingly high CO$_2$ 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 CO$_2$ data. The model setup HOL_ORB, i.e., a carbon cycle setup without anthropogenic CO$_2$ emissions or slow natural processes, leads to a more or less constant CO$_2$ 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 HOLPEAT 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 CO$_2$ emissions from CaCO$_3$ shallow water sedimentation in HOL_NAT then leads to an increase in atmospheric CO$_2$, not just compensating the C uptake by peatlands, but also releasing additional CO$_2$ 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
hypothesis. However, it becomes clear from the difference between experiments HOL_NAT and HOL_ANT that anthropogenic \( \text{CO}_2 \) emissions from land use changes only make a significant difference to atmospheric \( \text{CO}_2 \) after about 3 ka BP, which would shift to an even later time with smaller peat carbon accumulation (as in HOL_MPT). Anthropogenic emissions therefore cannot explain the 10 ppm rise in \( \text{CO}_2 \) observed in ice cores between 8 and 4 ka BP. For the earlier Holocene, \( \text{CO}_2 \) emissions from shallow water CaCO\(_3\) sedimentation are required instead. The continued rise in \( \text{CO}_2 \) after 2.5 ka BP, on the other hand, can only be explained if anthropogenic emissions are accounted for as well.

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

Modelled vegetation changes in our model compare well against tree cover reconstructions from Eurasia (Kleinen et al., 2011). They are also similar to vegetation changes obtained with other models, for example CLIMBA and Bern3D, as published by Brovkin et al. (2016). However, it is not possible to validate the modelled changes in terrestrial carbon storage since no direct proxy exists for carbon stored in terrestrial ecosystems. The \( \text{CO}_2 \) fertilisation effect displayed by CLIMBER2-LPJ as well as other DGVMs, which leads to increases in biomass with increasing \( \text{CO}_2 \), seems well-understood at the leaf level (De Kauwe et al., 2014), but...
may be overestimated in models because constraining mechanisms such as nutrient limitation are not taken into account (Reich et al., 2014).

Our CaCO$_3$ accumulation model seems to capture the late Holocene sedimentation, within good agreement with Milliman (1993). Nonetheless the increase in accumulation rate due to the rate of sea level rise during the earlier Holocene is relatively uncertain. This is due to uncertainties in the parameterisation, as well as uncertainties in the rate of sea level rise. While both are plausible, there is considerable uncertainty with respect to magnitude and timing of the CO$_2$ emissions from CaCO$_3$ formation. Previous assessments nevertheless agree, though, that coral growth was stronger in the early Holocene (Ryan et al., 2001; Vecsei and Berger, 2004). The change in sea level we use as a model forcing agrees well with sea level reconstructions for the Holocene, though sea level stabilises up to 2 ka earlier in our model than it does in reconstructions (Masson-Delmotte et al., 2013, Fig. 5.17 f). A later stabilisation of the sea level, as in palaeoclimatic archives, would lead to a prolongation of the relatively larger emissions from CaCO$_3$ formation we model for the early Holocene. Therefore we may underestimate the emissions from shallow water CaCO$_3$ accumulation.

Finally, the modelled trajectory of $\delta^{13}$CO$_2$ for the Holocene has relatively high values between 4 ka BP and stays well within the present, as shown in Fig. 4b. These values are outside the range of the error bars estimated by Elsig et al. (2009). This result can be explained in three different ways: (a) Elsig et al. might have underestimated the true $2\sigma$ uncertainty, (b) we may have underestimated the $\delta^{13}$CO$_2$ band of the Monte Carlo based uncertainty assessment by Schmitt et al. (2012), if one considers experiment HOL_ANT, while the other experiments leave this range. While the general dynamics of $^{13}$C seem to be captured well by the LPJ model (Scholze et al., 2003), there is some uncertainty with regard to the $^{13}$C changes induced by the accumulation of peat, and (c) we may require an unknown additional source of isotopically depleted carbon to explain the trajectory of $\delta^{13}$CO$_2$. This latter explanation has been favoured by proponents of large anthropogenic emissions from land use changes, since CO$_2$ released from the biosphere would have such a depleted isotopic signature (Ruddiman et al., 2011). At 262 PgC cumulative emissions from land use changes, the scenario adopted here already assumes larger fluxes than other recent estimates. Stocker et al. (2014), for example, estimate the cumulative emissions by 2004 at 243 PgC. Besides, judging from Fig. 4b, the modelled atmospheric $\delta^{13}$CO$_2$ is higher than the measurements after about 4.5 ka BP, earlier than the bulk of the emissions in the scenario based on Kaplan et al.
Emissions from anthropogenic land use changes therefore do not appear to be a likely cause of the mismatch in $\delta^{13}$C, but we cannot rule out other isotopically depleted sources of C, such as methane emissions or the release of carbon from thawing permafrost soils. With regard to (b), we assume that the carbon uptake by peat accumulation has a similar signature in $\delta^{13}$C as the growth of C3 grass. Since photosynthesis in mosses generally follows the C3 pathway, this assumption appears reasonable, and values for $\delta^{13}$C in mosses reported in the literature (e.g. Waite and Sack, 2011) are in a similar range as values for other C3 vegetation. With regard to (a), finally, there are no reasons to believe (e.g. Waite and Sack, 2011). However, the cycling of $^{13}$C in peatlands seems to be less well understood than in other terrestrial systems, which makes the modelled $\delta^{13}$C changes induced by peat accumulation less certain. In addition we cannot rule out other isotopically depleted sources of C, such as methane emissions or the release of carbon from thawing permafrost soils, that measurement errors are underestimated by Elsig et al. (2009), forcing us to reject (a) as well. This leaves unknown sources of isotopically depleted C as the most likely explanation for the discrepancy in $\delta^{13}$C we have not accounted for in our model.

With regard to the evolution of atmospheric CO$_2$ during the Eemian, the fit between ice core data and model results is clearly better for experiment EEM_NAT than for EEM_ORB. While the model produces an initial decrease followed by an increase for EEM_ORB, EEM_NAT shows a nearly constant CO$_2$ concentration for the entire period of time we modelled, very close to the measurements by Schneider from Bereiter et al. (2013, 2015). Similarly, modelled $\delta^{13}$CO$_2$ is within the error bars 1σ uncertainty range of the ice-core measurements Schneider et al. (2013) Monte Carlo average for most of the time, though the model displays less change in $\delta^{13}$CO$_2$ than the MC average. Here the largest uncertainty in our setup again stems from the sea level history, leading to uncertainty with respect to magnitude and timing of CO$_2$ emissions that result from CaCO$_3$ sedimentation. However, the sea level forcing we use is similar to reconstructed global mean sea level as shown by Masson-Delmotte et al. (2013, Fig. 5.15 a and b) and should therefore be a reasonable approximation.

In our setup, and with the sea level forcing data we use, the CO$_2$ emissions from CaCO$_3$ sedimentation counterbalance the weathering-induced decrease in CO$_2$ shown in setup EEM_ORB for the early Eemian, while carbon uptake by peatlands compensates for the increase in CO$_2$ modelled in EEM_ORB during the second half of the Eemian.
For MIS 11 our model experiment MIS11_NAT displays a qualitatively similar evolution of atmospheric \( \text{CO}_2 \) as the ice core data, with an initial increase, followed by a decrease during the middle of the interglacial until the \( \text{CO}_2 \) concentration stabilises for the later part of the interglacial. This leads to a clearly better fit to the ice core measurements than setup MIS11_ORB, which shows a continuous slow decrease in atmospheric \( \text{CO}_2 \). Nonetheless there still are discrepancies in the timing and the magnitude of the changes in \( \text{CO}_2 \) between model and ice core data. This discrepancy is most likely again due to uncertainty in the sea level history that we use to force the model. If the increase in sea level before 410 ka BP were slightly less pronounced and the decrease in sea level after 405 ka BP slightly delayed, our model results would fit the ice core data even better.

Carbon uptake by peatlands does not change substantially, neither during any of the interglacials, nor between interglacials. In all cases we obtain a more or less linear rise in peatland carbon storage. Sea level reconstructions for MIS11 show considerable discrepancies between each other, making an evaluation of the quality of our forcing data very difficult. Between the four reconstructions of MIS11 sea level data that we considered (Waelbroeck et al., 2002; Rohling et al., 2010; Elderfield et al., 2012; Grant et al., 2014), there is a general qualitative agreement in timing: Sea level rises between 420 and 405 ka BP, followed by a decrease. While the sea level history from Waelbroeck et al. (2002) shows a sustained decrease, this decrease ends at about 390 ka BP in the other three reconstructions, followed by either a plateau or a slight increase. The four reconstructions furthermore disagree with regard to the magnitude of changes, with the sea level highstand at ~405 ka BP ranging from -10m to +30m relative to the present. Our forcing trajectory falls well within the range of these reconstructions. However, the magnitude and timing of changes in sea level has a large impact on the \( \text{CO}_2 \) emissions from shallow water \( \text{CaCO}_3 \) accumulation, making the latter relatively uncertain. If, for example, the increase in sea level before 410 ka BP were slightly less pronounced in our forcing data and the decrease in sea level after 405 ka BP slightly delayed, our model results would better fit the ice core data.

Our study has several other limitations. We imposed anthropogenic emissions from land use changes as a simple flux to the atmosphere without changing the land carbon stocks. This simplification modifies the uptake of carbon by the biosphere and should already be contained in the Kaplan et al. (2011) \( \text{CO}_2 \) emission estimate, but an inconsistency remains nonetheless. We also neglected the long-term memory of the carbonate compensation response to the
release of carbon from the deep ocean and the early interglacial carbon uptake by the terrestrial biosphere during deglaciation. While CLIMBER2-LPJ contains all relevant processes, we did not model this period transiently and our results therefore do not have contain the long-term memory signal in our results. Menviel and Joos (2012) found that these memory effects could be of the order of a few ppm for the Holocene. Furthermore, we assumed that the long-term carbon cycle was in equilibrium in the pre-industrial climate, but this assumption is a simplification as the balance between carbonate burial, weathering, and volcanic outgassing could be out of equilibrium for other climates. As follows from control simulations without forcings (not shown), these effects can be of the order of few ppm as well. Last but not least, several other mechanisms that are currently under discussion such as changes in permafrost carbon pools (Schneider von Deimling et al., 2012) or methane hydrate storages (Archer et al., 2009) are not accounted for, as modelling of these processes is still in an early stage and because of the lack of reliable constraints on the amplitude of interglacial changes in these potentially large carbon pools.

5 Conclusions

We show -- to our best knowledge for the first time -- how the trends in interglacial atmospheric CO$_2$, as reconstructed from ice cores, can be reproduced relatively well by a climate model with identical forcing parameterisation for three recent interglacials. For these trends in atmospheric CO$_2$ it is important to account not just for cannot be reproduced well if only the marine and terrestrial carbon cycle components, as implemented in most earth system models (Ciais et al., 2013), are considered. Instead, it is necessary to also consider the modelled CO$_2$ change is considerably improved if two slow processes of CO$_2$ change currently neglected in the most comprehensive carbon cycle models, namely the carbon accumulation in peatlands and the CO$_2$ release from CaCO$_3$ formation and burial accumulation in shallow waters. This, are accounted for as well. The latter process leads to an increase in atmospheric CO$_2$ during periods of constant or slowly rising sea level, while the former process leads to a decrease in atmospheric CO$_2$.

For the Holocene, we can explain the rise in atmospheric CO$_2$ between and 3 ka BP purely by natural forcings, while later in the Holocene, starting at about 3 ka BP, anthropogenic emissions from land use changes and fossil fuel use play an important role. The increase in atmospheric CO$_2$ during the early Holocene therefore is the result of enhanced shallow-water sedimentation accumulation of CaCO$_3$ due to rising sea level. For the Eemian, our carbon
cycle model also leads to a satisfactory simulation of yields an atmospheric CO$_2$, which history that is very close to agreement with the ice core data. Here the consideration of the slow carbon cycle processes also leads to an improvement over the conventional model approach that neglects these. For MIS 11, finally, the conventional model setup does not simulate the changes in CO$_2$ observed throughout MIS 11, while the model with consideration of the slow forcings can explain the magnitude of changes in atmospheric CO$_2$, though the timing of changes is slightly different from the ice core data. This discrepancy is possibly due to the sea level forcing history that we use to drive the shallow water CaCO$_3$ accumulation in our model, and which remains uncertain.

Despite the uncertainties discussed above, we can draw some robust conclusions with regard to the timing of CO$_2$ changes. Early during interglacials, when sea level still rises, shallow water accumulation of CaCO$_3$ and the related CO$_2$ release is larger than in periods of stagnating or receding sea level. The carbon uptake by peatlands, on the other hand, is a more or less constant forcing factor. This uptake balances the CO$_2$ emission from CaCO$_3$ precipitation during periods of constant sea level. A rising sea level therefore leads to atmospheric CO$_2$ increases, while a decline in sea level strongly reduces shallow-water CaCO$_3$ sedimentation and accumulation, leading to a reduction in atmospheric CO$_2$.

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

<table>
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<tr>
<th>Name</th>
<th>Interglacial</th>
<th>Initial CO₂ [ppm]</th>
<th>Initial δ¹³CO₂ [%]</th>
<th>Initial time [ka BP]</th>
<th>Peat accumulation</th>
<th>Coral CaCO₃ sedimentation</th>
<th>Anthropogenic land use emissions</th>
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<td>No</td>
<td>No</td>
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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 3: Holocene CO$_2$ concentration (a) and δ$^{13}$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_OBS (blue), model without coral and anthropogenic forcing HOLPEAT (green), experiments and ice core data. δ$^{13}$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). δ$^{13}$CO$_2$ data is from EDC (Elsig et al., 2009), with Monte Carlo average (MC) and uncertainty estimate (MC σ) (Schmitt et al., 2012).

Figure 4: 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 5: Holocene experiment HOL_ALL: sea level forcing (a) and shallow water CaCO$_3$ formation accumulation flux (b). (b) Also contains background CaCO$_3$ formation from HOL_OBS (blue), and mean ocean alkalinity (c) in the Holocene experiments. Plots have been smoothed for clarity.

Figure 6: Eemian CO$_2$ concentration (a) for experiments EEM_NAT (black) and EEM_OBS (blue) and δ$^{13}$CO$_2$ for EEM_NAT (b). Red error bars are CO$_2$ and δ$^{13}$CO$_2$ from EPICA Dome C.

Figure 7: Land carbon pools in Eemian experiment EEM_NAT (black) and EEM_OBS (blue): total biomass vegetation 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_OBS (blue line). Plots are smoothed for clarity.

Figure 9: MIS11 CO$_2$ concentration for experiments MIS11_NAT (black) and MIS11_OBS (blue), as well as CO$_2$ reconstruction from ice core (red) model experiments and ice core data. δ$^{13}$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). δ$^{13}$CO$_2$ data are from EDC (Schneider et al., 2013), with Monte Carlo average (MC) and uncertainty estimate (MC σ).

Figure 10: 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.
Figure 8: Land carbon pools in MIS11 experiment MIS11_NAT (black) and MIS11_ORB (blue); the Eemian experiments: total biomass vegetation carbon (a), total non-peat soil carbon (b), and cumulative C uptake by peatlands (c).

Figure 9: Sea level forcing (a), shallow water CaCO$_3$ accumulation flux (b), and mean ocean alkalinity (c) in the Eemian experiments. Plots have been smoothed for clarity.

Figure 10: MIS 11 CO$_2$ concentration (a) from model experiments and ice core data, and $\delta^{13}$ of CO$_2$ from model experiment MIS11_NAT. $\delta^{13}$ CO$_2$ model results have been smoothed for clarity. Ice core CO$_2$ data are from EDC and Vostok as compiled by Bereiter et al. (2015). $\delta^{18}$CO$_2$ from ice cores is not available.

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

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

Figure 13: Sea level forcing (a) and shallow water CaCO$_3$ formation (b). Also contains background CaCO$_3$ formation from MIS11_ORB (blue line). Plots are, and mean ocean alkalinity (c) in the MIS 11 experiments. Plots have been smoothed for clarity.