Response to the comments by the reviewers.

We thank both reviewers for their effort and commitment to review our manuscript in great detail. This is very much appreciated. We considered all comments and modified the manuscript in response. In particular, we extended the introduction and discussion to refer the reader to other simple climate models in the literature and to better explain the main features of BernSCM. Additional figures have been added as requested. The text flow in section 3 and 4 has been reorganized. The main conclusions and results remain unchanged compared to the previously submitted version.
For convenience, we repeat the comments by the reviewers below. The answers are given in indented text.
A revised version with the changes highlighted is attached to this response.

H. Metzler (Referee)

General comments
This very interesting paper describes the Bern Simple Climate Model (BernSCM) v1.0. BernSCM simulates relations between CO2 emissions, atmospheric CO2, radiative forcing (RF), global mean surface air temperature (SAT), as well as carbon and heat fluxes between atmosphere, ocean, and land biosphere. It is a reduced form coupled carbon-climate model that emulates more complex coupled models by replacing complicated components with nearly linear behavior by impulse response functions (IRFs). This (to the best of my knowledge, novel) approach leads to a coupled carbon climate model which is easy to understand and needs only low computational cost to be run. Comparisons with results from two multi-model intercomparison studies (C4MIP; IRFMIP from Joos et al., 2013) show that BernSCM simulations give representative results with respect to current knowledge about carbon-climate interactions. I am convinced that this manuscript can be scientifically important in two ways: 1) The practical application of the model itself or extended versions in its own right or as part of bigger models can lead to advances in multiple directions. 2) The theoretical foundations of the manuscript based on IRFs provide an interesting perspective on the theory of ecological modelling. Very appealing is the interpretation of the IRFs as representing parallel systems with multiple boxes, for example. Apart from some minor exceptions, the manuscript and in particular the appendix and the provided Fortran code of the model are carefully prepared. The authors took care that the BernSCM model and its implementation can be reproduced. Furthermore, the manuscript is well organized and the results are nicely presented.

> Thank you for these nice words

There are some technical problems with the equations that describe the model and I suspect an inherent theoretical problem as soon as the IRFs become time-dependent or depend on other states of the system (e.g., temperature, CO2). While the technical problems can be solved easily, I am not sure about the theoretical issue, as I will explain in more detail below.

Even if the theoretical issues cannot be completely resolved, I consider this manuscript worth for publication in Geoscientific Model Development, if the authors make the readers aware of the situation.

> Please see the answer to the specific comments below.

Specific comments

General explanation
The theoretical idea of this manuscript is very appealing. The IRFs used to substitute complex model components are provided by earlier simulations of highly complex models and just plugged into BernSCM. This makes the model structure pretty simple and the model can be used to understand ongoing processes on a global level without getting lost in distracting details, for example. Furthermore, the computation is very fast due to the use of the IRFs. This speed is even improved by disassembling the IRFs into their most important time scales, which allows an interpretation of the substituted IRFs as describing an underlying parallel multi-box model. This approach allows a very fast recursive computation which is carefully explained in great detail in the appendix and implemented in the provided Fortran package of BernSCM v1.0.

Thank you for these remarks.

The unit issue

In some equations the units do not fit. The main reason is that ε has been given the wrong explanation and the wrong unit in Table 2. The correct unit is GtC/ppm and a better description could be "mass of C per atmospheric concentration". This solves the unit problem in equations (5) and (25).

Thank you for spotting this error. We changed the text on table 2 to read:
"Atmospheric mass of C per mixing ratio 2.123 GtC/ppm "

Equation (8) should then be $p_{CO_2A} = m_A \cdot \varepsilon^{-1}$ (8)

As far as I could see from the code, it is implemented correctly.

Equation corrected as proposed.

In equation (7) the units do not give the desired (Table 1) μ mol/kg. To that end the unit of $M_{\mu mol}$ needs to be changed to gC/μ mol.

Table 2 entry corrected to read:
"mass of DIC per micromole 12.0107 10^{-6} gC/μ mol"

Very confusing is also the use of different time units like in equation (9). Carbon fluxes are measured per year and heat fluxes per second (W=J/s). Nevertheless, the integral limits are in both cases t_0 and t. I could not find any correction term in the manuscript. In the code, this correction seems to be made.

We prefer to continue to use units of "year" for carbon fluxes and units of "Watt" for heat fluxes. These units are commonly used in the literature. For example, carbon emissions are typically tabulated as annual emissions in GtC/yr, while GHG radiative forcing are given in W/m2.
The following explanation is added after eq. 16:
"Note that for compatibility with commonly used units, carbon fluxes are expressed in Gt per *year*, while heat fluxes are expressed in Joule per *second* (Watt) in equations (15) and (16), respectively."

In equation (10), the unit results in W, not in PetaW as stated in Table 1. Also here a correction term is necessary. Again, this seems not to be an issue in the code.

The unit is now indicated as W. Note that the equation does not depend on the unit used for heat fluxes.

Linear equations and IRFs

Equation (14) is only true if $m(t_0) = 0$. The general equation for the state m at time t is

$$m(t) = r(t - t_0)\, m(t_0) + \int_{t0}^{t} f(t')\, r(t - t')\, dt' \qquad (2)$$

Since the authors use equation (14) to compute perturbations with an equilibrium value $m(t_0) = 0$, this does not lead to problems, but the way equation (14) is described is mathematically not correct. I have the feeling this happened, because the authors from the beginning had a perturbation with equilibrium equal to zero in mind, but started the section then with a slightly more general set up. Line 17 on page 5 does not mention perturbations.

We have changed the notation by extending the lower integration limit to negative infinity. This avoids the issues mentioned by the reviewer, and can be applied to perturbations or totals (the latter is necessary to capture the response of terrestrial carbon stocks to warming).

The infinite time scale issue

When equation (20) is inserted in (14) to obtain equation (21), α_∞ somehow disappears.
As soon as $\alpha_\infty \neq 0$ (ocean IRF), a term is missing in equation (21). The equation should then look like

$$m(t)=\int_{t0}^{t} f(t')\alpha_{\infty}dt' + \sum_k \int_{t0}^{t} f(t')\alpha_k \exp(-(t-t')/\tau_k)\ \ dt' \qquad (3)$$

If $f(t') > c > 0$, then the first part of the equation goes to infinity as $t \to \infty$ and the perturbation grows indefinitely. If some constant share from a constant input is never going to be decayed, this share accumulates forever. Also the carbon coming from carbon conversion in the ocean model (page 7, line 20), is going to decay at some point. A constant share of remaining carbon should not result from a multiplication with an input flux coming from the atmosphere. The same explosion effect can be seen in equation (A1). Also in Equation (A14) the to α_∞ associated term 1/2 α_∞ $(\Delta t)^2$ is missing. It cannot already be included in the present sum, because $B_\infty = 1/2\ \alpha_\infty$ is never going to be multiplied with an exponential. I do not know how this problem is handled in the implementation of the model.

- The term for the infinite time scale is now explicitly added in eq. (21) to (24) and in section A1
- The term for the infinite time scale is correctly treated in the code.
- We added the following text on p6, line 22: "We emphasize that the implementation considering only the partitioning of excess carbon between atmosphere, land and ocean (hence $a_\infty \neq 0$), neglecting ocean sediment-interactions and weathering flux perturbations, is only valid for time scales shorter than about 2,000 years. "

The theoretical issue

In my opinion, the theoretical foundations of this model are sound as long as the substitute IRFs are time-independent and also independent of other state variables. However, the great power of BernSCM emerges when temperature or CO2 dependencies are explicitly allowed in the IRFs. I am not perfectly sure, if the theory behind the IRF approach is still valid in this case, even though the simulations show reasonable results.

I think it is important to stress the fact, that equation (14) works for time-dependent forcings, but for time-independent processes only. The impulse response function r here depends only on the difference $t - t_0$ of the time t at which we are interested in the perturbation $m(t)$ and the time t_0 at which the input $f(t_0)$ came into the system.

The absolute time t_0 is not used by the impulse response function r. Consequently, the underlying black-boxed process which can be modelled by this approach is assumed to be time-independent (has constant coefficients). If now the IRF depends additionally on temperature, the impulse response function needs to "know" the current time t_0 and becomes $r(t', t-t')$. A different forcing function f leads to a different system state which then results in a different IRF. The system is inherently non-linear, even though it looks linear. From the code (and unfortunately only from there) I could see that the IRFs are provided as a set of coefficients a_k and a set of time scales τ_k. Probably these numbers result from an analysis of complex simulations (e.g. HRBM). If so, the numbers come from a non-linear model and this very IRF is representative only for this very model run. In a non-linear setting, a different model run (initial value, temperature, sensitivities) could theoretically lead to a very different IRF which is then going to be ported to BernSCM. The analysis in section 4 shows that this does not have drastic influence here, probably in part because the external emission forcing was chosen to be the same (SRES A2).

Additional to this possible dilemma, the IRF comes with additional numbers for temperature sensitivity. These numbers are used in each time step to adapt the IRF in dependence of current mean surface air temperature. As mentioned above, $r(t - t_0)$ becomes $r(t', t-t')$. In the derivation of equation (A7), which is crucial for the numerical implementation, this leads to a problem. The term

$$R_i = 1/\Delta t \int_{t_{i-1}}^{t_i} r(x)dx$$

Becomes

$$\tilde{R}_i = 1/\Delta t \int_{t_{i-1}}^{t_i} \tilde{r}(t-x, x)dx$$

and the integration becomes much more difficult, in particular if both a_k and τ_k are temperature dependent.

I am not sure whether this problem can be discussed away by numerical means or even by purely theoretical considerations, but at the moment I have a strange gut feeling about this issue.

We agree with the reasoning of the reviewer. The violation of linearity for temperature-sensitive IRF-parameters was pointed out by both reviewers. This issue is possibly related to a badly placed remark on this temperature sensitivity in the context of IRF-integrals on top of page 7. In fact, these parameters can be varied only in the context of a box-model interpretation, and we failed to point this out clearly. The text was modified and extended for clarification:

"The IRF representation is, strictly speaking, only valid if the described subsystem is linear. Then, the response function *r* does not depend on time and on state variables. In the BernSCM, major nonlinearities in the carbon cycle, namely air-sea gas exchange and the nonlinear carbonate chemistry and changes in NPP in response to changes in environmental conditions are treated by separate nonlinear equations (equations (4) and (5)), while surface-to-deep ocean transport of carbon and heat and respiration of carbon in litter and soils are viewed as approximately linear processes using IRFs. Yet ocean circulation and the respiration of carbon from soil and litter is likely to change under global warming, violating assumption of linearity. In practice, the IRF representation remains a useful approximation as long as the impact of associated nonlinearities on simulated atmospheric CO2 and temperature remain moderate.

The interpretation of the IRF representation as a box model provides a starting point for considering nonlinearities in the response. To account for nonlinearities, the response time scales τ_k and the coefficients a_k may be gradually adjusted as a function of state variables such as temperature. As the integral form (13) involves integration over the whole history at each time step, changing parameters along the way would result in inconsistencies. In contrast, the differential or box-model form (21) does not depend on previous time steps. Changing the model parameters from one step to the next thus equates to applying a slightly different model at each time step. Within each time step, the parameters remain constant, and the solution for the linear case applies. As time steps are small compared to the whole simulation, this discretization yields accurate results, which is confirmed by the close agreement between the different time resolutions shown in Figure 6 (formerly 4).

Varying coefficients have been successfully implemented and tested for the HRBM land component and its decay IRF (Meyer et al., 1999). In this way, the enhancement of biomass decay by global warming is captured (s.a. Appendix A and section 3.1). In such a modification, the advantage of the IRF and the equivalent box model representation - the faithful representation of the characteristic response time scale of a model system - is largely maintained, while at the same time the impact of time and state-dependent system responses on simulated outcomes is approximated.

Technical corrections

The following corrections were all incorporated as suggested. Specific clarifications are provided below for a few points.

• in general: Punctuation around equations is missing very often, in particular full stops or commas after the equations when necessary.

Done. Comma added where appropriate (but not full stops)

• in general: Some abbreviations are never introduced, e.g., HRBM, HILDA. Sometimes the explanation of the abbreviation comes late in the text. This happens in particular when reading the figure captions and figures are referred to in more places. Maybe this is hard to circumvent without destroying the text flow.
• There is a mix of British (analyse) and American ("behavior") English.
• in general: It is difficult to find out which constant means what since tables 1 and 2 are not complete. Some terms are explained in the text, some in the tables.

The tables were revised and completed. Table 2-4 now list all model parameters.

• page 1, line 14: "in an spatially"
• page 2, line 34: "of BernSCM a an IAM component"
• page 2, line 35: "managment"

• page 3, line 1: “cycle assessments(Levasseur et al., 2016)”:
space before parenthesis

• page 3, line 26: What does “LULUC” stand for?
• page 4, line 4: “by” or colon missing at the end?
• page 4, line 7, equation (4): How is φ_{NPP} defined?

Function definitions are now given in a new Appendix 1. The purely symbolic equations 4 and 6 were deleted.

• page 4, line 11: “Ao” instead of A_O, “eps” is probably meant to be “ε”
• page 4, line 14, equation 8: How are the functions ψ and χ defined?

Function definitions are now given in a new Appendix 1. The purely symbolic equations 4 and 6 were deleted.

• page 5, line 1: “and the separation of SAT from radiative equilibrium”:
Does “separation” here refer to the difference between 1 and the ratio $\frac{\Delta T}{\Delta T^{eq}}$? To me the word “separation” is rather confusing in this context.

Replaced “separation” by “deviation”

• page 5, line 9, equation (12): I could not find a description of $p_{CO_2\,A0}$ anywhere.
Typo?

Typo in line 10 corrected

• page 7, line 5, equation (21): The lower limit of the integral should be t_0 instead of 0.

All integration limits have been changed to -infinity.

• page 7, line 15: “a_{O_k}” is called a_k in Figure 1 (blue box). Also in the red box the constants are called a_k. Only in the green box they are called a_{Lk}. Similar problems with τ.

Figure 1 was changed as suggested.

• page 7, line 20: Which model from Table 3 is here referred to? The Bern2.5D or the 4-box Siegenthaler and Joos?

Reference added on line 21

• page 8, lines 9-10: “here, the IRF substitutes for the HILDA ocean model, and the HRBM land biosphere model are used for the standard setup”: It took me a while to understand this phrase. Maybe an additional “for” in front of “the HRBM” and omitting the comma are helpful?
• page 8, lines 13-14: “and the dependency of land C on temperature (f_{decay}) increases with warming, eq. (2))”: From equation (2) I cannot see what happens with warming. Going to equation (19), I can see that it depends very much on f_{NPP}. This is defined in equation (4) and depends heavily on φ_{NPP}, which is not explained at all.

- Term “eq. (2)” deleted
- Functions are now given in the MS

• page 9, line 4, equation 25: An interchange of β_O and β_L on the right hand side makes it better comparable with the left hand side and the following text.
• page 9, line 7-8: “β is the change in carbon stored (in GtC)”:
Following Table 3, β has the unit GtC/ppm and in Friedlingstein et al. (2006) it is referred to as ”sensitivity of land carbon storage to atmospheric CO_2“. If I understood equation (25) correctly, this is a more precise and less confusing description. The same holds for gamma.
• page 9, line 26: I did not immediately recognise “airborne fraction” as a technical term. Maybe a short explanation could avoid confusing non expert readers.
• page 10, line 1: “(Figure. 3)”

• page 10, line 1-2: Why different units?
• page 10, lines 3-5: Looking at Figure 3 (upper panel), after 100 years all simulated values are greater than 0.3. Where does the value 0.3 from the text come from?

Value changed from 0.3 to 0.4

• page 10, lines 5-7: "For AF simulated with BernSCM, the standard coupled setup is close to the IRFMIP multimodel median, but the BernSCM uncertainty range is asymmetric. The IRFMIP multi-model range is similarly asymmetric.:" The word "but" confuses me, because the "IRFMIP multi-model range is similarly asymmetric".

Text changed.

• page 12, line 21: "structural simplicitly"
• page 13, line 15, equation (A4): The integral limits are interchanged. They do not change with the parameter transformation, because $dt_0 = -dx$ makes for a second sign change.

The equation has been removed in response to reviewer 2.

• page 15, line 2, equation (A14):
Maybe it is better to write (Δt^2).

Equation removed in response to reviewer 2

• page 15, line 13: "explicite"
• page 15, line 20: "Equations (A1,A2)"
Space between A1 and A2?
• page 16, line 13, equation (A20):
Maybe $(m_{S_n} - m_{S_{n-1}})$ is correct? I am not sure.

The equation in the manuscript is correct. An increase of CO2 in the surface layer reduces C uptake; the term belongs to $-\varepsilon\, p^{CO2}_{S,n-1}$ (hence negative).

• page 17, line 12: "explicite"
• page 17, line 19: "Equations (11,10, A28)"
Space between equation numbers?
• page 17, line 24, equation (A30): Is it correct that f_{O_n} appears twice in this formula?

It should be $f_{O_{n-1}}$ in one case. Thank you for spotting this mistake.

• Figure 4: For me it is impossible to differentiate between dashed and dashed-dotted lines here. Maybe a different colour/line-style scheme could help here. Since the differences resulting from the use of different numerical schemes are almost invisible anyway, one could even go without trying to make them visible and simply mention that the differences are small. On the other hand, the point the authors want to emphasize here, is that due to the very small differences, the fastest scheme can be implemented. This leads to the entire appendix and the Fortran implementation. So it is rather an important point.

Figure 4 (now 6) has been updated as suggested.

• Table 1: "f_A: net flux to atmosphere flux"
• Table 1: "f_{deep}: Flux mixed layer to deep":
Why does "Flux" start with capital F? Missing "ocean" at the end?
• Table 2: Capitalization of first word in second column inconsistent?
• Table 2: From the units I think c_p should be called "specific heat capacity".
• Code: Why is in the file parLandHRBM.inc the first weight negative? If I understood correctly, those weights are the a_{L_k} values which here nicely sum to one, but how do you distribute a negative share of incoming carbon?

This follows from the original reference (Meyer et al., 1999). The response of several thousand reservoirs (boxes) of the original HRBM model is approximated by five boxes arranged in parallel. The timescales are 0.2, 1.4, 8.9, 74.1 and 253.7 yr. The pool with the short overturning time of 0.2 yr exhibits a very small (-1.3 GtC) negative inventory at equilibrium as noted by the reviewer. This solution is typical for models where material is transferred through successive reservoirs. The two boxes with the smallest time scales of 0.2 yr and 1.4 yr may be combined for decadal to century scale scenario calculations. They yield together a positive inventory of more than 30 GtC.

Anonymous Referee #2

This paper by Strassmann and Joos presents the reimplementation of the Bern Simple Climate Model (BernSCM), a reduced form model of the anthropogenic perturbation of the carbon-climate system. This is a historic model for the community, since it and its offspring have been used since the IPCC SAR. This new implementation is useful for the community, especially as this paper focuses on transparency and the model's codeis provided in an open-source format.

Thank you

Being an old model, the BernSCM ignores some relatively recent developments in climate sciences and modeling. In itself, it is not so much of a problem, as the authors leave the door open to further development of the model, both in the manuscript and in the model's code. However, mention and discussion of these caveats is required, especially regarding some specific points I develop below.

Please see our answers below to the specific points raised by the reviewer.

I also believe that the paper could benefit from a more careful rewriting, especially for some sections that I had to read several times – and I am still not 100% sure of what is done in some parts of the paper! In all honesty, some parts give the impression that the authors were in a rush for writing the paper.

We added additional text and references as detailed below to help the reader and the reviewer to better understand the content of the manuscript. We note that reviewer 1 came to a different conclusion and states that "*Apart from some minor exceptions, the manuscript and in particular the appendix and the provided Fortran code of the model are carefully prepared*."

The following text was added in the introduction (p2, l17 of submitted MS); please see also answer to the specific points:
"The BernSCM (Figure 1) is designed to compute decadal-to-millennial scale perturbations in atmospheric CO_2, in climate and in fluxes of carbon and heat relative to a reference state, typically preindustrial conditions. The uptake of excess, anthropogenic carbon from the atmosphere is described as a purely physico-chemical process (Prentice et al., 2001). As in pioneering modeling approaches with box-type (Revelle and Suess, 1957;Oeschger et al., 1975) and general ocean circulation models (Sarmiento et al., 1992;Maier-Reimer and Hasselmann, 1987) modification of the natural carbon cycle through potential changes in circulation and the marine biological cycle (Heinze et al., 2015) are not explicitly considered. While such modifications and their potential socio-economic consequences are vividly discussed in the literature (Gattuso et al., 2015), associated climate-CO2 feedbacks are likely of secondary importance. Estimated uncertainties in the marine carbon uptake due to climate change, including warming-driven changes in CO_2 solubility, are found to be smaller in magnitude than uncertainties arising from imperfect knowledge of surface-to-deep physical transport (see Figure 2d,e in (Friedlingstein et al., 2006). The exchange of CO_2 between the atmosphere and the surface ocean is described by two-way fluxes, from the atmosphere to the surface ocean and vice versa, and the net flux of CO_2 into the ocean is proportional to the air-sea partial pressure difference. CO_2 reacts with water to form carbon and bicarbonate ions (Dickson et al., 2007;Orr et al., 2015) and acid-base equilibria are here described using the well-established Revelle factor formalism (Siegenthaler and Joos, 1992;Zeebe and Wolf-Gladrow, 2001). The first order climate-carbon feedback of a decreasing solubility in warming water is considered. Surface-to-deep exchange, the rate limiting step of ocean carbon and heat uptake, is described using an IRF. On time scales of up to a few millennia processes associated with ocean sediments and weathering can be neglected. In such a "closed" ocean-atmosphere-land biosphere system, excess CO_2 is partitioned between the ocean and the atmosphere and a substantial fraction of the emitted CO_2 remains in the atmosphere and in the surface

ocean in a new equilibrium (Joos et al., 2013). This corresponds to a constant term (infinitely long removal time scale) in the IRF representing surface-to-deep mixing. On multi-millennial time scales, excess anthropogenic CO_2 is removed from the ocean-atmosphere-land system by ocean-sediment interactions and changes in the weathering cycle (Archer et al., 1999;Lord et al., 2016) and the IRF is readily adjusted to account for these processes, important for simulations extending over many millennia.

The BernSCM simulates global mean surface temperature and the heat uptake by the planet. The latter is equivalent to the net top-of-the-atmosphere energy flux. Changes in the Earth's heat storage in response to anthropogenic forcing are dominated by warming of the surface ocean and the interior ocean (Stocker et al., 2013b) due to their large heat capacity in comparison with that of the atmosphere and their large thermal conductivity in comparison to that of the land surface. Consequently, the atmospheric and land surface heat capacity is formally lumped with the heat capacity of the surface ocean in the BernSCM. The uptake of heat by the ocean (or planet) is, as for carbon, formulated as a two-way exchange flux. The flux of heat from the atmosphere into the surface ocean is taken to be proportional to the radiative forcing resulting from changes in CO_2 and other agents (Etminan et al., 2016). The upward loss of heat from the surface is proportional to the product of the simulated surface temperature perturbation and the (prescribed) climate sensitivity, λ, (Siegenthaler and Oeschger, 1984;Winton et al., 2010).

As with carbon, surface-to-deep transport is the rate limiting step for ocean heat uptake and thus for the adjustment of surface temperature to radiative forcing. This transport is key to determine the lag between realized warming and equilibrium warming (Frölicher and Paynter, 2015). Again this transport is described using an IRF. This IRF encapsulates the finite volume of the entire ocean. It also represents the range of transport time scales associated with advection, diffusion and convection ranging from decades for the ventilation of thermocline to more than a millennium for deep Pacific ventilation as evidenced by transient tracers such as CFCs and radiocarbon (Olsen et al., 2016). The simulated surface ocean temperature perturbation, taken as a measure of global mean surface air temperature change, may be combined with spatial patterns of change in temperature, precipitation or any other variable of interest to compute regionally explicit changes ((Joos et al., 2001;Stocker et al., 2013a;Hooss et al., 2001) (Figure 1).

Non-CO_2 radiative forcing may be prescribed, e.g., following estimates from complex climate-chemistry models (Myhre et al., 2013) or from simple emission driven non-CO2 chemistry-radiative forcing modules (Smith et al., 2017;Joos et al., 2001) and reconstructions of solar and volcanic forcing (Jungclaus et al., 2017;Eby et al., 2012) and considering the forcing efficacy of non-CO_2 agents relative to CO_2 forcing (Hansen et al., 2005). Climate sensitivity characterizing the response to radiative forcing, is a free parameter in the BernSCM. Climate sensitivity may change under increasing warming, particularly in high emission scenarios (Pfister and Stocker, 2017;Geoffroy et al., 2012a;Gregory et al., 2015). Here, climate sensitivity is assumed to be time-invariant and a potential state dependency of climate sensitivity is not considered. This may be changed when more solid information on state dependency becomes available or for the purpose of sensitivity analyses. Similarly, ocean heat uptake efficacy (Winton et al., 2010), influencing the atmospheric temperature response to ocean heat uptake forcing, is set to one here."

We added the following text in the discussion section on page 11:"The BernSCM does not explicitly distinguish between surface atmosphere and surface ocean temperature to compute global mean surface air temperature perturbation. This is in contrast to some energy balance calculations used to analyze results from state-of-the-art Earth System Models (e.g., (Geoffroy et al., 2012b). The BernSCM approach follows earlier work of (Siegenthaler and Oeschger, 1984). It is further guided by the similarity in reconstructions of marine night time air and sea surface temperature perturbations (Stocker et al., 2013b) that are consistent with the short, monthly relaxation time scale for air-sea heat exchange. The focus of the BernSCM is on the representation of the transport of heat from the surface into the thermocline and the deep ocean on decadal to multi-century time scales, while information on seasonal and spatial changes such as on land-sea air temperature differences or polar amplification may be obtained by applying suitable spatial perturbation patterns as derived from state-of-the-art models."

So I fear publication can only be recommended if the few scientific issues I raise below are answered/discussed, and if the text itself is improved.

Major points:

1. My first point concerns the use of the same IRF parameters for the ocean carbon cycle and the climate system. If I understand it well, the function r_O is the same for determining the ocean C sink and the temperature change, e.g. in equations (15) and (16). Although it would seem intuitive to use the same function, because – obviously – we are talking about the (same) world's ocean in both cases, I see several issues in doing so.

First, I am not quite sure one can assume that the diffusion process is the same for heat and for actual material such as carbon. (The assumption seems more reasonable for convection.) But more importantly, the biological pump does not affect heat transport, while it does for carbon. (Although, I am not sure whether there was a biological pump at all in the models used to calibrate the r_O function – another thing worth being mentioned.)

The model does not include a representation of the marine biological cycle as discussed in the answer above.

Second, global patterns of heat uptake vs. carbon uptake are different. This means that one unit of incoming f_O is dispatched differently than one of f_{HO}, at the scale of the global surface ocean. Therefore, it is likely that each of them is affected differently by the oceanic circulation. For the climate response, it is also known that this pattern affects an internal feedback (the ocean heat uptake feedback) in a way that changes the apparent time-scales of the climate response, see e.g. Geoffroy et al. (2013b) and references therein.

We modified the text in the discussion (p11, l 31 of submitted MS) to read.
"Ocean transport is known to vary under climate change with some consequences for heat and carbon uptake (Joos et al., 1999). Here, we applied time-invariant ocean transport parameters ($a_{o,k}$, $\tau_{o,k}$). It is in principle possible to represent temperature dependency of ocean transport in a similar way as it is done for the climate dependency of heterotrophic respiration for the HRBM land biosphere substitute model (Meyer et al., 1999). In the current BernSCM version, the same IRF parameters are applied for the transport of carbon and heat from the surface ocean to the interior ocean. Thereby, it is implicitly assumed that the spatial pattern of change is the same for temperature and carbon. This appears to be a reasonable first-order approximation on decadal-to-century timescales as perturbations in temperature and carbon show similar patterns with decreasing perturbations from the surface to depth. In future efforts, one may differentiate the ocean IRF for heat and carbon, in particular when more information from long-term multi-century to millennial-scale ESM simulations becomes available. The application of the same IRF for carbon and heat in individual model runs implies that modelled carbon and heat transport tend to be physically consistent. In contrast, some other simple models employ different transport parameters for heat and carbon and varied these parameters independently in probabilistic studies."

Third, the typical climate IRF only has two time-scales (e.g. Geoffroy et al., 2013a), and these are quite different from the time-scales from Joos et al. (1996). And, maybe more importantly, the typical two-box climate model implied by the typical climate IRF (Geoffroy et al., 2013a) includes a bidirectional exchange of energy between the surface and deep oceans. This is not the case in the assumed formulation presented here. There are some fundamental reasons for not having this bidirectional exchange for the carbon cycle: the so-called 'ocean invasion' is a slow process, and ultimately there is a sink of C in the deep ocean that involves geological chemical reactions (and time-scales). But can this be also applied to the climate system and heat transport?

The statement by the reviewer is incorrect. The BernSCM employs bidirectional exchange of carbon and heat. For example, rearranging equation 10 of the submitted MS yields the ocean heat uptake as difference between Radiative Forcing, *RF*, and the response, $\lambda\ \Delta T$, with λ the climate sensitivity in W m^{-2} K^{-1}:

$$F_o^H = RF - \lambda \cdot \Delta T \quad \text{with} \quad \lambda = RF / \Delta T_{eq}$$

Similarly, the net flux of carbon into the ocean is the results of an uptake flux proportional to the partial pressure (or more correctly the fugacity) of CO_2 in the atmosphere and a return flux proportional to the partial pressure of CO_2 in the surface ocean (Eq. 5). This feature is now discussed in the discussion (see answer to general comment above).

In addition, the following text is added (p5, line 7 of submitted MS):
"Equation (10) describes ocean heat uptake as difference between *RF* and the climate systems response, $\lambda\ \Delta T$, with $\lambda = RF/\ \Delta T_{eq}$ climate sensitivity expressed in W m^{-2} K^{-1}.

In the discussion, we added (p 11, l 32):
"The BernSCM model may be extended to model perturbation in the signatures and exchange fluxes of the carbon isotopes ^{13}C and ^{14}C as demonstrated in earlier work (Joos et al., 1996). This was not implemented here to keep the code as simple as possible and as most potential users are likely concerned with the evolution of climate and atmospheric CO_2."

There is no fundamental reason to not consider the bi-directional flux for carbon. Neglecting the return flux of excess carbon from the surface ocean to the atmosphere corresponds to assuming an infinitely large ocean and infinitely fast mixing between the surface and the deep ocean. Such an assumption leads to erroneous and misleading results.

Further, there is no reason to assume that heat transport by the ocean is governed by two time scales only.

We added the following text in the discussion p11, l 32:
"A distribution of time scales applies to ocean transport processes as evidenced by observations of transient and time dependent tracers such as chlorofluorocarbons, bomb-produced and natural radiocarbon and biogeochemical tracers (Olsen et al., 2016;Key et al., 2004). This continuum is sometimes approximated by one time scale, also termed heat uptake efficiency (e.g. (Gregory et al., 2009)) by two time scales, as in (Geoffroy et al., 2012b). The one to two time scale approximations were used to analyze relatively short Earth System Model simulations that do not yet reveal the multi-century response time scales of the deep ocean. We note that the equivalent ocean depth of the simple energy balance model of (Geoffroy et al., 2012b) for their AOGCM ensemble is only 1182 m compared to a mean ocean depth of about 3800 m. The ocean IRFs used in the BernSCM are derived from long simulations with ocean-only or simplified models. The range of distinct time scales used to construct the IRF faithfully approximates the sub-annual to multi-century response continuum of the parent models as shown in earlier work (Joos et al., 1996). Further, the BernSCM IRF model represents the heat capacity of the entire ocean."

Therefore, I believe this is an assumption made by the authors that r_O can be applied to the climate system as well. Despite all of the above, it may still be acceptable. But it should be presented as such, and it also warrants a discussion in the text. Additionally, the response to a step of radiative forcing (typically 4x CO2) of this climate model has to be compared to that of more complex models. I strongly suggest adding a (sub)figure in which the BernSCM climate response is compared to that of CMIP5 models, taken e.g. from Geoffroy et al. (2013b). This would complement figure 3.

The new Figure 3 demonstrates that the BernSCM temperature response falls well within the response of more complex models.
Thank you for suggesting this additional figure. We carried out additional simulations where CO_2 is prescribed to increase either exponentially (linear increase in RF) or abruptly to reach 4xCO2 (Figure 3). We compare the outcome in terms of realized warming fraction with the compilation by Frölicher and Paynter (2015) for EMICs and CMIP5 AOGCMs (please refer to newly added section 3.2).

2. I was very troubled by section 3 and how the carbon-climate feedbacks are represented/ investigated with BernSCM, in relation with C4MIP. At first, I thought BernSCM was trying to emulate the C4MIP models' sensitivities (which would have been a new feature).

In the end, my understanding is that the uncertainty range provided e.g. in table 4 is obtained by combining variations of: (i) the ocean model, 2 options; (ii) the land model, 2 options; (iii) the experimental setup, i.e. coupled/uncoupled/Tonly/Conly, 4 options. That is a total of 2× 2× 4=16 configurations. But my concern, here, is that I think that turning a process on or off can hardly be considered a new configuration of the model. Therefore, although the results shown e.g. in figure 3 or 4 are interesting, the ranges provided in table 4 are artificial and misleading.

More generally speaking, the text should make it clear that there are not many parameterizations available for the model, and so it does not cover the full range of existing multi-model uncertainty (and therefore, it cannot be used in a probabilistic fashion). Again, it is not so much of a problem in itself, but this has to be made very clear.

Done. We incorporated section 3 into section 4 "Illustrative simulations with the BernSCM" and reorganized section 4 to avoid a potential misunderstanding. This is done by adding the following subsection headings in section 4:
4 Illustrative simulations with the BernSCM
4.1 Model setup for sensitivity analyses and uncertainty assessment
4.2 Fraction of realized warming and idealized forcing experiments
4.2 Impulse response experiment
4.3 Carbon cycle-climate feedbacks
In section 4.1 we included the existing text from p 8 line 5 to 29 of section 3 followed by the following text: "We performed simulations with these different setups. In section 4.2, we probe the time scales of the temperature response in simulations where atmospheric CO_2 is abruptly (instantaneously) quadrupled or by increasing CO2 radiative forcing linearly within 140 years. In section 4.3, we probe the response of the coupled system to a pulse-like release of 100 GtC into the atmosphere. Finally in section 4.4, we analyze carbon cycle-climate feedbacks relying on simulations over the industrial period and for the SRES A2 scenario. BernSCM results are compared with the results from three multi-model intercomparison projects: the Climate Model Intercomparison Project 5 (CMIP5) with results as summarized by (Frölicher and Paynter, 2015); an analysis of carbon dioxide and climate impulse response functions …"

Section 3.2 describes the results for the 4xCO2 simulations requested by reviewer 2 (please see the revised manuscript (attached).

Section 4.3 on IRF experiments includes the text from page 9, line 25 to p10, l17 of the originally submitted MS

Section 4.4 on feedback analyes includes the text from p8, l30 to p9, l16 followed by the paragraph on p10, l18 to p10, l22. Then the paragraph on p9, l17 to l19 is added before continuing with the text from p10, l23 to p11, l2.

The caveat that only a limited set of model versions is available is already explicitly dicussed in section 5, p11, line 27 or original MS ("Currently, a limited set of substitute models is available …")

We somewhat disagree with the reviewer on the potential use in probabilistic assessment. As with any model, the parameters of the BernSCM (including those of the land and ocean IRF) can be varied using Latin Hypercube sampling or similar and simulation results weighted with observational constraints as for example demonstrated by Steinacher et al. (2013). We note that spatially-explicit and dynamic Earth System Models of Intermediate Complexity offer a much greater potential in probabilistic assessment than the current crop of simple models.

3. I have some trouble with the way the solving of the differential system is presented, but more importantly I believe there is a mistake with how the temperature-dependent parameters are implemented.

I am not convinced by the lengthy demonstration of appendix A1. Equations (A2) and (A3) are the 'results' of this section, and I believe the following demonstration is not needed. Equations (A2) and (A3) can simply be obtained by using the 'exponential integrator' method to solving a first-order differential system. Although not everyone may know this method, it could be summed up in one or two equations (and appropriate references) rather than be re-demonstrated from scratch.
Equations (A2) and (A3) are simply obtained by 'reminding the reader' that the solution to the differential system:

$$\frac{\mathrm{d}m}{\mathrm{d}t} = -\frac{m(t)}{\tau} + aF(t) \tag{1}$$

can be discretized by multiplying by $\exp(\delta t/\tau)$ and integrating between t_n and $t_{n+1} = t_n + \delta t$:

$$m_{n+1} = \exp-\frac{\delta t}{\tau} m_n + \int_0^{\delta t} \exp(-\frac{\delta t - s}{\tau}) aF(t_n + s)\mathrm{d}s \tag{2}$$

where $m_{n+1} = m(t_{n+1}) = m(t_n + \delta t)$, and $\delta t$is the time step.

The above equation is exact, but can hardly be solved. It is usual to assume that $F$is constant over the small time period of δt, which leads to the solution:

$$m_{n+1} = \exp(-\frac{\delta t}{\tau}) m_n + \tau(1 - \exp(-\frac{\delta t}{\tau})) aF(t) \tag{3}$$

which is basically equation (A2) and (A3) combined. $\tilde{}$ $t$remains to be chosen, e.g. to be t_n (forward method), t_{n+1} (backward), or any other fancier method possible. When assuming δt= 10 yr and *a* $F(t)$ is linear between t_n and t_{n+1}, one immediately finds the δt^2 equations.

So far, no fundamental problem with the authors' equations and text. I just believe it could be written in a more efficient and straightforward way. But a problem arises when one assumes that the time-scale τvaries with time (through e.g. temperature) so that we have in fact $\tau = \tau_0 + \Delta\tau(t)$. The exponential integrator method can still be applied, albeit by using τ_0 and not τin the exponential function.
To do so, it is easier to rewrite the differential equation as:

$$\frac{\mathrm{d}m}{\mathrm{d}t} = -\frac{m(t)}{\tau_0 + \Delta\tau} + aF(t) \tag{4}$$

$$= -\frac{m(t)}{\tau_0} + -\frac{m(t)\Delta\tau}{\tau_0(\tau_0 + \Delta\tau)} + aF(t) \tag{5}$$

which completely changes the exponential integrator form:

$$m_{n+1} = \exp-\frac{\delta t}{\tau_0} m_n + \int_0^{\delta t} \exp(-\frac{\delta t - s}{\tau_0}) aF(t_n + s)\mathrm{d}s + \int_0^{\delta t} \exp(-\frac{\delta t - s}{\tau_0}) \frac{m(t_n + s)\Delta\tau(t_n + s)}{\tau_0(\tau_0 + \Delta\tau(t_n + s)} aF(t_n + s)\mathrm{d}s \tag{6}$$

leading to:

$$m_{n+1} = \exp(-\frac{\delta t}{\tau_0}) m_n + \tau(1 - \exp(-\frac{\delta t}{\tau_0})) aF(t') + (1 - \exp(-\frac{\delta t}{\tau_0})) m(t') \frac{\Delta\tau(t')}{\tau_0 + \Delta\tau(t')} \tag{7}$$

The latter equation raises the issue that it is virtually impossible to use with a backward approach since $\Delta\tau(t_{n+1})$ is not known. But a bigger issue is that, if I understand it correctly, the authors do not use this equation nor any equivalent. I believe they simply apply the equation of the case with constant ⌒but with a value of ⌒that changes through time. That is, they use the following equation:

$$m_{n+1} = \exp(-\frac{\delta t}{\tau_0 + \Delta\tau(t)})m_n + (\tau_0 + \Delta\tau(t))(1 - \exp(-\frac{\delta t}{\tau_0 + \Delta\tau(t)}))aF(t') \quad (8)$$

instead of the one above.
Unless the authors can prove the difference between the two is negligible, I am afraid there is a fundamental mistake in the solving of the model.

We agree that appendix A1 was unnecessary lengthy. The section was rewritten in a simpler way according to the reviewer's suggestions. Thank you.

The violation of linearity for temperature-sensitive IRF-parameters was pointed out by both reviewers. This issue is possibly related to a badly placed remark on this temperature sensitivity in the context of IRF-integrals on top of page 7. In fact, these parameters can be varied only in the context of a box-model interpretation, and we failed to point this out clearly. The following text was added to clarify this point (please see the response to reviewer 1 for the full text changes):

"The interpretation of the IRF representation as a box model provides a starting point for considering nonlinearities in the response. To account for nonlinearities, the response time scales τ_k and the coefficients a_k may be gradually adjusted as a function of state variables such as temperature. As the integral form (13) involves integration over the whole history at each time step, changing parameters along the way would result in inconsistencies. In contrast, the differential or box-model form (21) does not depend on previous time steps. Changing the model parameters from one step to the next thus equates to applying a slightly different model at each time step. Within each time step, the parameters remain constant, and the solution for the linear case applies. As time steps are small compared to the whole simulation, this discretization yields accurate results, which is confirmed by the close agreement between the different time resolutions shown in Figure 6 (formerly 4)."

4. I believe the model should be completely described in the paper. I mean: formulations for e.g. functions $p\mathrm{CO_2s}$, ϕ, χ, as well as all the parameter values should be given. The model is relatively simple, and there are not that many parameters. Even if the values can be accessed in the code, the fact that this paper is a model description makes it necessary to be as exhaustive as possible.

Done. Parameters for the formulations are now given in the appendix and corresponding tables.

Minor points:
p. 1 (sec. 1): SCMs have many more usages than what is given here. Generally speaking, I find that the citations of this paper are too self-centered. I think everyone acknowledges the importance of the original Joos et al. (1996) paper, but much has been done since then regarding IRFs.

Done. We have complemented the list of potential applications on p. 2 l.2 and provide references to other simple box-type and IRF models: "Another application of simple models (e.g., (Enting, 1990;Enting et al., 1994;Oeschger et al., 1975;Siegenthaler and Oeschger, 1984;Huntingford et al., 2010;Smith et al., 2017;Tanaka et al., 2007;Bruckner et al., 2003;Joos and Bruno, 1996;Hooss et al., 2001;Urban and Keller, 2010;Good et al., 2011;Wigley and Raper, 1992;Raupach et al., 2013;Boucher and Reddy, 2008) is to compare, analyze or emulate more complex models ((Meinshausen et al., 2011;Raper et al., 2001;Geoffroy et al., 2012a;Thompson and Randerson, 1999;Geoffroy et al., 2012b). Simple models also play a significant role in previous assessments of the Intergovernmental Panel on Climate Change (e.g., (Harvey et al., 1997))."

p. 3, l.13: The "essentially linear behavior" is an assumption of the model.

Done. Text modified to read: "but in first order linear behavior" p.3, l. 16: IRFs are indeed equivalent to box-models, albeit with constant parameters!

Thank you for your confirmation

p.3, l.25: The non-inclusion of LULCC could be discussed a little.

Done. The following text is added: “Human impacts on the land biosphere exchange including land use and land use changes are not simulated in the present version, and treated as exogenous emissions (*e*). These emissions may be prescribed based on results from spatially-explicit terrestrial models.”

p.6 l.10: It is probably better to give all the equations, even if very similar.

Done. Text modified to read: “Similarly, equation (16) closes the heat budget equation (9) for the surface ocean.”

p.6 l.20: “conversation” => “conservation” (probably many typos I missed...)

Done. Typo corrected

p.6 l.22: I don’t think it is 10^4 or 10^5 kyr. Unit is probably yr.

Done. Unit correct to yr

p.7 l.2: At this stage, it is very unclear whether the response based on HRBM is a usual linear IRF calibrated with climate-carbon feedbacks on, so that those are linearized within the IRF, or if the time-scales of the response are indeed interactively changed by temperature during the simulation. Note also that I don’t think the name “IRF” can be given to a model with time-varying parameters. I believe an IRF is the integrated form of the differential equation, which can be obtained only when the parameters do not vary with time. When they do, there is no integrated form, and the model is just a box model.

Done. Text modified to read: “In contrast, the parameters of the IRF-derived box model representation of the HRBM land biosphere are interactively modified during the simulation by a temperature dependent factor. In this way, the enhancement of biomass decay by global warming is captured (s.a. Table 3 and section 3).”

p.7 l. 13: Similarly, I would question the fact that the equation shows that IRF and box model are equivalent. I think they are per definition. The only difference being that one is the integrated form of the other.

Done. Text changed to read: “Equation (22) represents the IRF by a box model, ...”

p.7 l. 20: Can cite Li et al. (2009) who provide a nice discussion on the (over)interpretation of those parameters.

Done. Reference to (Li et al., 2009)added.

p.7 l.25: It is more than ‘they can be viewed’. Per construction, IRFs show the exponential eigenmodes of the system they are calibrated upon. Raupach (2014) or Enting (2007) provide some insights on this.

Done. Text modified to read: “.. are equivalent to .. and may also be interpreted in the context of the Laplace transformation (Raupach et al., 2013;Enting, 2007)”.

p.8 (sec. 3): I really find this section difficult to apprehend. It would benefit from some re-organizing, e.g. with a subsection on the beta/gamma framework, and then one on what it gives when applied to BernSCM. This is also the part that made me wonder whether C4MIP models were emulated or simply used for comparison.

Done. Please see answer to major comment 2.

p. 8, l.16: Table 3 does not provide any parameter value

All functional forms and parameters are now given in the new Appendix A and corresponding tables.

p.8, l. 28-29: Please, name those simulations "T-only" and "C-only". The dash makes a lot of difference when reading the text that follows!

Done.

p.9, l.9: I don't think alpha is the "transient climate sensitivity" in the usual sense. Find another name.

Done. Text modified to read "linear transient climate sensitivity to CO_2 (^{o}C per ppm)" as in Friedlingstein et al. (2006).

p.9, l.10: Which original paper?

Done. Reference to Friedlingstein et al. (2006) added.

p.9, l.31: The "combinations" remain quite unclear.

The combinations are given in the caption of Figure 4 (formerly 3).

p.10, l.1: Inconsistent temperature units (this is in the whole paper).

Done. Notation adjusted.

p.10, l.2: More important comment related to my first major points. The choice of a climate sensitivity does not affect the time-scales of the climate response. However, it is known that a higher climate sensitivity implies a slower climate system (e.g. Baker Roe, 2011).

Done. Please see the new figure to gauge the time scales of the climate system response in the model

p.10, l.9-10: The last bit of this sentence is very uninformative.

Done. Text deleted.

p.10, l.12: 3.2K

Done. Unit added.

p.10, l.15-17: I believe the fundamental reasons exposed in my major point number 1 also explain a lot, here. Hence the need to compare the climate response alone, and not coupled to the carbon cycle as in figure 3.

A new figure that compares the climate response in isolation is added (new Figure 3).

p.10, l.23: Those sensitivities are not defined

Done. Sentence deleted in the revision process

p.10, l.29-32: I don't see the point of those sentences. Yes, the obtained sensitivities are zero. But this is per construction, since the uncoupled cases are used to investigate the sensitivity. This relates to my major point 2.

We prefer to keep this information in the text. Please see above for the response to major point 2

p.11, l.4: I believe it is 0.5K, according to figure 4. Also these values are for a fixed climate sensitivity. So I wonder how informative they are.

Done. Text on page 11 line 3 to 5 deleted p.11 (sec. 5): I don't find all the discussion about BernSCM/C4MIP very convincing, for the reason already exposed above.

Done. We shortened the discussion and deleted the text from line18 to 23.

p.12, l.1: Yes, but that requires building EOFs on more complex models. Mention and citations needed here.

Text modified to read: "A potential future application .."

p.12, l.3: Note that regarding precipitation (and likely cloud cover as well), we now know that the response is forcing dependent (e.g. Shine et al., 2015; and referencestherein).

Done. We added the following text "Patterns of change are generally similar across models for temperature, whereas patterns in precipitation are more uncertain and show greater variability between models (Knutti and Sedlacek, 2013) and are forcing dependent (Shine et al., 2015). We also note that natural variability strongly influences the space-time evolution of climate change (Deser et al., 2012). Patterns may be scaled with changes in global mean surface air temperature as indicated in Figure 1 or dependencies on radiative forcing may be considered (Shine et al., 2015)."

p.12, l.10: Yes. But simple models usable in a probabilistic fashion already exist out there.

Done. Text added ", although more sophisticated models are available for observation-constrained probabilistic quantification of climate targets (Steinacher et al., 2013;Holden et al., 2010;Steinacher and Joos, 2016)"

p.12, l.23: GWPs and other metrics require inclusion of non-CO2 species. So I'm not sure the sentence here is relevant.

Text clarified to read: " .. by applying emissions- or concentration-driven simulations."

p. 12, l.26: I don't like the use of "fixed", here. It is e.g. not influenced by external factors such as climate change.

Done. "fixed" replaced by "ocean transport not influenced by climate change."

p. 13 (sec. A1): As I wrote in my major points 3, I believe this section could be more straightforward.

We agree. The section was rewritten in a simpler way according to the reviewer's suggestions. Thank you.

p. 13 (sec. A2): This section is awfully complicated! It makes me wonder about several things, and I could not find the answer. Couldn't a solver be used for the backward method? Is the backward method solved with an exact solution, or is the method proposed an approximation? Does it have to be that complicated?
Also, I find the equations extremely difficult to follow. There are four (!!) levels of notation: U,V,Wrefer to p_{fk} which refer to A_k which refer to the original parameters $\frown_k$ and a_k. I am convinced this part could be written (and implemented in the code?) in a much simpler way

We agree that a solver would be simpler and the equations are complicated. However, we decided for this solution to make the model self-contained and more portable, and we included the equations as such for completeness. The equation systems were solved with a symbolic mathematics software to minimize the risk of mistakes. The number of notation levels has been reduced by the simplification of section A2 (formerly A1).

p.12 (sec. A3): Again, not completely clear how the climate-carbon feedback is implemented. See major point 3.

The section now refers to the formulas for the variable HRBM parameters listed in the new appendix A1.

p.25 (fig. 3): A representation of the land and ocean fractions could be provided. Also, see major point 1: the climate response alone should be shown somewhere (be it within figure 3 or separately).

Done. Two new figures were added, one showing the climate response for an idealized forcing experiment (Figure 3), and one showing the land, ocean, and airborne fractions for the IRFMIP pulse experiment (Figure 5).

p.26 (fig. 4): Maybe show ranges from C4MIP?

Done as suggested.

p.29 (tab. 3): I don't find this table very informative. Parameter values and functional forms should be provided instead.

Done. A new appendix A was added containing all model parameters and functional forms.

p.30 (tab. 4): Using the words "parameters" is one of the things that made me wonder whether C4MIP models were used as input to BernSCM or just to compare outputs. I would call that e.g. "metrics".

Done. Word changed as suggested.

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The Bern Simple Climate Model (BernSCM) v1.0: an extensible and fully documented open source reimplementation of the Bern reduced form model for global carbon cycle-climate simulations

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Abstract. The Bern Simple Climate Model (BernSCM) is a free open source reimplementation of a reduced form carbon cycle-climate model which has been used widely in previous scientific work and IPCC assessments. BernSCM represents the carbon cycle and climate system with a small set of equations for the heat and carbon budget, the parametrization of major nonlinearities, and the substitution of complex component systems with impulse response functions (IRF). The IRF approach allows cost-efficient yet accurate substitution of detailed parent models of climate system components with near linear behaviourbehavior. Illustrative simulations of scenarios from previous multi-model studies show that BernSCM is broadly representative of the range of the climate-carbon cycle response simulated by more complex and detailed models. Model code (in Fortran) was written from scratch with transparency and extensibility in mind, and is provided as open source. BernSCM makes scientifically sound carbon cycle-climate modeling available for many applications. Supporting up to decadal timesteps with high accuracy, it is suitable for studies with high computational load, and for coupling with, e.g., Integrated Assessment Models (IAM). Further applications include climate risk assessment in a business, public, or educational context, and the estimation of CO_2 and climate benefits of emission mitigation options.

1 Introduction

Simple climate models (SCM) consist of a small number of equations, which describe the climate system in an a spatially and temporally highly aggregated form. SCMs have been used since the pioneering days of computational climate science, to analyse analyze the planetary heat balance (Budyko, 1969; Sellers, 1969), and to clarify the role of the ocean and land compartments in the climate response to anthropogenic forcing through carbon and heat uptake (e.g., Oeschger et al., 1975; Siegenthaler and Oeschger, 1984b; Hansen et al., 1984). Due to their modest computational demands, SCMs enabled pioneering research using the limited computational resources of the time, and continue to play a useful role in the hierarchy of climate models today.

Recent applications of SCMs are often found in research where computational resources are still limiting. Examples include probabilistic or optimization studies involving a large number of simulations, or the use of a climate component as part of a detailed interdisciplinary model. SCMs are also much easier to understand and handle than large climate models, which makes them useful as practical tools that can be used by non-climate experts for applications where detailed spatio-temporal physical modeling is not essential. This applies to interdisciplinary research, educational applications, or the quantification of the impact of emission reductions on climate change.

An important application of SCMs is in Integrated Assessment Models (IAMs). IAMs are interdisciplinary models that couple a climate component with an energy-economy model, to simulate emissions and their climate consequences. Another application of simple models (e.g., Boucher and Reddy, 2008; Bruckner et al., 2003; Enting et al., 1994b; Good et al., 2011; Hooss et al., 2 to compare, analyze or emulate more complex models (Geoffroy et al., 2012b, a; Meinshausen et al., 2011; Raper et al., 2001; Thompson a Simple models also play a significant role in previous assessments of the Intergovernmental Panel on Climate Change (e.g., Harvey et al., 19 The comprehensive scope and sweeping interdisciplinarity of such models raise the challenge of maintaining a high and balanced scientific standard across all model components, especially when human resources are limited. This may apply particularly to the climate component, as IAMs are mostly used within the economic and engineering disciplines. Climate and carbon cycle representation are central parts of an IAM and have been critically assessed in the literature (Joos et al., 1999a; Schultz and Kasting, 1997; Vuuren et al., 2009).

BernSCM is a zero-dimensional global carbon cycle-climate model built around impulse-response representations of the ocean and land compartments, as described previously in Joos et al. (1996a, b); Meyer et al. (1999)Joos et al. (1996a); Meyer et al. (1999). The linear response of more complex ocean and land biosphere models with detailed process descriptions is captured using impulse-response functions (IRFs). These IRF-based substitute models are combined with nonlinear parametrizations of carbon uptake by the surface ocean and the terrestrial biosphere as a function of atmospheric CO_2 concentration and global mean surface temperature. Pulse response models have been shown to accurately emulate spatially resolved, complex models (Joos et al., 1996a, b; Meyer et al., 1999; Joos et al., 2001; Hooss et al., 2001a)(Joos et al., 1996a; Joos and Bruno, 1996; Meyer et al., 19

BernSCM (Figure 1) is designed to compute decadal-to-millennial scale perturbations in atmospheric CO_2 in climate and in fluxes of carbon and heat relative to a reference state, typically preindustrial conditions. The uptake of excess, anthropogenic carbon from the atmosphere is described as a purely physico-chemical process (Prentice et al., 2001). As in pioneering modeling approaches with box-type (Oeschger et al., 1975; Revelle and Suess, 1957) and general ocean circulation models (Maier-Reimer and Hasse of the natural carbon cycle through potential changes in circulation and the marine biological cycle (Heinze et al., 2015) are not explicitly considered. While such modifications and their potential socio-economic consequences are vividly discussed in the literature (Gattuso et al., 2015), associated climate-CO_2 feedbacks are likely of secondary importance. Estimated uncertainties in the marine carbon uptake due to climate change, including warming-driven changes in CO_2 solubility, are found to be smaller in magnitude than uncertainties arising from imperfect knowledge of surface-to-deep physical transport (see Figure 2d,e in Friedlingstein et The exchange of CO_2 between the atmosphere and the surface ocean is described by two-way fluxes, from the atmosphere to the surface ocean and vice versa, and the net flux of CO_2 into the ocean is proportional to the air-sea partial pressure difference. CO_2 reacts with water to form carbon and bicarbonate ions (Dickson et al., 2007; Orr et al., 2015), and acid-base equilibria are here described using the well-established Revelle factor formalism (Siegenthaler and Joos, 1992; Zeebe and Wolf-Gladrow, 2001).

The first order climate-carbon feedback of a decreasing solubility in warming water is considered. Surface-to-deep exchange, the rate limiting step of ocean carbon and heat uptake, is described using an IRF. On time scales of up to a few millennia, processes associated with ocean sediments and weathering can be neglected. In such a closed ocean-atmosphere-land biosphere system, excess CO_2 is partitioned between the ocean and the atmosphere and a substantial fraction of the emitted CO_2 remains in the atmosphere and in the surface ocean in a new equilibrium (Joos et al., 2013). This corresponds to a constant term (infinitely long removal time scale) in the IRF representing surface-to-deep mixing. On multi-millennial time scales, excess anthropogenic CO_2 is removed from the ocean-atmosphere-land system by ocean-sediment interactions and changes in the weathering cycle (Archer et al., 1999; Lord et al., 2016), and the IRF is readily adjusted to account for these processes, important for simulations extending over many millennia.

BernSCM simulates global mean surface temperature and the heat uptake by the planet. The latter is equivalent to the net top-of-the-atmosphere energy flux. Changes in the Earth's heat storage in response to anthropogenic forcing are dominated by warming of the surface ocean and the interior ocean (Stocker et al., 2013b) due to their large heat capacity in comparison with that of the atmosphere and their large thermal conductivity in comparison to that of the land surface. Consequently, the atmospheric and land surface heat capacity is formally lumped with the heat capacity of the surface ocean in the BernSCM. The uptake of heat by the ocean (or planet) is, as for carbon, formulated as a two-way exchange flux. The flux of heat from the atmosphere into the surface ocean is taken to be proportional to the radiative forcing resulting from changes in CO_2 and other agents (Etminan et al., 2016). The upward loss of heat from the surface is proportional to the product of the simulated surface temperature perturbation and the (prescribed) climate sensitivity λ (Siegenthaler and Oeschger, 1984a; Winton et al., 2010).

As with carbon, surface-to-deep transport is the rate limiting step for ocean heat uptake and thus for the adjustment of surface temperature to radiative forcing. This transport is key to determine the lag between realized warming and equilibrium warming (Frölicher and Paynter, 2015). Again, this transport is described using an IRF. This IRF encapsulates the finite volume of the entire ocean. It also represents the range of transport time scales associated with advection, diffusion and convection ranging from decades for the ventilation of thermocline to more than a millennium for deep Pacific ventilation as evidenced by transient tracers such as CFCs and radiocarbon (Olsen et al., 2016). The simulated surface ocean temperature perturbation, taken as a measure of global mean surface air temperature change, may be combined with spatial patterns of change in temperature, precipitation or any other variable of interest to compute regionally explicit changes (Hooss et al. (2001b); Joos et al. (2001); Stocker et al. (

Non-CO_2 radiative forcing may be prescribed, e.g., following estimates from complex climate-chemistry models (Myhre et al., 2013) or from simple emission driven non-CO_2 chemistry-radiative forcing modules (Joos et al., 2001; Smith et al., 2017) and reconstructions of solar and volcanic forcing (Eby et al., 2012; Jungclaus et al., 2017) and considering the forcing efficacy of non-CO_2 agents relative to CO_2 forcing (Hansen et al., 2005). Climate sensitivity characterizing the response to radiative forcing, is a free parameter in the BernSCM. Climate sensitivity may change under increasing warming, particularly in high emission scenarios (Geoffroy et al., 2012a; Gregory et al., 2015; Pfister and Stocker, 2017). Here, climate sensitivity is assumed to be time-invariant and a potential state dependency of climate sensitivity is not considered. This may be changed when more solid information on state dependency becomes available or for the purpose of sensitivity analyses. Similarly, ocean heat uptake efficacy (Winton et al., 2010), influencing the atmospheric temperature response to ocean heat uptake forcing, is set to one here.

The present version 1.0 of BernSCM is fundamentally analogous to the Bern Model as used already in the IPCC Second Assessment Report, Bern-SAR (whereas different versions of the Bern model family were used in the more recent IPCC reports). BernSCM represents the relevant processes more completely than Bern-SAR, thanks to additional alternative representations of the land and ocean components, which contain a more complete set of relevant sensitivities to temperature and atmospheric CO_2.

Here, BernSCM model simulations are compared to previous multimodel studies. The model is run for an idealized atmospheric pulse CO_2 emission experiment of Joos et al. (2013)~~,~~ ; for an idealized CO_2 forcing experiment similar to simulations from the Climate Model Intercomparison Project 5 (CMIP5); and for the SRES A2 emission scenario used in the C4MIP study (Friedlingstein et al., 2006a).

Together with this publication, BernSCM v1.0 is provided as an open source Fortran code for free use. The code was also rewritten from scratch, with flexibility and transparency in mind. The model is comprehensively documented, and easily extensible. New alternative model components can be added using the existing ones as a template. A range of numerical solution schemes is implemented. Up to decadal timesteps are supported with high accuracy, suitable for the coupling with, e.g., emission models of coarse time resolution. However, the published code is a ready-to-run standalone model which may also be useful in its own right.

BernSCM offers a physically sound carbon cycle-climate representation, but it is small enough for use in IAMs and other computationally tasking applications. In particular, the support of long time steps is ideally suited to the application of BernSCM ~~a~~ an IAM component, as these complex models often use time steps on the order of 10 years.

BernSCM also offers a tool to realistically assess the climate impact of carbon emissions or emission reductions and sinks, for example in aviation, forestry (Landry et al., 2016), blue carbon ~~managment~~management, peat development (Mathijssen et al., 2017), life cycle assessments (Levasseur et al., 2016), or to assess the interaction of climate engineering interventions such as terrestrial carbon dioxide removal with the natural carbon cycle (Heck et al., 2016).

In this paper, we describe the model equations (section 2 and appendix B), ~~uncertainty assessment (section 3.1),~~ illustrative simulations in comparison with previous multi-model studies and uncertainty assessment (section 3), followed by a discussion (section 4) and ~~conlusions~~ conclusions (section 5).

2 The BernSCM model framework and equations

BernSCM simulates the relation between CO_2 emissions, atmospheric CO_2, radiative forcing (RF), and global mean Surface Air Temperature (SAT) by budgeting carbon and heat fluxes globally between the atmosphere, the (abiotic) ocean, and the land biosphere compartments. Given CO_2 emissions and non-CO_2 RF, the model solves for atmospheric CO_2 and SAT (e.g., in the examples of section 3), but can also solve for carbon emissions (or residual uptake) when atmospheric CO_2 (or SAT and non-CO_2 RF) is prescribed, or for RF when SAT is prescribed.

The transport of carbon and heat to the deep ocean, as well as the decay of land carbon result from complex, but ~~essentially linear behaviour~~ in first order linear behavior of the ocean and land compartments. These are represented in BernSCM using impulse response functions (IRF, or Green's function). The IRF describes the evolution of a system variable after an initial perturbation, e.g., the pulse-like addition of carbon to a reservoir. It fully captures linear dynamics without representing the underlying physical processes (Joos et al., 1996a). More illustratively, the model can be considered to consist of box models, which are an equivalent representation of the IRF model components (Figure 1).

The net primary production (NPP) of the land biosphere and the surface ocean carbon uptake depend on atmospheric CO_2 and surface temperature in a nonlinear way. These essential nonlinearities are described by parametrizations linking the linear model components.

2.1 Carbon cycle component

The budget equation for carbon is

$$\frac{\mathrm{d}m_A}{\mathrm{d}t} = e - f_O - \frac{\mathrm{d}m_L}{\mathrm{d}t}, \tag{1}$$

where m_A denotes the atmospheric carbon stored in CO_2, e denotes CO_2 emissions, f_O the flux to the ocean, m_L the land biosphere carbon stock, and t is time. Here, m_L refers to the (potential) natural biosphere. Human impacts on the land biosphere exchange ~~(LULUC)~~ including land use and land use changes are not simulated in the present version, and treated as exogenous emissions (e). These emissions may be prescribed based on results from spatially-explicit terrestrial models. An overview of the model variables and parameters is given in tables A1 and A2.

The change in land carbon is given by the balance of net primary production (NPP) and decay of assimilated terrestrial carbon,

$$\frac{\mathrm{d}m_L}{\mathrm{d}t} = f_{\mathrm{NPP}} - f_{\mathrm{decay}} \tag{2}$$

Decay includes heterotrophic respiration (RH), fire and other disturbances due to natural processes.

Carbon is taken up by the ocean through the air-sea interface (f_O) and distributed to the mixed surface layer (m_S) and the deep ocean interior (f_{deep}):

$$f_O = \frac{\mathrm{d}m_S}{\mathrm{d}t} + f_{\mathrm{deep}} \tag{3}$$

Global NPP (f_{NPP}) is assumed to be a function of the partial pressure of atmospheric CO_2 (p^{CO_2}) and the SAT deviation from preindustrial equilibrium ~~,~~

~~$$f_{\mathrm{NPP}} = \varphi_{\mathrm{NPP}}(p_A^{CO_2}, \Delta T)$$~~

(functions for the implemented land components are given in Appendix A).

The net flux of carbon into the ocean is proportional to the gas transfer velocity (k_g) and the CO_2 partial pressure difference between surface air and seawater:

$$f_O = k_g A_O \varepsilon (p_A^{CO_2} - p_S^{CO_2}), \tag{4}$$

where Ao A_O is ocean surface area and eps a unit conversion factorε the atmospheric mass of C per mixing ratio of CO_2.

The global average perturbation in surface water $\Delta p_S^{CO_2}$ is a function of dissolved inorganic carbon (change (ΔDIC) in the surface ocean at constant alkalinity (Joos et al., 1996a) , and SAT (Takahashi et al., 1993).

$$\Delta p_S^{CO_2} = \psi(\Delta \mathrm{DIC})\,\chi(\Delta T)$$

; ΔDIC and $p_A^{CO_2}$ are related to model variables (cf. tables A1, A2see Appendix A),

$$\Delta \mathrm{DIC} = \frac{m_S}{H_{\mathrm{mix}} A_O\, \varrho\, M_{\mu\mathrm{mol}}\, 10^{-15} \mathrm{Gt/g}} \tag{5}$$

$$p_A^{CO_2} = m_A \cdot \varepsilon^{-1} \tag{6}$$

The carbon cycle equation set is closed by the specification of f_{decay} and f_{deep} (section 2.3), as well as ΔT, i.e., the coupling to the climate component (section 2.2).

2.2 Climate component

BernSCM simulates the deviation in global mean SAT from the preindustrial state. SAT is approximated by the temperature perturbation of the surface ocean ΔT, which is calculated from heat uptake by the budget equation

$$\frac{\mathrm{d}\Delta T}{\mathrm{d}t} c_S = f_O^H - f_{\mathrm{deep}}^H, \tag{7}$$

where c_s is the heat capacity of the surface layer, f_O^H is ocean heat uptake, and f_{deep}^H is heat uptake by the deep ocean (and accounts for the bulk of the effective heat capacity of the ocean). Continental heat uptake is neglected due to the much higher ratio of heat conductivity to heat capacity heat conductivity of the ocean in comparison to the continent.

f_O^H is taken to be proportional to RF (Forster et al., 2007) and the separation deviation of SAT from radiative equilibrium ($\Delta T = \Delta T^{\mathrm{eq}}(RF)$; see table A2 for parameter definitions),

$$f_O^H = \mathrm{RF}\left(1 - \frac{\Delta T}{\Delta T^{\mathrm{eq}}}\right)\frac{A_O}{a_O} \tag{8}$$

This relation follows from the assumption that feedbacks are linear in ΔT (e.g., Hansen et al., 1984). ΔT^{eq} is given by

$$\Delta T^{\mathrm{eq}} = \mathrm{RF}\frac{\Delta T_{2\times}}{\mathrm{RF}_{2\times}}, \tag{9}$$

where $\Delta T_{2\times}$ is climate sensitivity (defined as the equilibrium temperature change corresponding to twice the preindustrial CO_2 concentration). Climate sensitivy Equation (8) describes ocean heat uptake as the difference between RF and the climate system's response, $\lambda \cdot \Delta\mathrm{T}$, with $\lambda = RF/\Delta T^{\mathrm{eq}}$ the climate sensitivity expressed in W m^{-2} K^{-1}.

Climate sensitivity is an external parameter, as the model does not represent the processes determining equilibrium climate response. RF of CO_2 is calculated as (Myhre et al., 1998)

$$\mathrm{RF}_{CO_2} = \ln\left(\frac{p_A^{CO_2}}{p_{A0}^{CO_2}}\right)\frac{\mathrm{RF}_{2\times}}{\ln(2)}, \tag{10}$$

where $p_A^{CO_2}$ $p_{A0}^{CO_2}$ is the preindustrial reference concentration of atmospheric CO_2, and $RF_{2\times}$ is the RF at twice the preindustrial CO_2 concentration. RF of other GHGs, aerosols etc. can be parametrized in similar expressions involving GHG and pollutant emissions and concentrations (Prather et al., 2001). In the provided BernSCM code, non-CO_2 RF is treated as an exogenous boundary condition. Total RF is then

$$\mathrm{RF} = \mathrm{RF}_{\mathrm{CO_2}} + \mathrm{RF}_{\mathrm{nonCO_2}} \tag{11}$$

The calculation of f_{deep}^H (section 2.3) completes the climate model.

2.3 Impulse response model components

The response of a linear system to a time-dependent forcing f can be expressed by

$$m(t) = \int_{-\infty}^{t} f(t')r(t-t')\,\mathrm{d}t' \tag{12}$$

where equilibrium is assumed for $t \leq t_0$. The function r is the system's impulse response function (IRF), as can be shown by evaluating the integral for a Dirac impulse ($f(t') = \delta(t')$). The IRF indicates the fraction remaining in the system at time t of a pulse input at a previous time t'. Because of linearity of the integral, any physically meaningful integrand f can be represented as a sequence of such impulses of varying size.

In BernSCM, an IRF is used to calculate the perturbation of heat and carbon in the mixed surface ocean layer (mixed layer IRF, (Joos et al., 1996b)(Joos et al., 1996a). For carbon,

$$m_S(t) = \int_{-\infty}^{t} f_O(t')r_O(t-t')\,\mathrm{d}t', \tag{13}$$

and similarly, for heat

$$\Delta T(t)\,c_S = \int_{-\infty}^{t} f_O^H(t')r_O(t-t')\,\mathrm{d}t' \tag{14}$$

where the initial SAT deviation is zero. This approach has been shown to faithfully reproduce atmospheric CO_2 and SAT as simulated with the models from which the IRF is derived (Joos et al., 1996b)(Joos et al., 1996a). For temperature, the linear approach works since relatively small and homogeneous perturbations of ocean temperatures do not affect the circulation strongly and can be treated as a passive tracer (Hansen et al., 2010). Note that for compatibility with commonly used units, carbon fluxes are expressed in Gt per *year*, while heat fluxes are expressed in Joule per second (Watt) in equations 13 and 14, respectively.

Equation (13) closes the ocean C budget equation (3), as can be seen by taking the derivative with respect to time (using $r(0) = 1$),

$$\frac{\mathrm{d}m_S}{\mathrm{d}t} = f_O(t) - \underbrace{\left(- \int\limits_{-\infty}^{t} f_O(t')\,\frac{\mathrm{d}r_O}{\mathrm{d}t}(t-t')\,\mathrm{d}t' \right)}_{f_{\mathrm{deep}}} , \tag{15}$$

where f_{deep} is the flux to the deep ocean. Similarly, equation (14) closes the budget equation for ocean heat uptake heat budget equation (7) . for the surface ocean,

$$\frac{\mathrm{d}\Delta T}{\mathrm{d}t} c_S = f_O^H(t) - \underbrace{\left(- \int\limits_{-\infty}^{t} f_O^H(t')\,\frac{\mathrm{d}r_O}{\mathrm{d}t}(t-t')\,\mathrm{d}t' \right)}_{f_{\mathrm{deep}}^H} \tag{16}$$

Another IRF is used for the carbon m_L in living or dead biomass reservoirs of the terrestrial biosphere,

$$m_L(t) = \pm(t_0) \int\limits_{-\infty}^{t} f_{\mathrm{NPP}}(t') r_L(t-t')\,\mathrm{d}t' \tag{17}$$

Again, equation (17) closes the budget equation for the land biosphere (2), as shown by the derivative with respect to time,

$$\frac{\mathrm{d}m_L}{\mathrm{d}t} = f_{\mathrm{NPP}}(t) - \underbrace{\left(- \int\limits_{-\infty}^{t} f_{\mathrm{NPP}}(t')\,\frac{\mathrm{d}r_L}{\mathrm{d}t}(t-t')\,\mathrm{d}t' \right)}_{f_{\mathrm{decay}}} \tag{18}$$

The time derivative of the land IRF is also known as the decay response function (e.g., Joos et al., 1996a).

The above IRFs can be expressed as a sum of exponentials,

$$r(t) = a_\infty + \sum_k a_k e^{-t/\tau_k} \tag{19}$$

where the constant term a_∞ corresponds to an infinite decay timescale.

The ocean IRF contains a positive constant coefficient a_∞, indicating a fraction of the perturbation that will remain indefinitely (implied by carbon conversation conservation in the ocean model). $CaCO_3$ compensation by sediment dissolution and weathering (Archer et al., 1998) (Archer et al., 1999) are not considered here, but could be described using analogous elimination processes with time scales on the order of 10^4 to 10^5 kyr (Joos et al., 2004). yr (Joos et al., 2004). We emphasize that the implementation considering only the partitioning of excess carbon between atmosphere, land and ocean (hence $a_\infty \neq 0$), neglecting ocean sediment-interactions and weathering flux perturbations, is only valid for time scales shorter than about 2,000 years. In land biosphere models, in contrast, organic carbon is lost to the atmosphere by oxidation to CO_2 at non-zero rates, and consequently all timescales are finite (i.e., $a_\infty = 0$), and the IRF tends to zero (Figure 2).

Presently the parameters of the ocean mixed layer IRF are fixed. A possible change of ocean transport due to global warming is not captured. In contrast, the HRBM land biosphere IRF is temperature-dependent, and captures the enhancement of biomass decay by global warming (s.a. Table ?? and section 3.1).

Inserting formula (19) in the pulse response equation (12) yields (f is a perturbation flux when $a_\infty \neq 0$)

$$m(t) = \sum_k \int_{-\infty}^{t} f(t')\, a_k e^{-(t-t')/\tau_k}\, \mathrm{d}t' + \int_{-\infty}^{t} f(t')\, a_\infty\, \mathrm{d}t' \tag{20}$$

Thus the expression (12) separates into a set of independent integrals m_k corresponding to the number of time scales of the response. Taking the time derivative of expression (20) reveals the equivalence to a diagonal system of linear differential equations,

$$\frac{\mathrm{d}m_k}{\mathrm{d}t} = f(t)\, a_k - m_k/\tau_k; \qquad \frac{\mathrm{d}m_\infty}{\mathrm{d}t} = f(t)\, a_\infty$$
$$m = \sum_k m_k + m_\infty \tag{21}$$

The direct numerical evaluation of the equation (12) involves integrating over all previous times at each timestep. The differential form (21) allows a recursive solution, which is much more efficient, especially for long simulations (the recursive solution implemented in BernSCM is described in appendixAppendix B).

Equation (21) shows the IRF to be equivalent to represents the IRF by a box model, whereby each box m_k receives a fraction a_k of the input f, and has a characteristic turnover time τ_k (Figure 1). For the mixed ocean surface layer the carbon content of box k is given by:

$$\frac{\mathrm{d}m_{S_k}}{\mathrm{d}t} = f_O(t)\, a_{O_k} - m_{S_k}/\tau_{O_k}; \qquad \frac{\mathrm{d}m_{S_\infty}}{\mathrm{d}t} = f_O(t)\, a_{O_\infty} \tag{22}$$

and the change in total carbon content in the mixed layer is:

$$m_S = \sum_k m_{S_k} + m_{S_\infty} \tag{23}$$

Similar equations describe the heat content in the ocean surface layer, as well as the carbon stored in the land biosphere (Figure 1).

The timescales of an IRF describing a linear system are equivalent to the inverse eigenvalues of the model matrix of that system and may also be interpreted in the context of the Laplace transformation (Enting, 2007; Raupach, 2013). For example, the timescales of the mixing layer IRF are the inverse eigenvalues of a matrix describing a diffusive multilayer ocean model (Hooss et al., 2001a). A large model matrix yields a spectrum of many eigenvalues and timescales and corresponding model boxes. In practice, IRFs are approximated with fewer fitting parameters and, equivalently, timescales (4-6 in the case of BernSCM). Joos et al. (1996a) used IRFs combined from two or more functions to minimize the number of parameters needed for an accurate representation. In BernSCM, simple IRFs of the form (19) are used exclusively. This allows adequate accuracy and a consistent interpretation as a multibox model.

Thinking of IRF components as box models is conceptually meaningful. The simple Bern 4 box biosphere model (cf. table ??)(Siegenthal for example, contains boxes corresponding to ground vegetation, wood, detritus, and soil . The HRBM land component(Appendix A). The High-Resolution Biosphere Model (HRBM) land component (Meyer et al., 1999), on the other hand, is abstractly defined by an IRF, but corresponds to boxes which correlate with biospheric reservoirs. However, since different box models may show a similar response, in practice the coefficients a_k and time scales τ_k may not be uniquely defined by the IRF, and should be interpreted primarily as abstract fitting parameters (Enting, 2007)(Enting, 2007; Li et al., 2009).

The timescales of an IRF describing a linearsystem may be thought of as IRF representation is, strictly speaking, only valid if the described subsystem is linear. Then, the response function r does not depend on time and on state variables. In the BernSCM, major nonlinearities in the carbon cycle, namely air-sea gas exchange and the nonlinear carbonate chemistry and changes in NPP in response to changes in environmental conditions are treated by separate nonlinear equations (equations (4) and (5)), while surface-to-deep ocean transport of carbon and heat and respiration of carbon in litter and soils are viewed as approximately linear processes using IRFs. Yet ocean circulation and the respiration of carbon from soil and litter is likely to change under global warming, violating assumption of linearity. In practice, the IRF representation remains a useful approximation as long as the impact of associated nonlinearities on simulated atmospheric CO_2 and temperature remain moderate.

The interpretation of the IRF representation as a box model provides a starting point for considering nonlinearities in the response. To account for nonlinearities, the response time scales τ_k and the coefficients a_k may be gradually adjusted as a function of state variables such as temperature. As the integral form (12) involves integration over the whole history at each time step, changing parameters along the way would result in inconsistencies. In contrast, the differential or box-model form (21) does not depend on previous time steps. Changing the model parameters from one step to the inverse eigenvalues of the model matrix of that system. For example, the timescales of the mixing layer IRF are the inverse eigenvalues of a matrix describing a diffusive multilayer ocean model (Hooss et al., 2001a). A large model matrix yields a spectrum of many eigenvalues and timescales and corresponding model boxes. In practice, IRFs are approximated with fewer fitting parameters and, equivalently, timescales (4-6 in next thus equates to applying a slightly different model at each time step. Within each time step, the parameters remain constant, and the solution for the linear case applies. As time steps are small compared to the whole simulation, this discretization yields accurate results, which is confirmed by the close agreement between the different time resolutions (Table A5).

Varying coefficients have been successfully implemented and tested for the HRBM land component and its decay IRF (Meyer et al., 1999). In this way, the case of BernSCM) . Joos et al. (1996a) used IRFs combined from two or more functions to minimize the number of parameters needed for an accurate representation. In BernSCM, simple IRFs of the form (19) are used exclusively.This allows adequate accuracy and a consistent interpretation as a multibox model enhancement of biomass decay by global warming is captured (s.a. Appendix A and section 3.1). In such a modification, the advantage of the IRF and the equivalent box model representation - the faithful representation of the characteristic response time scale of a model system - is largely maintained, while at the same time the impact of time and state-dependent system responses on simulated outcomes is approximated.

3 ~~Carbon cycle uncertainty assessment~~Illustrative simulations with the BernSCM

3.1 Model setup for sensitivity analyses and uncertainty assessment

The carbon cycle-climate uncertainty of simulations with BernSCM can be assessed in two ways. First, to assess structural uncertainty, different substitute models for the ocean and land components can be used~~(Table ??)~~. Currently, this approach is quite limited by the set of available substitute models (see Appendix A). Second, parameter uncertainty can be assessed by varying the temperature and CO_2 sensitivities of the model, based on a standard set of components that represent the key dependencies as completely as possible (here, the IRF substitutes for the ~~HILDA ocean model, and~~ High-Latitude Exchange/Interior Diffusion-Advection (HILDA) ocean model (Joos et al., 1996a) and for the HRBM land biosphere model ~~are used for~~ (Meyer et al., 1999) are used in the standard setup).

The uncertainties of the global carbon cycle concern the sensitivity of the ~~modelled~~ modeled fluxes of carbon and heat to changing atmospheric CO_2 and climate. Key uncertainties strongly affecting the overall climate response are associated with land C storage: ~~The~~ the dependency of NPP on CO_2 (CO_2 fertilization~~, eq. ??~~), and the dependency of land C on temperature (f_{decay} increases with warming~~, eq. (2)) give~~). This gives rise to large and opposed carbon ~~fluxes~~ flux perturbations which are both very uncertain in magnitude (Le Quéré et al., 2016). While all substitute land models available for BernSCM include CO_2 fertilization, only the HRBM substitute model represents temperature sensitivity of biomass decay (~~IRF parameters are temperature-dependent; Table ??~~Appendix A2).

As for the ocean, the uncertainty of heat uptake into the surface ocean is treated in terms of climate sensitivity (eq. 8). The efficiency of the uptake of heat (f^H_{deep}) and carbon (f_{deep}) into the deep ocean is not sensitive to temperature, as the currently available substitute models all represent a fixed circulation pattern (IRF/box-model parameters are not temperature dependent, Appendix A1). The nonlinear chemistry of CO_2 dissolution in the surface ocean (eq. 4), which determines the sensitivity of ocean C uptake to atmospheric CO_2, is scientifically well established (Dickson et al., 2007; Orr and Epitalon, 2015), and is not treated as an uncertainty in BernSCM. The temperature sensitivities of NPP ~~(eq. ??)~~ and CO_2 dissolution in the surface ocean ~~(eq. ??)~~ are treated as uncertain here, but have secondary influence on the climate response.

Similar to previous studies using models from the Bern family (Plattner et al., 2008; Joos et al., 2001; Meehl et al., 2007; Van Vuuren et al., 2008), the parameter uncertainty range is assessed using the following setups:

"coupled": All temperature and CO_2 sensitivities at their standard values

"uncoupled": All sensitivities zero (except from the ocean CO_2 dissolution chemistry)

~~**"Conly":**~~

"C-only": Only CO_2 dependencies considered (CO_2 fertilization)

~~**"Tonly":**~~

"T-only": Only temperature dependencies considered in land module (NPP, decay~~and ocean C uptake~~)

Climate models with explicit and detailed carbon cycle components exhibit a wide range of responses, as shown in the intercomparison studies of climate models with a detailed carbon cycle, C4MIP (Friedlingstein et al., 2006a) and CMIP5 (Jones et al., 2013). The authors analysed the feedback of carbon cycle-climate models using linearized sensitivity measures. These are derived from a simulation with temperature dependence ("coupled") and one without ("uncoupled"; note that these names have adifferent meaning in BernSCM) . Total We performed simulations with these different setups. In section 4.2, we probe the time scales of the temperature response in simulations where atmospheric CO_2 emissions for the "coupled" (left hand side)and "uncoupled" (right hand side)simulations can be expressed as

$$\Delta C_A^c(\varepsilon + \beta_L + \beta_O + \alpha(\gamma_L + \gamma_O)) = \Delta C_A^u(\varepsilon + \beta_O + \beta_L)$$

where ΔC_A is the cumulative change in atmospheric is abruptly (instantaneously) quadrupled or by increasing CO_2 radiative forcing linearly within 140 years. In section 4.3, we probe the response of the coupled system to a pulse-like release of 100 GtC into the atmosphere. Finally in section 4.4, we analyze carbon cycle-climate feedbacks relying on simulations over the industrial period and for the SRES A2 scenario. BernSCM results are compared with the results from three multi-model intercomparison projects: the Climate Model Intercomparison Project 5 (in ppm)in the coupled (c) or uncoupled (u)case, and the terms in brackets represent the total sensitivity of C storage to ΔC_A; in particular, β is the change in carbon stored (in GtC) on land (L) or in the ocean (O) in responseto atmospheric CMIP5) with results as summarized by Frölicher and Paynter (2015); an analysis of carbon dioxide and climate impulse response functions (Joos et al., 2013, here referred to as IRFMIP), and the C4MIP Climate-Carbon Cycle Feedback Analysis (Friedlingstein et al., 2006a).

3.2 Fraction of realized warming and idealized forcing experiments

The climate response of BernSCM is illustrated using idealized simulations with prescribed forcing. One series of simulations (a) was run for CO_2 change, γ is similarly the change in carbon storage in response to warming, and α is the transient climate sensitivity with respect to atmospheric concentration increasing exponentially from the preindustrial value by 1% per year over 140 years to approximately four times the preindustrial concentration, corresponding to a linear increase in RF (Figure 3, panel a); in a second series of simulations (b), CO_2 concentration; ε converts ppm to GtC (cf. Table A2; the formula in the original paper implies identical units for atmospheric and stored carbon).

The climate-carbon cycle feedback is measured by the feedback parameter g, defined by

$$\frac{\Delta C_A^c}{\Delta C_A^u} = \frac{1}{1-g}$$

and is thus estimated by

$$g = -\frac{\alpha(\gamma_L + \gamma_O)}{\varepsilon + \beta_O + \beta_L}$$

Thus the feedback strength scales with the assumed climate sensitivity and the temperature sensitivities, and is reduced by -induced sinkswas abruptly increased to four times the preindustrial concentration (Figure 3, panel b).

The BernSCMsensitivity setups can be expressed in terms of the C4MIP sensitivityparameters: Tonly corresponds to $\beta_L = 0$, Conly to $\gamma_L = \gamma_O = 0$, and uncoupled to $\beta_L = \gamma_L = \gamma_O = 0$. This can be used to estimate climate-carbon cycle feedback g captured in BernSCM. A comparison of the uncertainty ranges for BernSCM (including structural and parameter uncertainty) and Frölicher and Paynter (2015) compare similar simulations of Earth System Models (ESM) performed within the Coupled Model Intercomparison Project Phase 5 (CMIP5), and Earth System Models of Intermediate Complexity (EMIC) (Joos et al., 2013). As a model comparison metric sensitive to the long-term climate response, Frölicher and Paynter (2015) use the fraction of realized warming, defined by the ratio of the temperature response at a given year and the C4MIP ensemble is shown in section 3.

4 Illustrative simulations with BernSCM

equilibrium temperature for the corresponding RF. They show that the smaller realized warming of ESMs in comparison to EMICs (Figure 3) is connected to a higher long-term warming response; this implies an increase in the coefficient relating global warming to cumulative carbon emissions on multi-centennial timescales and suggests a lower quota on allowed emissions for a given global warming target (Frölicher and Paynter, 2015). The realized warming fraction simulated with BernSCM is in good agreement with the responses of the ESMs (and lower on average than that of the EMICs). The validity of the IRF approach has also been shown by Good et al. (2011) using a SCM to reconstruct and interpret AOGCM projections. For the 150-year time scale of the CMIP5 experiments, Geoffroy et al. (2012b, a) show that the climate response of AOGCMs is well captured by a two-layer energy balance model with two effective response time scales.

In this section, simulations with BernSCM are compared with the results from two multi-model intercomparison projects: an analysis of Carbon dioxide and climate impulse response functions (Joos et al., 2013, here referred to as IRFMIP), and the C4MIP Climate–Carbon Cycle Feedback Analysis (Friedlingstein et al., 2006a)In BernSCM, the fraction of realized warming depends primarily on the choice of climate sensitivity, and is qualitatively similar for the different model setups. Such a clear relationship is not seen in the EMS and EMICS. Thus the structural uncertainty and model differences of complex models is not fully represented in BernSCM. The BernSCM climate response to abrupt warming (Figure 3, panel b) is qualitatively similar, especially on multi-centennial time scales.

3.1 Impulse response experiment

Coupled carbon cycle-climate models can be characterized and compared based on their response to a CO_2 emission pulse to the atmosphere (Joos et al., 2013). In IRFMIP, the The airborne fraction (AF) denotes the fraction of emissions found in the atmosphere at at a given time. In IRFMIP, the AF for a pulse of 100 GtC, emitted on top of current (i.e., year 2010) atmospheric CO_2 concentrations, was simulated by a set of 15 carbon cycle-climate models of different complexity. For three of these models (Bern3D-LPJ, GENIE, MAGICC), ensembles sampling the parameter uncertainty of these models are included in IRFMIP. Thus, IRFMIP captures structural as well as parameter uncertainty.

The IRFMIP pulse experiment was repeated with BernSCM, exploring parameter uncertainty of the carbon cycle (section 3.1), as well as structural uncertainty, using the ocean model IRFs HILDA and Princeton (Sarmiento et al., 1992) in various combinations with the land biosphere components HRBM and Bern-4box (Figure~~.~~ 4). Simulations were run for equilibrium climate sensitivities of 3°C (standard setup), 2 ~~K~~°C, and 4.5 ~~K~~°C.

The AF simulated with BernSCM broadly agrees with the set of simulations from IRFMIP. 100 years after the pulse, it is ~~0.30~~ 0.40 (0.34–0.57) for a climate sensitivity of ~~3K~~ 3°C (for coupled setup with uncertainty range in brackets). Climate sensitivity uncertainty only slightly affects the upper end of this range (Figure 4). For AF simulated with BernSCM, the standard coupled setup is close to the IRFMIP multimodel median~~, but the~~. The BernSCM uncertainty range is asymmetric~~. The~~, like the IRFMIP multi-model range~~is similarly asymmetric~~. For the MAGICC and GENIE ensembles, the medians also correspond with the BernSCM standard case, while the uncertainty ranges are more symmetric~~, which may be related to the method used to sample the parameter uncertainties~~.

The BernSCM SAT response also broadly agrees with IRFMIP. The standard coupled simulation is somewhat lower than the IRFMIP median, which is explained in part by the climate sensitivity (3 ~~K~~°C) being slightly lower than the IRFMIP average (3.2°C). The short term temperature response of BernSCM in particular is on the lower side of the IRFMIP range, suggesting stronger ocean mixing. The quickest initial temperature increase of the BernSCM simulations is obtained with the Princeton ocean model component (dashed lines), which shows a slower initial mixing to the deep ocean than the other implemented components (Figure 2). The comparability of the SAT projections is limited, as the range of climate sensitivities considered in the BernSCM simulations (2-4.5 ~~K~~°C) differ somewhat from that of the IRFMIP multimodel set (1.5-4.6 ~~K~~°C) and the single model ensembles (1.9-5.7 ~~K~~°C), and are compounded with RF differences resulting from the uncertainty in atmospheric CO_2.

Figure 5 shows how the added carbon is redistributed within the Earth system. In the coupled setup, the fraction of the initial pulse sequestered by the land and by the ocean increases over the first century, while the airborne fraction decreases. After 100 years, slightly more than 20% of the added carbon is stored in the land and about 40% in the ocean. The ocean continues to sequester excess carbon in the following centuries to become the dominant sink for excess carbon. In contrast, the land returns part of the sequestered carbon back to the atmosphere and ocean as decreasing atmospheric CO_2 is reducing the modeled CO_2 fertilization of the land biosphere. In the T-only setup, where CO_2 fertilization is not operating, the land is a source of carbon to the atmosphere due to accelerated soil turnover in response to warming. The largest land sink is simulated in the C-only setup, where soil turnover timescales remain invariant and CO_2 fertilization is on. The different BernSCM setups span a range of plausible land biosphere and ocean responses to continued anthropogenic CO_2 emissions as reflected in the simulated range in the airborne fraction (Figure 4a, 5)."

3.2 Carbon cycle-climate feedbacks

Climate models with explicit and detailed carbon cycle components exhibit a wide range of responses, as shown in the intercomparison studies of climate models with a detailed carbon cycle, C4MIP (Friedlingstein et al., 2006a) and CMIP5 (Jones et al., 2013). The authors analyzed the feedback of carbon cycle-climate models using linearized sensitivity measures. These are derived from a simulation with temperature dependence ("coupled") and one without ("uncoupled"; note that these names have a different meaning in BernSCM). Total CO_2 emissions for the "coupled" (left hand side) and "uncoupled" (right hand side) simulations can be expressed as

$$\Delta C_A^c(\varepsilon + \beta_L + \beta_O + \alpha(\gamma_L + \gamma_O)) = \Delta C_A^u(\varepsilon + \beta_L + \beta_O) \tag{24}$$

where ΔC_A is the cumulative change in atmospheric CO_2 (in ppm) in the coupled (c) or uncoupled (u) case, and the terms in parentheses represent the total sensitivity of C storage to ΔC_A; in particular, β is the sensitivity of carbon storage to atmospheric CO_2 (in GtC/ppm) on land (β_L) or in the ocean (β_O). γ is similarly the sensitivity in carbon storage to climate change, and α is the linear transient climate sensitivity to CO_2 (°C per ppm) as in Friedlingstein et al. (2006b); ε converts ppm to GtC (cf. Table A2; the formula in the original paper implies identical units for atmospheric and stored carbon).

The climate-carbon cycle feedback is measured by the feedback metric g, defined by

$$\frac{\Delta C_A^c}{\Delta C_A^u} = \frac{1}{1-g} \tag{25}$$

and is thus estimated by

$$g = -\frac{\alpha(\gamma_L + \gamma_O)}{\varepsilon + \beta_O + \beta_L} \tag{26}$$

Thus the feedback strength scales with the assumed climate sensitivity and the temperature sensitivities, and is reduced by CO_2-induced sinks.

The C4MIP study used a SRES A2 emission scenario to compare the carbon cycle sensitivities of a range of models. As in the C4MIP exercise, BernSCM was run for SRES A2 without any non-CO_2 forcings (Figure 6; prescribed historical and scenario emissions were smoothed with the R smooth.spline function (R Core Team, 2015) for 41 degrees of freedom for use with different time steps). Land use was treated as an exogenous CO_2 emission, while the land model simulates an undisturbed biosphere.

The BernSCM sensitivity setups can be expressed in terms of the C4MIP ~~results can be compared to the BernSCM simulations using the carbon cycle sensitivity parameters defined in section 3.1 (Table 1)~~sensitivity metrics: T-only corresponds to $\beta_L = 0$, C-only to $\gamma_L = \gamma_O = 0$, and uncoupled to $\beta_L = \gamma_L = \gamma_O = 0$. This can be used to estimate climate-carbon cycle feedback g captured in BernSCM. The sensitivity ~~parameters~~ metrics for the BernSCM standard simulation (HILDA-HRBM with coupled carbon cycle) lie within the C4MIP range (Table 1). The uncertainty range for BernSCM, however, is not congruent with the multi-model range of C4MIP. Maximum and standard sensitivity for BernSCM are practically identical. Notably, this sensitivity is smaller (absolutely) than the C4MIP average for the land carbon response to CO_2 increase and warming. The resulting gain g is also smaller, though this results in large part from the lower climate sensitivity in BernSCM (which corresponds to 2.5 ~~K~~ °C as used for the Bern-CC model contribution to C4MIP). The lower end (in absolute terms) of the BernSCM carbon cycle sensitivity range is, on the other hand, zero per definition for all but the ocean-CO_2 sensitivity β_O (see section 3.1). As a consequence, the climate-carbon cycle feedback range also includes zero. In contrast, the C4MIP range does not include zero for all sensitivity parameters.

The land carbon uptake until 2100, under the different BernSCM configurations, varies over 500 GtC (Figure 6), more than three times the range of ocean uptake (180 GtC). This partly reflects the limited coverage of the uncertainty in ocean mixing, but also the fact that the land carbon sink is, together with the land use-related source, the most uncertain item in the budget (Le Quéré et al., 2009).

Together, the uncertainties in the carbon cycle sensitivities amount to a range of about 200 ppm in the projected atmospheric for this scenario around 2100; the SAT range, after emissions have ceased in 2100, reaches roughly 1 K. Thus the carbon cycle uncertainty range amounts to about 1/3 of the total anthropogenic perturbation for both and SAT.

4 Discussion

We simulated illustrative scenarios from two recent multi-model studies, C4MIP and IRFMIP, to compare BernSCM to the literature of carbon-cycle climate models. The results show that BernSCM is broadly representative of the current understanding of the global carbon cycle-climate response to anthropogenic forcing (in a time-averaged sense that does not address internal variability). The BernSCM uncertainty range in CO_2 and SAT projections is broadly similar to the ranges spanned by probabilistic single-model ensembles, and multi-model "ensembles of opportunity" such as the 15 IRFMIP models. The shown BernSCM uncertainty range consists mainly of parameter uncertainty and to a small extent of structural uncertainty. For the standard, coupled model setup, the sensitivities of ocean and land carbon uptake to changing CO_2 and climate (Table 1) of BernSCM are within the range of the detailed carbon cycle models in C4MIP. However, as some C4MIP models show much higher sensitivities, the BernSCM range does not capture the full C4MIP multi-model range. On the other hand, the C4MIP set is unlikely to sample uncertainty exhaustively, as each model contributed only a single, "most likely" simulation. Thus it does not include zero (or weak) sensitivities, whereas the BernSCM range does.

Figure 4 illustrates the importance of a systematic appraisal of uncertainty considering not only the "most likely" model setups, as the standard coupled response in and SAT is near the lower end of the range, and may thus understate the impact. This is even more the case if the key processes are not implemented fully. For example, the early model version Bern-SAR, which was used for the Global Warming Potential (GWP) estimates in the IPCC second assessment report (Schimel et al., 1996) and more recently for integrated assessment (e.g. Hijioka et al., 2006), lacks temperature sensitivity of land carbon uptake (corresponding to the Conly setup) and coincides with the lower end of the BernSCM range.

As Figure 6 shows, solutions with different timesteps and numerical schemes as implemented in BernSCM are largely equivalent for a sufficiently smooth forcing. This offers the flexibility to opt for simplicity of implementation or maximum speed as required by the application (see also Appendix B).

BernSCM does not explicitly distinguish between surface atmosphere and surface ocean temperature to compute global mean surface air temperature perturbation. This is in contrast to some energy balance calculations used to analyze results from state-of-the-art Earth System Models (e.g., Geoffroy et al., 2012b). The BernSCM approach follows earlier work of Siegenthaler and Oeschger (1984a). It is further guided by the similarity in reconstructions of marine night time air and sea surface temperature perturbations (Stocker et al., 2013b) that are consistent with the short, monthly relaxation time scale for air-sea heat exchange. The focus of the BernSCM is on the representation of the transport of heat from the surface into the thermocline and the deep ocean on decadal to multi-century time scales, while information on seasonal and spatial changes such as on land-sea air temperature differences or polar amplification may be obtained by applying suitable spatial perturbation patterns as derived from state-of-the-art models.

Currently, a limited set of substitute models is available and included with BernSCM. The simple structure and open source policy of BernSCM allows users to address these current limitations according to the needs of their applications. More components can be added using the existing ones as a template. This requires the specification of the IRF and the parametrization of gas exchange for the surface ocean, or NPP for the land biosphere, respectively (as described in Joos et al., 1996a; Meyer et al., 1999). For the ocean component, it

Ocean transport is known to vary under climate change with some consequences for heat and carbon uptake (Joos et al., 1999b). Here, we applied time-invariant ocean transport parameters (a_{Q_k}, τ_{Q_k}). It is in principle possible to represent temperature dependency of ocean transport in the same a similar way as it is done for the climate dependency of heterotrophic respiration for the HRBM land biosphere component (Meyer et al., 1999)substitute model (Meyer et al., 1999). In the current BernSCM version, the same IRF parameters are applied for the transport of carbon and heat from the surface ocean to the interior ocean. Thereby, it is implicitly assumed that the spatial pattern of change is the same for temperature and carbon. This appears to be a reasonable first-order approximation on decadal-to-century time scales as perturbations in temperature and carbon show similar patterns with decreasing perturbations from the surface to depth. In future efforts, one may differentiate the ocean IRF for heat and carbon, in particular when more information from long-term multi-century to millennial-scale ESM simulations becomes available. The application of the same IRF for carbon and heat in individual model runs implies that modeled carbon and heat transport tend to be physically consistent. In contrast, some other simple models employ different transport parameters for heat and carbon and varied these parameters independently in probabilistic studies.

A distribution of time scales applies to ocean transport processes as evidenced by observations of transient and time dependent tracers such as chlorofluorocarbons, bomb-produced and natural radiocarbon and biogeochemical tracers (Key et al., 2004; Olsen
This continuum is sometimes approximated by one time scale, also termed heat uptake efficiency (e.g., Gregory et al., 2009) and by two time scales, as in (Geoffroy et al., 2012b). The one-to-two time scale approximations were used to analyze relatively short Earth System Model simulations that do not yet reveal the multi-century response time scales of the deep ocean. We note that the equivalent ocean depth of the simple energy balance model of Geoffroy et al. (2012b) for their AOGCM ensemble is only 1,though this has not been done yet 182 m compared to a mean ocean depth of about 3,800 m. The ocean IRFs used in the BernSCM are derived from long simulations with ocean-only or simplified models. The range of distinct time scales used to construct the IRF faithfully approximates the sub-annual to multi-century response continuum of the parent models as shown in earlier work (Joos et al., 1996a). Further, the BernSCM IRF model represents the heat capacity of the entire ocean.

One The BernSCM model may be extended to model perturbation in the signatures and exchange fluxes of the carbon isotopes ^{13}C and ^{14}C as demonstrated in earlier work (Joos et al., 1996a). This was not implemented here to keep the code as simple as possible and as most potential users are likely concerned with the evolution of climate and atmospheric CO_2.

A potential future application of BernSCM is to use it as an emulator of the global long-term response of complex climate-carbon cycle models by adding the corresponding substitute model components. Additionally, pattern scaling can be applied to transfer the global mean temperature signal into spatially resolved changes in surface temperature, precipitation, cloud cover, etc., exploiting the correlation of global SAT with regional and local changes (Hooss et al., 2001a). This allows to drive spatially explicit models, e.g., of terrestrial vegetation (as in Joos et al., 2001; Strassmann et al., 2008) or climate change-related impacts ~~(as in Hijioka et al., 2009).~~ (e.g., as in Hijioka et al., 2009). Patterns of change are generally similar across models for temperature, whereas patterns in precipitation are more uncertain and show greater variability between models (Knutti and Sedlacek, 2013) and are forcing dependent (Shine et al., 2015). We also note that natural variability strongly influences the space-time evolution of climate change (Deser et al., 2012). Patterns may be scaled with changes in global mean surface air temperature as indicated in Figure 1 or dependencies on radiative forcing may be considered (Shine et al., 2015)

The addition of further alternative model components will extend the structural uncertainty that can be represented with BernSCM. A sufficient coverage of structural uncertainty could allow the interpolation between alternative model components, to represent uncertainty with scalable parameters (and removing the distinction between structural and parameter uncertainty). Such a parametrization of the uncertainty would enhance the possibilities for probabilistic applications of BernSCM, although more sophisticated models are available for observation-constrained probabilistic quantification of climate targets (Holden et al., 2010; Steinacher and Joos, 2016; Steinacher et al., 2013).

5 Conclusions

BernSCM is a reduced-form carbon cycle-climate model that captures the characteristics of the natural carbon cycle and the climate system essential for simulating the global long term response to anthropogenic forcing. Simulated atmospheric CO_2 concentrations and SAT are in good agreement with results from two comprehensive multi-model ensembles. Process detail is minimal, due to the use of IRFs for system compartments that can be described linearly, and nonlinear parametrizations governing the carbon fluxes into these compartments. This framework allows, in particular, to represent the wide range of response time scales of the ocean and land biosphere, and the nonlinear chemistry of CO_2 uptake in the surface ocean - both essential for reliably simulating the global climate response to arbitrary forcing scenarios.

Due to its structural ~~simplicitly~~ simplicity and computational efficiency, BernSCM has many potential applications. In combination with pattern scaling, BernSCM can be used to project spatial fields of impact-relevant variables for applications such as climate change impact assessment, coupling with spatially explicit land biosphere models, etc. With alternative numerical solutions of varying complexity and stability to choose from, applications range from educational to computationally intensive integrated assessment modeling. BernSCM also offers a model-based alternative to GWPs for estimation of the climate impact of emissions and can be used to quantify climate benefits of mitigation options by applying emissions- or concentration-driven simulations.

The generic implementation of linear IRF-components offers a transparent, extensible climate model framework. Current limitations concern the number of available substitute models (limiting the uncertainty range represented), and ~~a fixed ocean~~

transport ocean transport not influenced by climate change. An addition of further alternative model components, and more flexible representation of sensitivities in terms of continuously variable parameters would further increase the models usefulness, for example for probabilistic applications.

Code availability. The source code of the Bern Simple Climate Model is available from the github repository at https://doi.org/10.5281/zenodo.1038117

Appendix A: Implementation of the pulse-response modelModel parameters and parametrizations

A1 DiscretizationOcean

For the solution of the pulse-response equation (12), two discrete approximations are implemented, which both correspond to the differential equation system (21). Currently available ocean components include substitute models for the High-Latitude Exchange/Interior Diffusion-Advection model (HILDA Joos et al., 1996a), Bern2D (Stocker et al., 1992), and the Princeton GCM (Sarmiento et al., 1992). Ocean model parameters of the equations described in the main text are listed in Table A3 for the mixed-layer IRF/box models and in Table A2 for other equations. The IRF/box model parameters given here are recalculated by fitting a sum of 6 exponential functions and one constant to the original response functions as given in Joos and Bruno (1996). The original functions treated the first few years separately; the approximation to a purely exponential form simplifies the equations and has a negligible effect on accuracy. The parametrization of ocean surface CO_2 pressure is the same for all available ocean components and is given below.

First, f can be taken as constant over a sufficiently short timestep $\Delta t = t_i - t_{i-1}$. This approximation yields the system

$$m_n = m_{\infty n} + \sum_k m_{kn}$$

$$m_{kn} = m_{kn-1}\, e^{-\Delta t/\tau_k} + A_k\, f_{n-\frac{1}{2}}\, \Delta t\, e^{-\Delta t/\tau_k}$$

$$m_{\infty n} = m_{\infty n-1} + A_\infty\, f_{n-\frac{1}{2}}\, \Delta t$$

where the subscript n indicates the state at time t_n, and $f_{n-\frac{1}{2}}$ is the value of f at midpoints between t_{n-1} and t_nOcean surface CO_2 pressure perturbations are fitted as a function of the globally averaged unperturbed surface temperature T^* and perturbations in DIC by Joos et al. (1996a) using carbonate chemistry coefficients summarized by Millero (1995):

$$\Delta\, p_S^{CO_2}\Big|_{T^*} = (1.5568 - 1.3993 \cdot 10^{-2}\, T^*)\, \Delta\mathrm{DIC} + (7.4706 - 0.20207\, T^*)\, 10^{-3}\, \Delta\mathrm{DIC}^2 - (1.2748 - 0.12015\, T^*)\, 10^{-5}\, \Delta\mathrm{DIC}^3 + (2.4491 - 0.12639\, T^*)\, 10^{-7}\, \Delta\mathrm{DIC}^4 - (1.5468 - 0.15326\, T^*)\, 10^{-10}\, \Delta\mathrm{DIC}^5$$

The expression holds for unperturbed global average surface water temperature T^* between 17.7 and 18.3°C and for $\Delta p_S^{CO_2}$ between 0 and 1320 ppm.

Second, for longer timesteps, a better approximation is obtained by assuming linear variation of f over each time step. This yields

$$m_{kn} = m_{kn-1}\,e^{-\Delta t/\tau_k} + (A_k\,f_{n-1} + B_k\,(f_n - f_{n-1}))\,\Delta t\,e^{-\Delta t/\tau_k}$$
$$m_{\infty n} = m_{\infty n-1} + (A_\infty\,f_{n-1} + B_\infty\,(f_n - f_{n-1}))\,\Delta t$$

The coefficients in the above equations are given by

$$A_k = a_k\,\frac{\tau_k}{\Delta t}\,(e^{\Delta t/\tau_k} - 1)$$
$$A_\infty = a_\infty$$
$$B_k = a_k\,\frac{\tau_k}{\Delta t}\left(1 - \frac{\tau_k}{\Delta t}(1 - e^{-\Delta t/\tau_k})\right)\,e^{\Delta t/\tau_k}$$
$$B_\infty = \frac{a_\infty}{2}$$

In the following, equations (B2-??) are derived. Ocean surface CO_2 pressure for global surface temperature perturbation ΔT (Takahashi et al., 1993):

$$p_S^{CO_2} = p_S^{CO_2}\Big|_{T^*} \cdot e^{0.0423\Delta T}$$

We substitute t' by $t - x$ in equation (12) to get

$$m(t) = \int_t^{t_0} f(t-x)r(x)\,\mathrm{d}x$$

A2 Land biosphere

Currently available land biosphere components include substitute models for the High-Resolution Biosphere Model (Meyer et al., 1999) and the 4Box biosphere model (Siegenthaler and Joos, 1992).

Taking f to be constant over one time step,

$$m(t_n) \simeq \sum_{i=1}^{n} f(t_n - t_{i-\frac{1}{2}}) \int_{t_{i-1}}^{t_i} r(x)\,\mathrm{d}x,$$

where the midpoint value $t_{i-\frac{1}{2}}$ is used for accuracy. The integral in (??)can be evaluated explicitly, to define an adapted, discrete IRF R_iFor the HRBM model, temperature-dependent IRF/box model parameters as given by Meyer et al. (1999) are implemented:

$$\tilde{a}_k = \frac{a_k\,e^{s_{a_k}T}}{\sum_j a_j\,e^{s_{a_j}T}},$$
$$\tilde{\tau}_k = \tau_k\,e^{-s_{\tau_k}T},$$

where $\tilde{a}_k$, $\tilde{\tau}_k$ are the adjusted and a_k,

$$R_i = 1/\Delta t \int_{t_{i-1}}^{t_i} r(x)\,\mathrm{d}x$$

where Δt is the length of the (constant)time step. Evaluating the integral using the τ_k the unperturbed parameters. The IRF/box model parameter values for HRBM and the 4box model are listed in Table A4.

Net primary production for HRBM is given by (Meyer et al., 1999):

$$\begin{aligned}\mathrm{NPP}(p)|_{\Delta T=0} = & -e^{3.672801} + e^{-0.430818}\cdot p - e^{-6.145559}\cdot p^2 + e^{-12.353878}\cdot p^3 - e^{-19.010800}\cdot p^4 + e^{-26.183752}\cdot p^5 \\ & - e^{-34.317488}\cdot p^6 - e^{-41.553715}\cdot p^7 + e^{-48.265138}\cdot p^8 - e^{-56.056095}\cdot p^9 + e^{-64.818185}\cdot p^{10}\end{aligned}$$

where p is atmospheric CO_2 pressure. The model includes growth enhancement by SAT increase (but without a dynamical vegetation):

$$\mathrm{NPP}(p,\Delta T) = \mathrm{NPP}_0\cdot(1 + 0.11780208\tanh(\Delta T/50.9312421) + 0.002430513\cdot\tanh(\Delta T/8.85326739))$$

Net primary production for the 4Box model is described after (Enting et al., 1994b; Schimel et al., 1996):

$$\mathrm{NPP} = \mathrm{NPP}_0 + \mathrm{NPP}_0 * \beta * \log(p^{CO_2}/p_0^{CO_2})$$

where NPP_0 is undisturbed NPP.

Appendix B: Implementation of the pulse-response model

B1 Discretization

For the solution of the pulse-response function (19) yields

$$R_i = A_\infty + \sum_k A_k e^{-t_i/\tau_k}$$

This allows to write equation (??)as

$$m_n = m_{\infty n} + \sum_k m_{kn}$$

$$m_{\infty n} = \sum_{i=1}^{n} \Delta t\, f_{n-i+\frac{1}{2}}\, A_\infty$$

$$m_{kn} = \sum_{i=1}^{n} \Delta t\, f_{n-i+\frac{1}{2}}\, A_k e^{-t_i/\tau_k}$$

To derive a recursive expression equation (12), two discrete approximations are implemented, using the separation by time scales in equation (20) or, equivalently, in the differential equation system (21). The recursive solution for a time step Δt can be obtained from equation (??) , split sums,

$$m_{\infty n} = \Delta t\left(f_{n-\frac{1}{2}}\,A_{\infty} + \sum_{i=2}^{n} f_{n-i+\frac{1}{2}}\,A_{\infty}\right)$$
$$m_{kn} = \Delta t\left(f_{n-\frac{1}{2}}\,A_{k}e^{-t_1/\tau_k} + \sum_{i=2}^{n} f_{n-i+\frac{1}{2}}\,A_{k}e^{-t_i/\tau_k}\right)$$

and replace indices $i = j+1$, setting $t_0 = 0$, $t_1 = \Delta t$, $t_{j+1} = t_j + \Delta t$

$$m_{\infty n} = \Delta t\left(f_{n-\frac{1}{2}}\,A_{\infty} + \sum_{j=1}^{n-1} f_{n-1-j+\frac{1}{2}}\,A_{\infty}\right)$$
$$m_{kn} = \Delta t\left(f_{n-\frac{1}{2}}\,A_{k}e^{-\Delta t/\tau_k} + \sum_{j=1}^{n-1} f_{n-1-j+\frac{1}{2}}\,A_{k}e^{-(t_j+\Delta t)/\tau_k}\right)$$

comparison with equation (??) yields the recursive differential system (B220) by substituting $t = t_n = t_{n-1} + \Delta t$, and $s = t' - t_{n-1}$,

$$m_n = m_{\infty n} + \sum_k m_{kn}$$
$$m_{kn} = m_{kn-1}\,e^{-\Delta t/\tau_k} + \int_0^{\Delta t} f(t_{n-1}+s)\,a_k e^{-(\Delta t - s)/\tau_k}\,\mathrm{d}s$$
$$m_{\infty n} = m_{\infty n-1} + \int_0^{\Delta t} f(t_{n-1}+s)\,a_{\infty}\,\mathrm{d}s \tag{B1}$$

where $m_n = m(t_n) = m(t_{n-1}+\Delta t)$.

First, f can be taken as constant over a sufficiently short timestep $\Delta t = t_i - t_{i-1}$. Evaluating equations (B1) yields

$$m_{kn} = m_{kn-1}\,e^{-\Delta t/\tau_k} + f(t^*)\,a_k\tau_k(1 - e^{-\Delta t/\tau_k})$$
$$m_{\infty n} = m_{\infty n-1} + f(t^*)\,a_{\infty}\Delta t \tag{B2}$$

where t^* is chosen to be t_{n-1} (for explicit forward solution) or t_n (for implicit backward solution).

Assuming now Second, for longer timesteps, a better approximation is obtained by assuming linear variation of f over each time stepin equation (12),

$$m(t_n) \simeq \sum_{i=1}^{n}\int_{t_{i-1}}^{t_i}\left(f_{n-i} + (f_{n-i+1} - f_{n-i})\frac{t_i - x}{\Delta t}\right) r(x)\,\mathrm{d}x,$$

$$= \sum_{i=1}^{n} f_{n-i} \int_{t_{i-1}}^{t_i} r(x)\,\mathrm{d}x + (f_{n-i+1} - f_{n-i}) \int_{t_{i-1}}^{t_i} \left(\frac{t_i - x}{\Delta t}\right) r(x)\,\mathrm{d}x$$

Substituting equation (??) for the first integral and using partial integration on the second, one obtains

$$m(t_n) \simeq \sum_{i=1}^{n} \left(f_{n-i} \Delta t\, R_i + \frac{f_{n-i+1} - f_{n-i}}{\Delta t} \left(\cdot \int_{t_{i-1}}^{t_i} \int_{t_{i-1}}^{x} r(y)\,\mathrm{d}y\,\mathrm{d}x \right) \right)$$

The double integral of the pulse-response function evaluates to

$$\int_{t_{i-1}}^{t_i} \int_{t_{i-1}}^{x} r(y)\,\mathrm{d}y\,\mathrm{d}x \quad = \sum_k B_k\, e^{-t_i/\tau_k}\, \Delta t^2$$

By a similar procedure as for the constant flux approximation, the recursive formulation (B3) is obtained. . This yields

$$m_{kn} = m_{kn-1}\, e^{-\Delta t/\tau_k} + f_{n-1}\, a_k \tau_k \left(\frac{\tau_k}{\Delta t}(1 - e^{-\Delta t/\tau_k}) - e^{-\Delta t/\tau_k} \right) + f_n a_k \tau_k \left(1 - \frac{\tau_k}{\Delta t}(1 - e^{-\Delta t/\tau_k}) \right)$$

$$m_{\infty n} = m_{\infty n-1} + \frac{f_{n-1} + f_n}{2}\, a_\infty \Delta t \tag{B3}$$

B2 Numerical schemes

For the solution of the BernSCM model equations, both explicit and implicit time stepping is implemented.

The stability requirement for the numerical solution depends on the equilibration time for the ocean surface CO_2 pressure $p_S^{CO_2}$. Due to the buffering of the carbonate chemistry, the CO_2 equilibration time is smaller than the gas diffusion time scale (~ 10yr) by a ratio given by the buffer factor. For undisturbed conditions (buffer factor $\simeq 10$) the equilibration time is about 1 yr. With increasing DIC, the buffer factor increases and the equilibration time shortens, making the equation system stiffer. Accordingly, when the model is solved explicitly with a time step of 1 yr, instability typically occurs after sustained carbon uptake by the ocean, which can occur in many realistic scenarios.

For the tested scenario range, the explicite explicit solution is stable at a time step on the order of 0.1 yr, for which the piecewise constant approximation is accurate. For larger step size, an implicit solution is required to guarantee stability.

The piecewise constant approximation is adequate for time steps up to 1 yr, and the piecewise linear approximation for up to decadal time steps. An overview of the performance of three representative settings (set at compile time) for the C4MIP A2 scenario is given in Table A5.

The explicit solution is only implemented for the piecewise constant approximation (B2) and is obtained by approximating $f_{n-\frac{1}{2}}$ with f_{n-1}.

For the implicit solution , for both the piecewise constant (B2) or and the piecewise linear approximation (B3), respectively, is solved for the quantities at t_n, approximating $f_{n-\frac{1}{2}}$ by f_n where applicable. Equations (B2,B3) are expressed in a common equation by substituting

$$m_{k\,n} = m_{k\,n-1}\, p_{mk} + f_n\, p_{fk} + f_{n-1}\, p_{fk}^{\mathrm{old}} \tag{B4}$$

~~with the following parameters for the piecewise constant approximation (B2),~~

$$p_{mk} = e^{-\Delta t/\tau_k}$$
$$p_{fk} = \Delta t A_k e^{-\Delta t/\tau_k}$$
$$p_{fk}^{\mathrm{old}} = 0$$

~~and for the piecewise linear approximation (B3),~~

$$p_{mk} = e^{-\Delta t/\tau_k}$$
$$p_{fk} = \Delta t B_k e^{-\Delta t/\tau_k}$$
$$p_{fk}^{\mathrm{old}} = \Delta t (A_k - B_k) e^{-\Delta t/\tau_k}$$

In the following, the implicit solution for the piecewise constant discretization is derived. Here, the fully implicit scheme for land and ocean exchange is discussed, but for stability, it is only crucial to treat ocean uptake implicitly. The parameters of equation (B4) for this case are

$$p_{mk} = e^{-\Delta t/\tau_k}$$
$$p_{fk} = a_k \tau_k (1 - e^{-\Delta t/\tau_k})$$
$$p_{fk}^{\mathrm{old}} = 0 \tag{B5}$$

Consider first the equation system for carbon, assuming temperature to be known (or neglecting temperature dependence of model coefficients). Equation (B4) is applied to land carbon exchange for the constant approximation (B5),

$$m_{L\,n} = m_L^{c\Delta} + \Delta f_{\mathrm{NPP}} \sum_k p_{fk\,L}$$
$$m_L^{c\Delta} = \sum_k m_{Lk\,n-1} p_{mk\,L} + f_{\mathrm{NPP}\,n-1} \sum_k p_{fk\,L} \tag{B6}$$

where $m_L^{c\Delta}$ is the land carbon stock obtained after one time step if NPP remained constant (“constant flux commitment”), and $\Delta f_{\mathrm{NPP}} = (f_{\mathrm{NPP}\,n} - f_{\mathrm{NPP}\,n-1})$ is the change in NPP over one time step.

For ocean carbon uptake,

$$m_{S\,n} = m_S^{c0} + f_{O\,n} \sum_k p_{fk\,O}$$
$$m_S^{c0} = \sum_k m_{Sk\,n-1}\, p_{mk\,O} \tag{B7}$$

where m_S^{c0} is the value of m_S after one time step if $f_{O\,n} = 0$ (“zero-flux commitment”).

To solve the implicit system, the nonlinear parametrizations need to be linearized around t_{n-1}. Linearizing ocean surface CO_2 pressure ~~(??)~~ as a function of surface ocean carbon and inserting in equation (4) yields

$$f_{O\,n} \simeq k_g A_O (m_{A\,n} - \varepsilon\, p^{CO_2}_{S,n-1}) + k_g A_O\, \varepsilon \frac{\mathrm{d}p_S^{CO_2}}{\mathrm{d}m_S}\bigg|_{n-1} (m_{S\,n-1} - m_{S\,n}) \tag{B8}$$

where equations (5,6) were used. Similarly, NPP ~~(??)~~ as a function of atmospheric carbon is linearized,

$$\Delta\, f_{\mathrm{NPP}\,n} \simeq \frac{\mathrm{d}f_{\mathrm{NPP}}}{\mathrm{d}m_A}\bigg|_{n-1} (m_{A\,n} - m_{A\,n-1}) \tag{B9}$$

using equation (6).

The system is completed with the discretized budget equation (1)

$$m_{A\,n} = m_{A\,n-1} + (e_{n-\frac{1}{2}} - f_{O\,n})\Delta t - (m_{L\,n} - m_{L\,n-1}) \tag{B10}$$

Here, $e_{n-\frac{1}{2}}$ is assumed to be known (though this only applies to the "forward" solution for atmospheric CO_2 from emissions, solving for emissions from CO_2 is also implemented in the model code).

After calculating the "~~commited~~committed" values $m_L^{c\Delta}{}_n$, $m_S^{c0}{}_n$ from the model state at t_{n-1}, equations (B7) through (B10) are solved

$$\begin{aligned}\Delta f_{\mathrm{NPP}} = \frac{\frac{\mathrm{d}f_{\mathrm{NPP}}}{\mathrm{d}m_A}\Big|_{n-1}}{UV+W}\bigg(& m_{L\,n-1} - m_L^{c\Delta} + \Delta t\, e_{n-\frac{1}{2}} + \Delta t\, k_g A_O \Big(\varepsilon\, p^{CO_2}_{S,n-1} - m_{A\,n-1} \\ & + \varepsilon \frac{\mathrm{d}p_S^{CO_2}}{\mathrm{d}m_S}\bigg|_{n-1} \Big[m_S^{c0} - m_{S\,n-1} + \sum_k p_{fk\,O} \big(\frac{m_{L\,n-1} - m_L^{c\Delta}}{\Delta t} + e_{n-\frac{1}{2}} \big) \Big] \Big) \bigg)\end{aligned} \tag{B11}$$

with the auxiliary variables

$$U = k_g A_O\, \varepsilon \frac{\mathrm{d}p_S^{CO_2}}{\mathrm{d}m_S}\bigg|_{n-1} \sum_k p_{fk\,O} + 1 \tag{B12}$$

$$V = \frac{\mathrm{d}f_{\mathrm{NPP}}}{\mathrm{d}m_A}\bigg|_{n-1} \sum_k p_{fk\,L} + 1 \tag{B13}$$

$$W = \Delta t\, k_g A_O \tag{B14}$$

and, after inserting into equation (B6),

$$f_{O\,n} = \frac{k_g A_O}{U+W}\Big(m_{A\,n-1} - \varepsilon\, p^{CO_2}_{S,n-1} - \varepsilon \frac{\mathrm{d}p_S^{CO_2}}{\mathrm{d}m_S}\bigg|_{n-1} (m_S^{c0} - m_{S\,n-1}) - (m_{L\,n} - m_{L\,n-1}) + \Delta t\, e_{n-\frac{1}{2}} \Big) \tag{B15}$$

The remaining variables are then calculated using equations (B7) and (B10), whereby first the components $m_{k\,n}$ are calculated as in equation (B4) and then summed. Finally, the ~~non-linear parametrisations (4,??)~~ nonlinear parametrizations are recalculated with the updated model state.

The order of these equations matters, as the updated variables are successively inserted into the following equations. The land part is solved first, and can be substituted by an ~~explicite~~ explicit step or a separate model, while keeping the ocean step implicit.

An implicit time step is also implemented for calculating SAT from RF (again, solving RF from SAT is also implemented but not discussed here). $\mathrm{RF}(t_n)$ can be assumed as known, as atmospheric $\mathrm{CO_2}$ is calculated first (i.e., no linearization necessary). Applying equation (B4) to temperature,

$$\Delta T_n c_S = \Delta T^{c\Delta} c_S + \Delta f_O^H \sum_k p_{fk_O}$$

$$\Delta T^{c\Delta} = \sum_k \Delta T_{k\,n-1} p_{mk_O} + f^H_{O\,n-1}/c_S \sum_k p_{fk_O} \tag{B16}$$

where $\Delta T^{c\Delta}$ is the "~~commited~~ committed temperature" for constant heat flux to the ocean, and $\Delta f_O^H = f^H_{O\,n} - f^H_{O\,n-1}$ is the change in heat flux over one time step. Equations (~~9,8,~~ 8, 9, B16) are solved for f_O^H,

$$f^H_{O\,n} = \frac{\mathrm{RF}_n - \frac{\mathrm{RF}_{2\times}}{\Delta T_{2\times}} \Delta T^{c\Delta} + f_{H n-1} \sum_k p_{fk_O} \frac{\mathrm{RF}_{2\times}}{\Delta T_{2\times} c_S}}{\frac{\mathrm{RF}_{2\times}}{\Delta T_{2\times} c_S} \sum_k p_{fk_O} + a_O/A_O} \tag{B17}$$

Temperature change ΔT_n then follows from equation (B16).

The case of piecewise linear approximation (~~??~~B3) differs from the piecewise constant one (~~B5~~ (B2)) only in a non-zero contribution of f_{n-1} and a slightly different budget equation,

$$m_{A\,n} = m_{A\,n-1} + \left(e_{n-\frac{1}{2}} - \frac{f_{O\,n} + f_{O\,n}}{2} \frac{f_{O\,n} + f_{O\,n-1}}{2} \right) \Delta t - (m_{L\,n} - m_{L\,n-1}) \tag{B18}$$

The first difference merely changes the calculation of "committed" changes, and only the second difference affects the solution of the implicit time step. In practice, however, this can be neglected without loss of accuracy, and thus equations (B11 – B15) and (B17) are also used to solve the piecewise linear system (while equation (B18) is used to close the budget).

B3 Temperature dependent parameters

~~Temperature change in general affects the behavior of the ocean and land biosphere compartments, which are represented by IRFs. Thus, IRF coefficients can be temperature dependent, as it is the case with the HRBM substitute land biosphere model. In the above derivations, the change of temperature over one time step was not considered~~BernSCM allows for temperature-dependent model parameters for IRF-based substitute models. This generalization of the IRF-approach is possible using a box-model form (section 2.3). Currently, temperature-dependent coefficients and time scales are implemented for the HRBM land biosphere substitute model (Appendix A2).

BernSCM updates any temperature-dependent model parameters by approximating the current temperature ΔT_n by the "committed" temperature $\Delta T^{c\Delta}$ as defined in equation (B16). Accuracy is further improved by substituting $\Delta T^{c\Delta}$ for ΔT_n in evaluating equation (B8) with temperature dependent ~~parametrisations (??)~~parametrizations.

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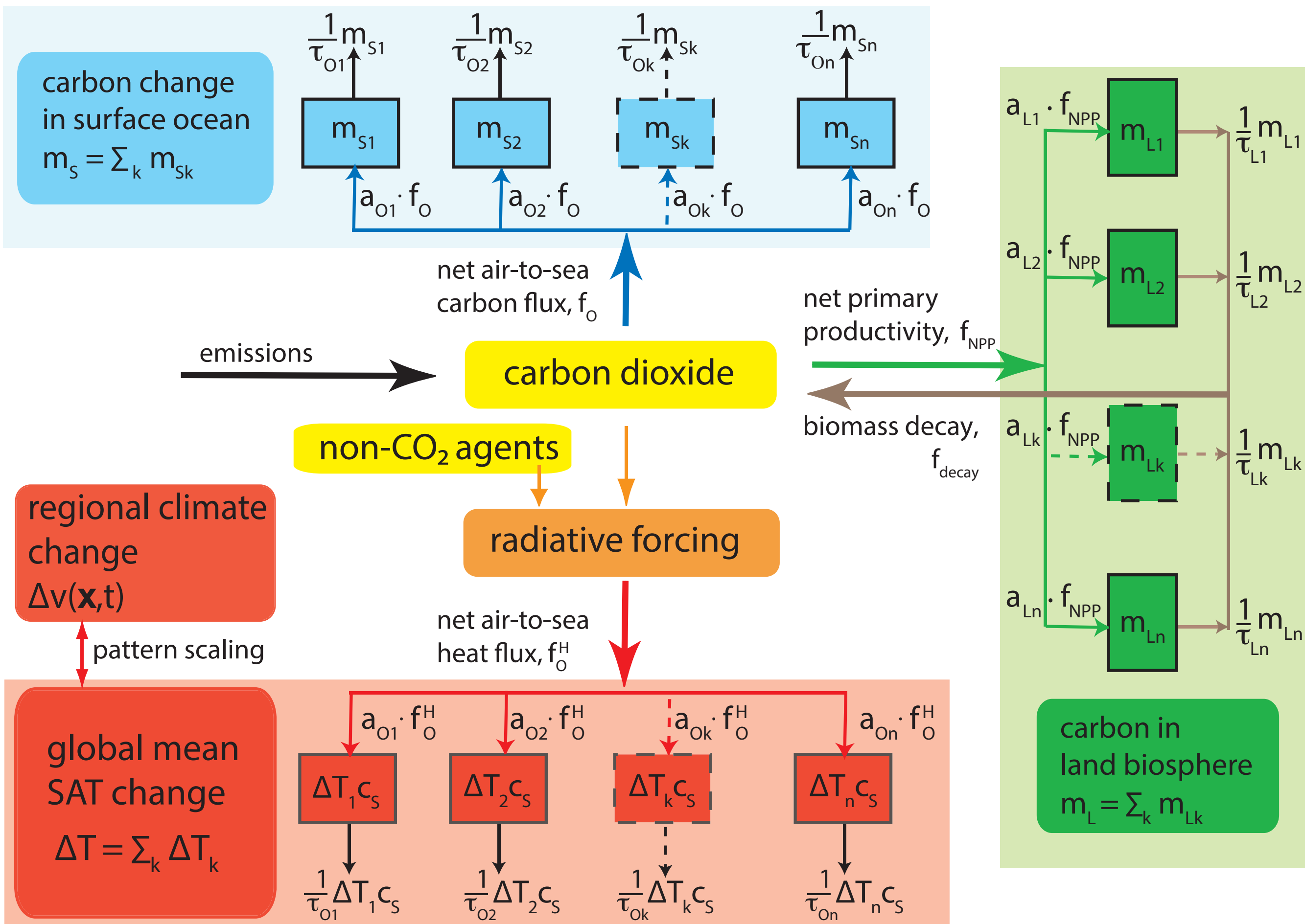

Figure 1. BernSCM as a box-type model of the carbon cycle-climate system based on impulse response functions. Heat and carbon taken up by the mixed ocean surface layer and the land biosphere, respectively, is allocated to a series of boxes with characteristic time scales for surface-to-deep ocean transport (~~τ~~τ_O) and of terrestrial carbon overturning (τ_L). The total perturbations in land and surface ocean carbon inventory and in surface temperature are the sums over the corresponding individual perturbations in each box, ($m_{Sk}, \Delta T_k, m_{Lk}$). Using pattern scaling, the response in SAT can be translated to regional climate change for fields $v(\mathbf{x}, t)$ of variables such as SAT or precipitation.

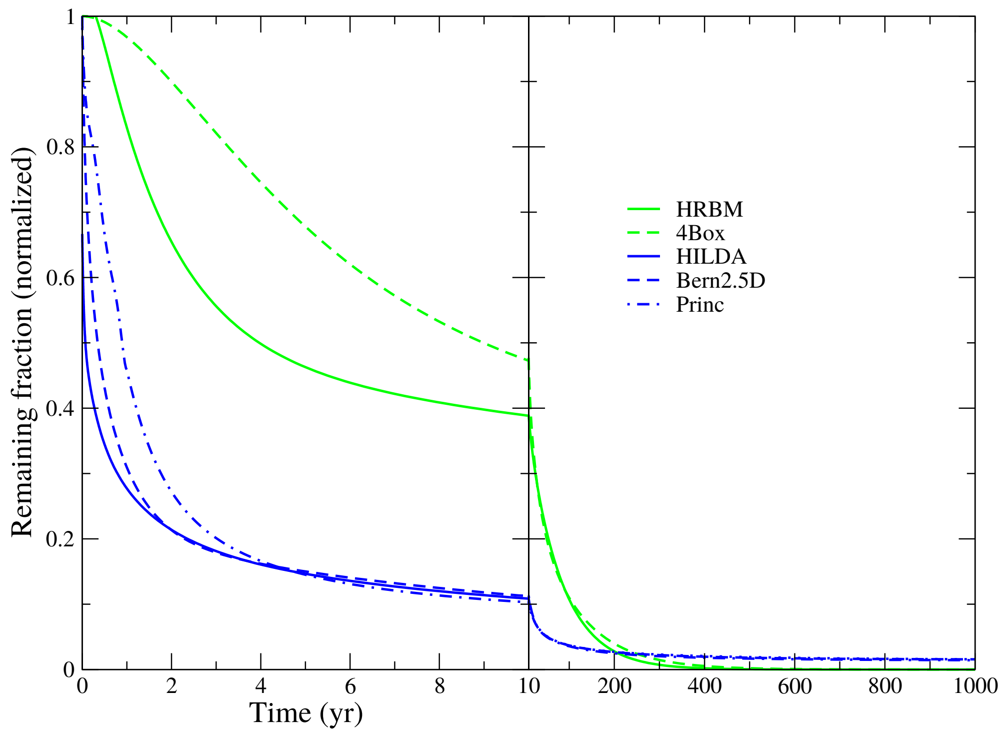

Figure 2. IRFs of ocean (blue) and land (green) model components (without temperature dependence). Ocean components are normalized to a common mixed layer depth of 50m (multiplied by H_{mix}/50m), causing initial response to deviate from 1.

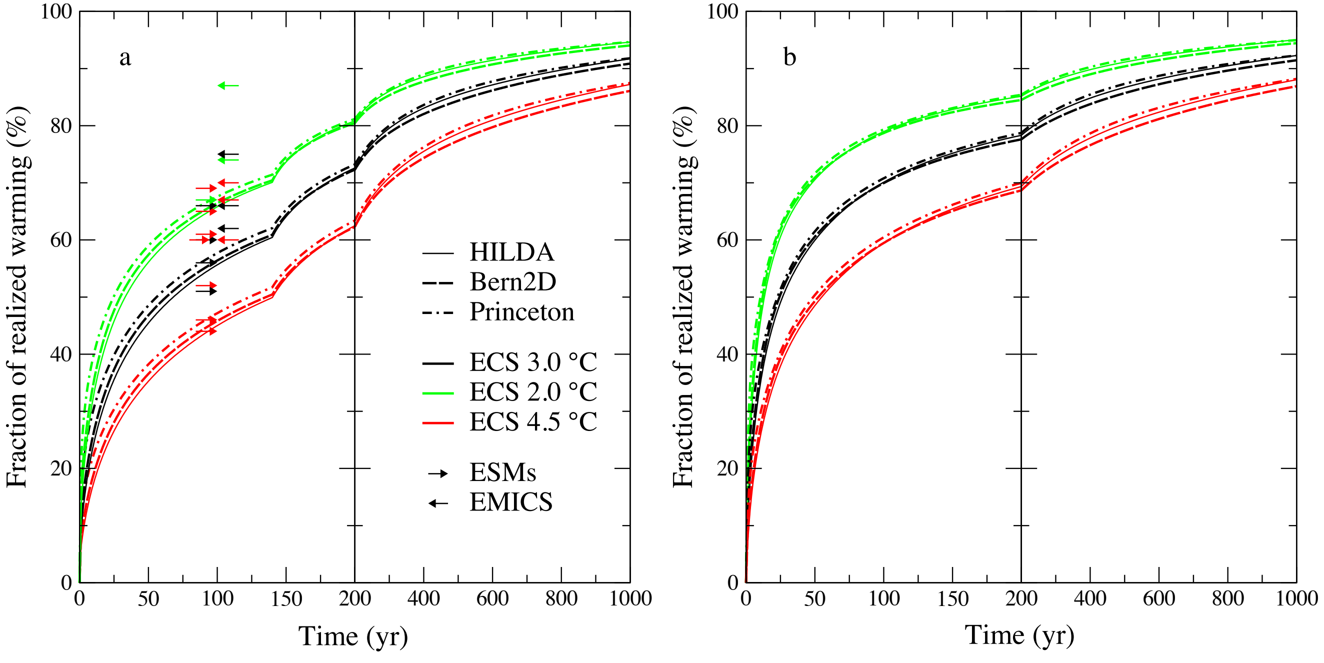

Figure 3. Fraction of realized warming (temperature divided by the equilibrium temperature for the current RF) for idealized experiments with prescribed atmospheric CO_2 concentration increase from preindustrial; panel a shows an exponential CO_2 increase by 1% per year over 140 years to approximately four times the preindustrial concentration (and linear increase in RF); panel b shows an abrupt increase to fourfold CO_2 concentration. BernSCM simulations are shown for climate sensitivities of 2, 3, and 4.5 K and the three available ocean model substitutes as indicated in the legend. Arrows in panel a indicate the corresponding warming fractions at year 99 compiled by (Frölicher and Paynter, 2015, SI Tables 1,2) for Earth System Models (ESM, right-pointing) and Earth System Models of Intermediate Complexity (EMICS, left-pointing); arrow colors indicate climate sensitivities below 2.5 K (green), between 2.5 and 3.5 K (black), and above 3.5 K (red).

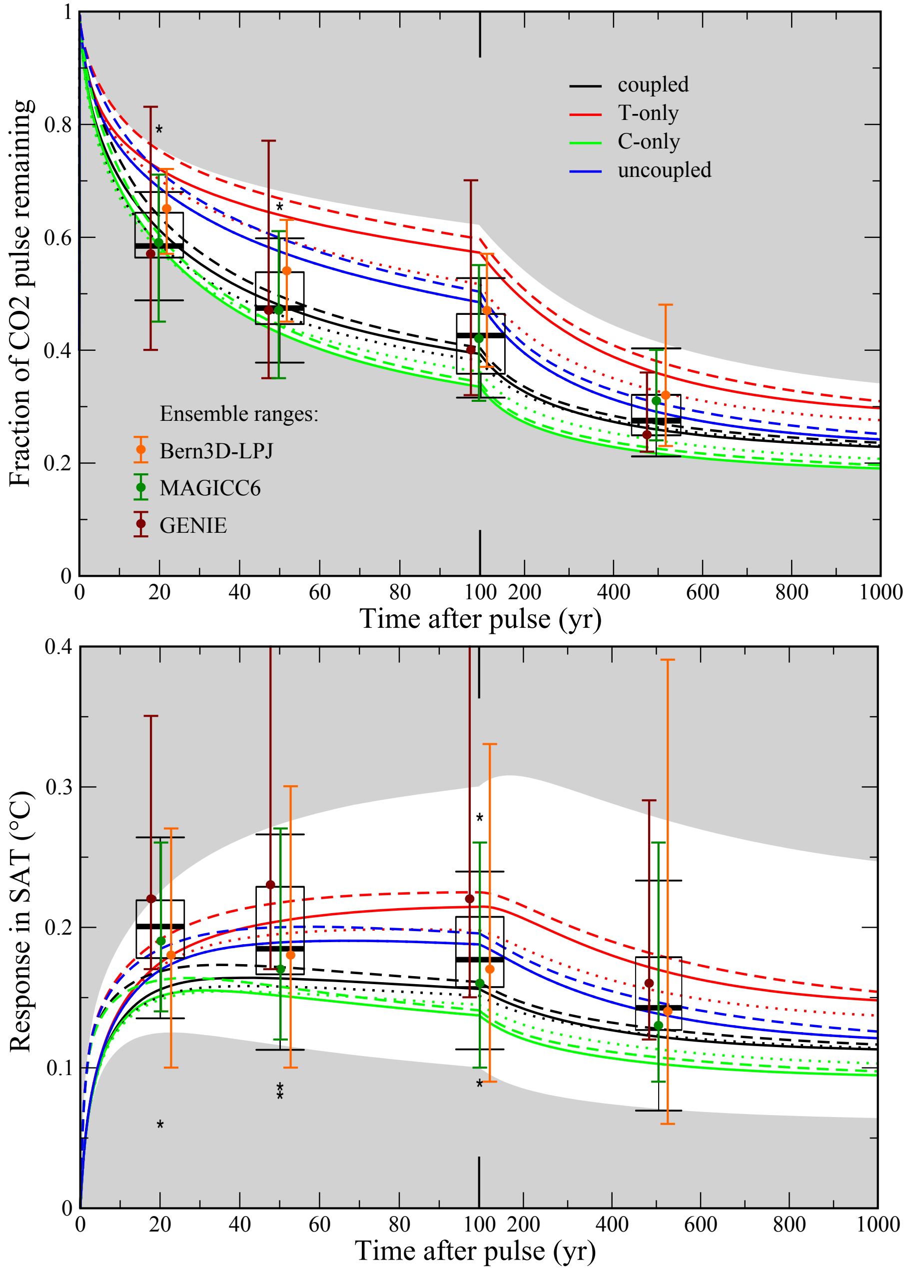

Figure 4. IRFMIP pulse response range compared to BernSCM range for parameter uncertainty (colors according to legend) and structural uncertainty, with model versions HILDA/HRBM (solid lines), HILDA/4box (dots), Princeton/HRBM (dashed). Standard climate sensitivity is 3 ~~K~~°C, and a climate sensitivity range of 2–4.5 ~~K~~°C is shown by the white area (envelope of all BernSCM runs). Single-model ensemble ranges from IRFMIP are included as errorbars indicating the 5-95% range and dots indicating the median. The multimodel IRFMIP range is shown by boxplots indicating median (bold black line), first quartiles (box), extreme values (whiskers) excluding outliers deviating from the median by more than 1.2 times the interquartile distance (asterisks).

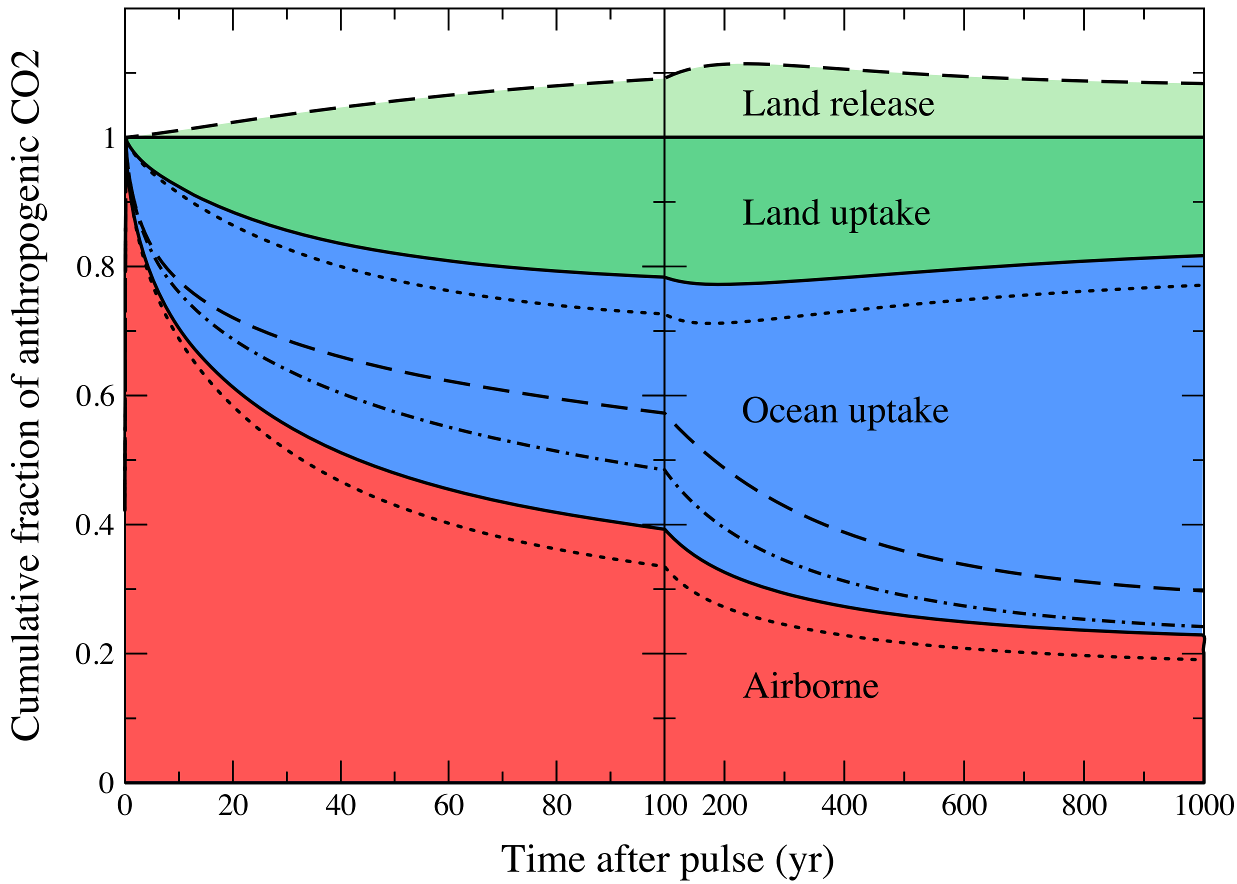

Figure 5. BernSCM simulations of the SRES A2 scenario used for C4MIPLand, ocean, with a climate sensitivity of 2.5C and airborne fractions of the HILDA/HRBM ocean/land components. Results for three numerical schemes are overlaid 100 GtC CO_2 pulse shown in Figure 4 for the same line style; i. 0.1 yr Euler forward timestep coupled (solid lines and colored areas), ii. 1 yr implicit timestep the T-only (dashed), iiithe C-only (dotted) and the uncoupled (dash-dotted) model setup. 10 yr implicit timestep with piecewise linear approximation of fluxes In the T-only case, the land biosphere exhibits a net release (dash-dotlight green shading);, and the difference at ocean uptake consists of the sum of this resolution area and the area delimited by the dashed line below the line at 1; for the uncoupled case, land uptake is only visible in C zero and ocean uptake extends from the dash-dotted line to unity.

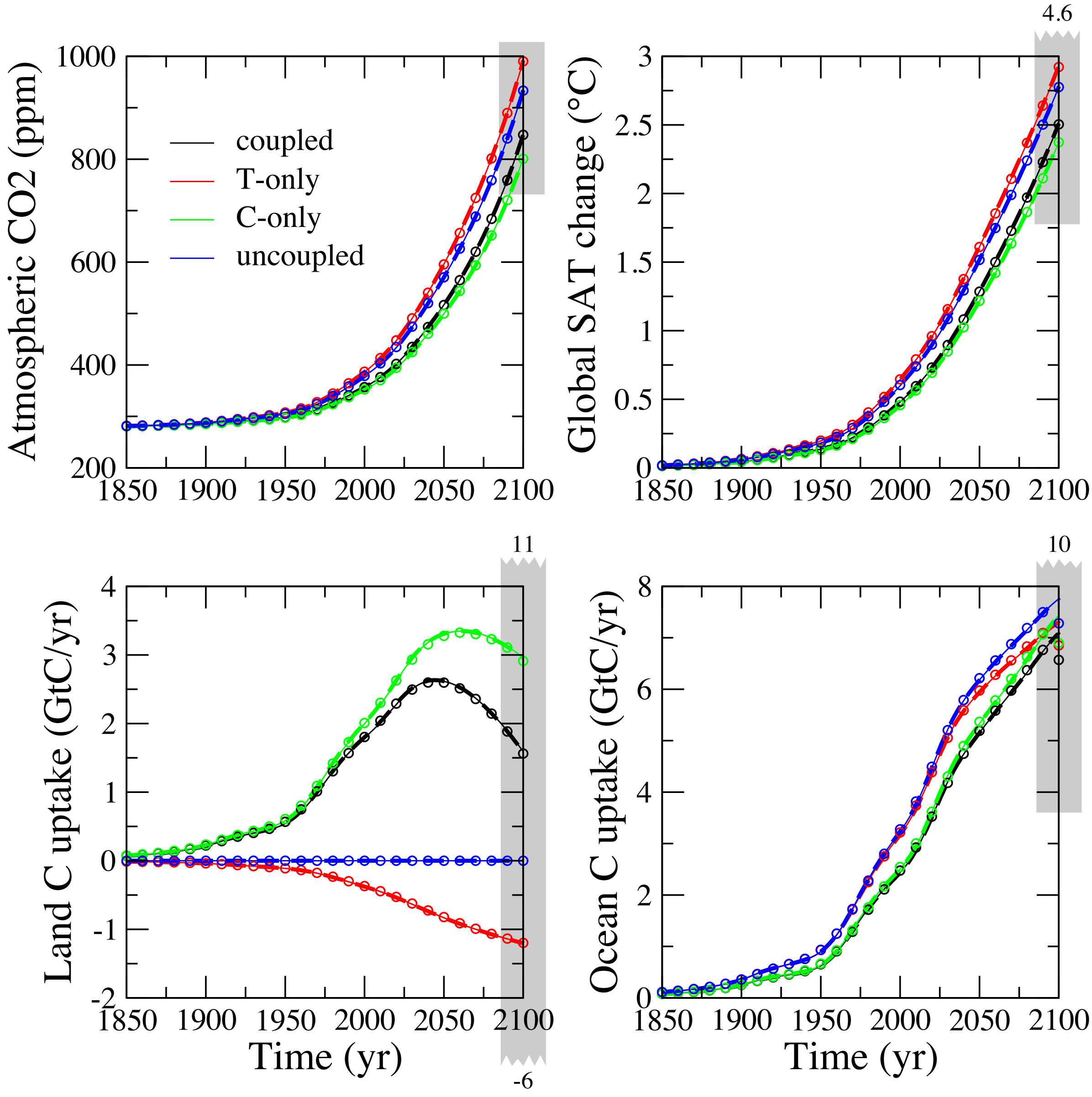

Figure 6. BernSCM simulations of the SRES A2 scenario used for C4MIP, with a climate sensitivity of 2.5°C and the HILDA/HRBM ocean/land components. Results for three numerical schemes are overlaid; 0.1 yr Euler forward timestep (solid thin line), ii. 1 yr implicit timestep (dashed bold line), iii. 10 yr implicit timestep with piecewise linear approximation of fluxes (circles); the difference at this resolution is only visible in the C uptake. The C4MIP model range at 2100 is indicated by grey bars; numbers above or below the bars indicate values outside of the chart range.

Table 1. C4MIP sensitivity metrics. The BernSCM range covers the carbon cycle settings as discussed in section 3.1, and different combinations of model components (HILDA-HRBM, HILDA-4box, Princeton-HRBM); the C4MIP range covers all participating models.

	α	β_L	β_O	γ_L	γ_O	g
Unit	$10^{-3}\ \frac{°C}{\mathrm{ppm}}$	$\frac{\mathrm{GtC}}{\mathrm{ppm}}$	$\frac{\mathrm{GtC}}{\mathrm{ppm}}$	$\frac{\mathrm{GtC}}{\mathrm{K}}$	$\frac{\mathrm{GtC}}{\mathrm{K}}$	10^{-2}
			BernSCM			
Standard	4.4	0.75	1.2	-46	-31	8.3
Range	4.1–4.6	**0**–0.75	1.0–1.2	-46–**0**	-31–**0**	0–8.4
			C4MIP ensemble			
Average	6.1	1.35	1.13	-79	-30	15
Range	3.8–8.2	0.2–2.8	0.8–1.6	-177– -20	-67– -14	4–31

Table A1. Model variables

Variable	Meaning	Unit
m_A	atmospheric Atmospheric CO_2 carbon	GtC
m_L	land Land biomass carbon	GtC
m_O ocean carbon perturbation GtC m_S	dissolved Dissolved inorganic C perturbation in ocean mixed layer	GtC
ΔDIC	perturbation Perturbation of dissolved inorganic C concentration in mixed layer	μmol/kg
$p_{A/S}^{CO_2}$	atmosphericAtmospheric/ocean surface CO_2 pressure	ppm
RF	Radiative forcing	Wm^{-2}
ΔT	global Global mean surface (ocean) temperature perturbation	K °C
ΔT^{eq}	Equilibrium ΔT for current RF	°C
e	CO_2 emissions	GtC/yr
f_A net flux to atmosphere flux GtC/yr f_O	air-sea Air-sea C flux	GtC/yr
f_{deep}	Flux Net C flux from mixed layer to deep the deep ocean	GtC/yr
f_{NPP}	NPP	GtC/yr
f_{decay}	decay Decay of terrestrial biomass C	GtC/yr
f_O^H	air-sea Air-sea heat flux	PetaW W
$f_{O,deep}^H$	Net heat flux from mixed layer to the deep ocean	W

Table A2. ~~Parameter definitions~~Model parameters

Parameter	Meaning	~~Value~~ Unit	HIL
H_{mix}	~~depth~~ Depth of mixed ocean surface layer	~~50-75~~[a] ~~m~~ m	7
A_O	Ocean surface area	~~$3.62\cdot10^{14}$ m^2~~ m^{-2}	3.62·
k_g	Gas exchange coefficient	yr^{-1} A_O^{-1}	1/9
T^*	Global average ocean surface temperature	°C	18
			All m
a_O	Ocean fraction of earth surface	-	0.
ε	Atmospheric ~~concentration per~~ mass of C per mixing ratio	~~2.123~~ GtC/ppm ~~/GtC~~	2.1
ϱ	~~density~~ Density of ocean water[b]	~~1028~~ kg/m^3	1028 (
c_p	Specific heat capacity of water	~~4000~~ J/kg/K	40
c_s	~~mixed~~ Mixed layer heat capacity	~~$c_p\,\varrho H_{\mathrm{mix}} A_O k_g$ gas exchange coefficient 1~~J/~~(9.06 yr·A_O)~~ K	$c_p\,\varrho H$
$M_{\mu\mathrm{mol}}$	~~micromol mass of DIC~~ Mass of DIC per micromole	gC/μmol	12.0107
~~$\mathrm{RF}_{2\times}$~~ $\mathrm{RF}_{2\times}$	RF per doubling of atm. CO_2	~~3.708~~ Wm^{-2}	3.7
$\Delta T_{2\times}$	Equilibrium climate sensitivity for CO_2 doubling	°C	fr

[a] ~~Range for included ocean components~~

[b] The first value is used in the climate component equations, the value in parentheses in the C cycle component equations.

Table A3. Mixed-layer IRFsubstitute model components currently implemented in BernSCM, and the corresponding implemented dependencies on atmospheric and SAT (references for the parametrisations used are given in the footnotes)./Box parameters

Ocean substitute model Ocean C dependent on: HILDA									
HILDA Joos (1992)	Input coefficients	a	(-)	0.27830	0.24014	0.23337	0.13733	0.051541	0.03
Princeton GCM Sarmiento et al. (1992)	Time scales	τ	(yr)	0.45254	0.03855	2.1990	12.038	59.584	237
Bern2.5D Stocker et al. (1992)									
	Input coefficients	[a], SAT[b] a	(-)	0.27022	0.45937	0.094671	0.10292	0.0392835	0.01
Land substitute model	Time scales	τ	(yr)	0.07027	0.57621	2.6900	13.617	86.797	337
4box Siegenthaler and Joos (1992) Princeton GCM									
	Input coefficients	[c] a	(-)	2.2745	-2.7093	1.2817	0.061618	0.037265	0.01
HRBM Meyer et al. (1999)	Time scales	τ	(yr)	1.1976	1.5521	2.0090	16.676	65.102	347

[a]Joos et al. (1996b)
[b]Takahashi et al. (1993)
[c]Enting et al. (1994a); Schimel et al. (1996)
[d]Meyer et al. (1999)

Table A4. ~~C4MIP sensitivity~~ Land C stock IRF/Box parameters~~. The BernSCM range covers the carbon cycle settings as discussed in section 3.1, and different combinations of model components (HILDA-HRBM, HILDA-4box, Princeton-HRBM); the C4MIP range covers all participating models.~~

HRBM							
Input coefficients	~~α~~ a	~~β_L~~ (-)	~~β_O~~ -0.15432	~~γ_L~~ 0.56173	~~γ_O~~ 0.074870	~~g~~ 0.41366	0.10406
~~Unit~~ Time scales	~~$10^{-3}\frac{\mathrm{K}}{\mathrm{ppm}}$~~ τ	~~$\frac{\mathrm{GtC}}{\mathrm{ppm}}$~~ (yr)	~~$\frac{\mathrm{GtC}}{\mathrm{ppm}}$~~ 0.20107	~~$\frac{\mathrm{GtC}}{\mathrm{K}}$~~ 1.4754	~~$\frac{\mathrm{GtC}}{\mathrm{K}}$~~ 8.8898	~~10^{-2}~~ 74.098	253.81
~~Standard~~ sensitivities	~~4.4~~ s_a	~~0.75~~ (-)	~~1.2~~ 0.14	~~-46~~ 0.056	~~-31~~ 0.072	~~8.3~~ 0.044	0.069
~~Range~~	~~4.1–4.6~~ s_τ	~~0–0.75~~ (-)	~~1.0–1.2~~ 0.056	~~-46–0~~0.079	~~-31–0~~0.057	~~0–8.4~~ 0.053	0.036
~~Average~~ height4Box							
Input coefficients	~~6.1~~ a	~~1.35~~ (-)	~~1.13~~ -1.5675	~~-79~~ 2.0060	~~-30~~ 0.26828	~~15~~ 0.29323	
~~Range~~ Time scales	~~3.8–8.2~~ τ	~~0.2–2.8~~ (yr)	~~0.8–1.6~~ 2.1818	~~-177–-20~~ 2.8571	~~-67–-14~~ 20	~~4–31~~ 100	

Table A5. Performance and accuracy for time steps 1–10 yr relative to a reference with a time step of 0.1 yr. The reference simulation is solved explicitly, otherwise an implicit solution was used. The average execution time of the time integration loop is given as a fraction of the explicit case. For atmospheric CO_2 and SAT, the root mean square difference to the ~~explicite~~ explicit case, divided by the value range over the simulation is given. All values are for the C4MIP A2 scenario (years 1700 – 2100), using the HILDA ocean component and the HRBM land component with standard temperature and carbon cycle sensitivities (coupled).

Δt	1yr	10yr
discretization	piecewise const.	piecewise lin.
execution time	15%	2 %
CO2 RMS/range	0.31‰	0.45‰
SAT RMS/range	0.52‰	0.53‰