

Cosmogenic radiocarbon production and glacial cycles control marine radiocarbon archive variations

Corresponding Author: Dr Martin Butzin

This file contains all reviewer reports in order by version, followed by all author rebuttals in order by version.

Attachments originally included by the reviewers as part of their assessment can be found at the end of this file.

Version 0:

Reviewer comments:

Reviewer #1

(Remarks to the Author)

This very well written and beautifully illustrated manuscript describes climate model simulations of radiocarbon for the last 80,000 years, with a focus on Marine Reservoir Ages. I have only minor suggestions for improvements.

- 1) I would suggest to include statistics of model – data fit, such as root-mean-squared errors, bias (difference in means), and correlation coefficients for all comparisons shown in Figs. 2, 3 and 4
- 2) How realistic is the AMOC simulation? Recent papers by Oppo (10.1002/2014PA002667), Repschläger (10.1016/j.quascirev.2021.107145) and Poeppelmeier (10.1038/s41561-023-01140-3) suggest the AMOC was not shutdown during HS1. May warrant a brief discussion.
- 3) The point made in lines 118-122 is a good one. Consider mentioning it in the abstract. I think it will be useful for other modelers who don't usually do that.

See attached annotated documents for more minor issues.

In the caption of Fig. S1 there is a typo for the age of HS1.

It would be good to make the data available at a repository.

Andreas Schmittner

Reviewer #2

(Remarks to the Author)

The review is provided as attached pdf file

Version 1:

Reviewer comments:

Reviewer #1

(Remarks to the Author)

I'm satisfied with the revisions and recommend publication.

Reviewer #2

(Remarks to the Author)

The reference to Galbraith et al. (2015) points to a paper which is not relevant to the present work. The right reference is

another work by Galbraith et al., also published in 2015.

Hence, this reference

Galbraith, E.D., Dunne, J.P., Gnanadesikan, A., Slater, R.D., Sarmiento, J.L., Dufour, C.O., de Souza, 410 G.F., Bianchi, D., Claret, M., Rodgers, K.B., Marvasti, S.S., 2015. Complex functionality with 411 minimal computation: Promise and pitfalls of reduced-tracer ocean biogeochemistry 412 models. *Journal of Advances in Modeling Earth Systems* 7, 2012–2028. 413 <https://doi.org/10.1002/2015MS000463>

should be replaced by this one

Galbraith, E. D., E. Y. Kwon, D. Bianchi, M. P. Hain, and J. L. Sarmiento, 2015. The impact of atmospheric pCO₂ on carbon isotope ratios of the atmosphere and ocean. *Global Biogeochem. Cycles*, 29, 307–324. doi: 10.1002/2014GB004929.

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Reply to Review #1

We thank Andreas Schmittner his constructive review. Please find our detailed answers to the specific issues further below. Our answers as well as the changes made in the revised manuscript are highlighted in blue. Please note that the figures showing meridional ^{14}C age distributions now also include benthic-Planktic (B-P) ^{14}C ages, regardless of his review.

Reviewer #1 (Remarks to the Author):

This very well written and beautifully illustrated manuscript describes climate model simulations of radiocarbon for the last 80,000 years, with a focus on Marine Reservoir Ages. I have only minor suggestions for improvements.

1) I would suggest to include statistics of model - data fit, such as root-mean-squared errors, bias (difference in means), and correlation coefficients for all comparisons shown in Figs. 2, 3 and 4.

We now report these uncertainty measures in Table S1 that can be found in the Supplementary Information (lines 69-73). For better readability, we only report the biases in the main text and refer to Table S1 for the other uncertainty values.

2) How realistic is the AMOC simulation? Recent papers by Oppo (10.1002/2014PA002667), Repschlaeger (10.1016/j.quascirev.2021.107145) and Poepelmeier (10.1038/s41561-023-01140-3) suggest the AMOC was not shutdown during HS1. May warrant a brief discussion.

We now briefly discuss that an AMOC shutdown during HS1 has been questioned in view of proxy data other than ^{231}Pa / ^{230}Th , and that Pöppelmeier et al. (2023) reject this hypothesis because a shutdown would lead to unrealistically high benthic-planktic (B-P) ^{14}C ages of more than 2500 years in the North Atlantic. Regarding ^{14}C , we think that the evidence against an AMOC shutdown is not really convincing because our B-P ^{14}C ages in the North Atlantic are much lower (which is shown in an additional Supplementary Figure, Fig. S5 (lines 61-68 in the Supplementary Information). In fact, our B-P ^{14}C ages during HS1 are in line with a new high-resolution ^{14}C record from the equatorial West Atlantic (Jena et al., 2026). See lines 183-191 in the revised manuscript.

3) The point made in lines 118-122 is a good one. Consider mentioning it in the abstract. I think it will be useful for other modelers who don't usually do that.

Done, see lines 28-30.

See attached annotated documents for more minor issues.

In the caption of Fig. S1 there is a typo for the age of HS1.

Corrected to 17.5 ka, see line 29 in the Supplementary Material.

It would be good to make the data available at a repository.

We are going to upload the data to the PANGAEA repository as soon as the manuscript is accepted for publication.

Annotated issues:

Line 78: What's the cutoff?

The cutoff is 50° and is now mentioned in line 80

Lines 91-92: Is that shown somewhere?

Not really (but it could be inferred from Fig. 3). This is now clarified in line 94.

Line 98: how realistic is this AMOC evolution?

See our reply to suggestion #2 further above.

Lines 99-106: Report statistics such as RMS errors and correlation coefficients

RMS errors of modelled $\Delta^{14}\text{C}$ are 40 ‰ for CLIMBER-X and 45 ‰ for LSG, respectively. Model biases (the differences in modelled and observed means) are -7 ‰ for CLIMBER-X and -25 ‰ for LSG. Errors and biases are rounded to integers. Correlation coefficients are $r = 0.912$ for CLIMBER-X and $r = 0.922$ for LSG. The biases are reported in the revision in lines 104-105, RMS errors and correlation coefficients can be found in Supplementary Table S1 (lines 69-73).

Lines 118-122: Good point

Thank you. Note that it has been slightly rephrased following the suggestion of Reviewer #2 (line 126).

Line 123: Report statistics such as mean (bias), RSME, r

The overall model bias is -379 ^{14}C years with an RMSE of 653 ^{14}C years; $r = 0.416$. In the main text we report only the bias (lines 131-133) and refer to Supplementary Table S1 for the other values (lines 69-73 in the Supplementary Information).

Lines 128-130: Could it also be an issue with the data? There are a few very high data points also in the tropics, which seem unrealistic.

This is a valid point but it is difficult for us as modellers to judge the data quality of observations. We now note that at some locations the observations exhibit questionably strong scatter (lines 144-145). Note that Figure 3c has been redrawn now showing the average of multiple observations at the same site; however, the resulting changes are barely visible.

Line 142: Report statistics

The overall bias is -483 ¹⁴C years with an RMSE of 649 ¹⁴C years; r = 0.753. In the main text we report only the bias (lines 152-153) and refer to Table S1 for all other numbers (lines 69-73 in the Supplementary Information).

Lines 172-173: Seems inconsistent with recent work by Oppo (10.1002/2014PA002667), Repschlaeger (10.1016/j.quascirev.2021.107145) and Poeppelmeier (10.1038/s41561-023-01140-3)

See our reply to suggestion #2 further above.

Line 176: What does the acronym stand for?

“Virtual Axial Dipole Moments”, referring to the palaeomagnetic production reconstruction. We decided to rename this experiment to “P_pmag” where “pmag” stands for “paleomagnetic reconstruction” (line 165).

Line 243: How long did the full run take?

Do you mean in real time? A simulation over 70 kyr takes about 10 days using 36 nodes at AWI's NEC Linux cluster “Albedo” which is now mentioned at lines 243-244.

Line 552: Specify if this is surface only or full depth average.

The results refer to the sea surface which is clarified in line 600.

Reference

Jena, P. S., Chiessi, C. M., Beese, I., Butzin, M., Dias, B. B., Campos, M. C., Lohmann, G., Mulitza, S., 2026 (in press). Centennial-scale intensifications of the Atlantic Meridional Overturning Circulation during Heinrich Stadial 1. Nature Communications, <https://doi.org/10.1038/s41467-026-73364-x>

Reply to Review #2

We thank Reviewer #2 for his / her constructive review. Please find our detailed answers to the specific issues further below. Our answers as well as the changes made in the revised manuscript are highlighted in blue. Please note that the figures showing meridional ^{14}C age distributions now also include benthic-Planktic (B-P) ^{14}C ages, regardless of his / her review.

Comments on the manuscript by Butzin et al.

"Marine radiocarbon reservoir age simulations for the past 50,000 years revisited" (COMMSENV-25-5679)

In this work the authors study the evolution of Marine Reservoir Age (MRA) over the past fifty thousand years with the help of an Earth system model of intermediate complexity.

There are two distinct parts in their work.

First they constrain the atmospheric $\Delta^{14}\text{C}$ to follow the Intcal20 and SHcal20 reconstructions. Under that scenario, they compare the MRAs with those previously obtained with an ocean global circulation model under different circulation scenarios (hence, the mention 'revisited' in the title of the paper). In this first part they also examine the impact of initial conditions on the final radiocarbon distribution in the ocean. In the second part, the model is forced with cosmogenic radiocarbon production in the atmosphere. Two sensitivity experiments were performed, one with a reconstructed rate, and one with a constant rate, allowing for disentangling the influence of ocean circulation changes and changes in the production rates on MRA. A very short discussion on the interhemispheric $\Delta^{14}\text{C}$ gradient then follows. The authors then conclude that the interhemispheric $\Delta^{14}\text{C}$ gradient was much larger during the Laschamps event (~ 41 ka BP) than predicted by the reconstructions.

The manuscript is well structured and clear to read. However, there are several points of concern.

I am not convinced by the explanation of the difference between the MRA predicted by the LSG and by CLIMBER-X (lines 95-99). Given that the air-sea exchange of radiocarbon is very slow, a more vigorous ocean ventilation would result in an increase of the difference between the radiocarbon content in the surface layer and in the atmosphere, since the faster renewal of surface waters would allow for less time to equilibrate. I suspect that the larger MRA obtained with the LSG are due to a more vigorous vertical mixing in that model. Did the authors examine the difference in background vertical mixing (explicit and numerical) and mixed layer depth between the LSG and CLIMBER-X?

In a homogeneous system (or averaged over large scales) the opposite is the case: A more vigorous meridional overturning circulation (MOC) decreases the gradient between the atmosphere (where there is the only source of ^{14}C) and the deep sea (sink) because the interior of the ocean is faster renewed. Having said that, we agree that mixing may play a more important role for MRAs than the MOC. In fact, the vertical diffusivity in the LSG model is significantly stronger than in CLIMBER-X, namely up to three times in the upper 500 m of the ocean. In addition, the LSG model does not include a specific mixed layer model while this is the case in CLIMBER-X.

As a result, mixed layer depths (considering a density threshold of $\Delta\sigma_\theta = 0.125 \text{ kg m}^{-3}$ with respect to the sea surface following Monterey and Levitus, 1997) in subpolar and subtropical subduction regions appear to be higher in the LSG simulations than in CLIMBER-X. However, the size of this effect is somewhat uncertain as only annual-mean temperatures and salinities are available from the LSG model (which was run for the last time in 2020 at a computer which is no longer in operation). We rephrased as follows (lines 97-102): “This is due to weaker vertical mixing in the upper ocean and stronger near-surface winds in CLIMBER-X. Both lead to a smaller isotopic disequilibrium between the sea surface and the atmosphere than in the LSG model, in which the upper (lower) bound of MRAs corresponds to the scenario with the weakest (strongest) $^{14}\text{CO}_2$ air-sea fluxes as well as the weakest (strongest) meridional overturning circulation in the Atlantic (AMOC).”

The transient climate simulation should be discussed further. Over the deglaciation, the results presented here differ in several aspects from previously published ones under the same ice-sheet scenario (Masoum et al., 2024). For example, in this study, the changes in AMOC at the end of Heinrich 1 event are much larger than obtained by Masoum et al. (2024) in the PaleoMist scenario.

The AMOC changes in our paper are much larger because we apply a more recent model version of CLIMBER-X than Masoum et al., 2024 (version 3 instead of version 2). Version 3 includes an updated parametrization of frictional processes in the bottom water layer of the ocean caused by the topographic drag. This is now clarified in line 254-256. See also our reply to Reviewer #1 and lines 183-191 in the revision regarding the realism of an AMOC shutdown.

How well is the present-day ocean state represented at the end of the long transient simulation? I agree that $\Delta^{14}\text{C}$ in 1990 is reasonably reproduced (Fig. 4), but how do temperature and salinity distributions compare to observations?

We now compare simulated temperatures and salinities with gridded observations according to World Ocean Atlas 2023 (WOA23; decadal averages of the period 1985-94 corresponding to the period of GLODAP bomb ^{14}C observations). Temperature biases are 0.43°C at the sea surface (SST) and -0.23°C considering zonal-mean values in the interior of the Atlantic and Pacific (ZMT). Salinity biases are 0.10 at the sea surface (SSS) and 0.01 considering zonal-mean values in the interior of Atlantic and Pacific (ZMS). Regional uncertainties are summarized in terms of root mean square errors (RMSE) and correlation coefficients (r). Values for temperature are $\text{RMSE}_{\text{SST}} = 2.0^\circ\text{C}$, $\text{RMSE}_{\text{ZMT}} = 1.51^\circ\text{C}$, $r_{\text{SST}} = 0.982$, and $r_{\text{ZMT}} = 0.979$. Regional uncertainties of salinities are $\text{RMSE}_{\text{SSS}} = 1.8$, $\text{RMSE}_{\text{ZMS}} = 0.37$, $r_{\text{SSS}} = 0.718$, and $r_s = 0.854$. See the new Figure S2 and lines 34-43 in the Supplementary Information.

The differences observed in 1950 and 1990 between experiments IntSH20 and SpinHist (Fig. 4) are not solely explained by the longer or shorter transient ^{14}C input history as claimed on lines 120-122. In 1950 or 1990, the ocean circulation obtained with IntSH20 is most probably very different from that obtained with SpinHist. Indeed, in IntSH20, the ocean has been adjusting to

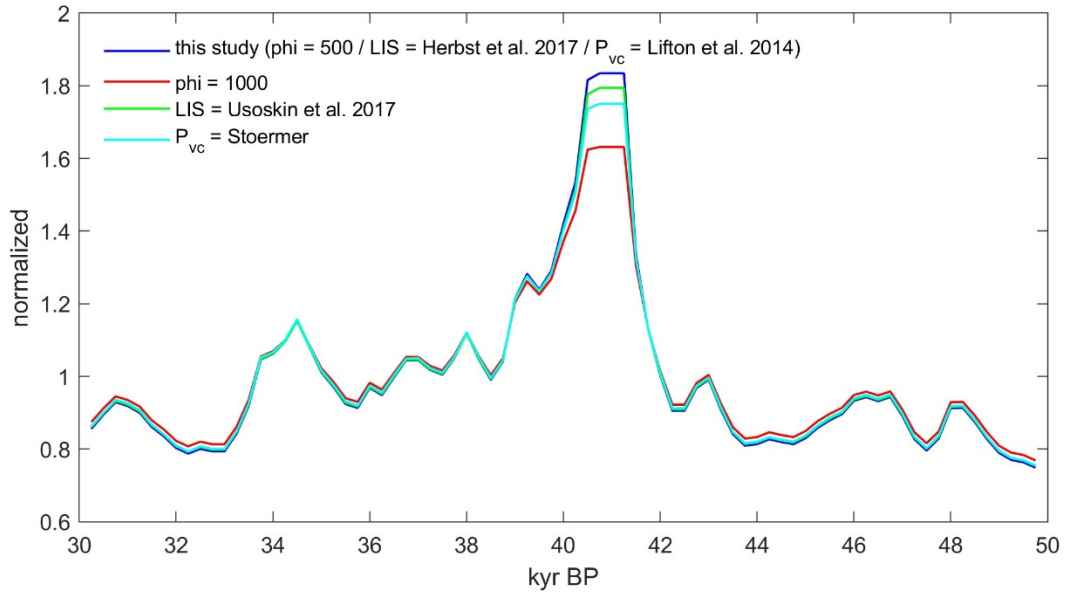
a varying climate forcing over 70 kyr in addition to the transient ^{14}C input. The SpinHist experiment which starts in 1850 from a long spin-up with steady climate forcing is not influenced by the long term climate trend. Differences in the radiocarbon distribution between IntSH20 and SpinHist also result from the different climate scenarios and are not only due to the duration of the transient atmospheric radiocarbon forcing.

We agree and modified the sentence accordingly to “This indicates that a thorough model evaluation with ^{14}C observations is misleading if a simulation does not include a sufficiently long (millennial rather than centennial) transient climate and ^{14}C input history”, line 126.

The reconstruction of the radiocarbon production rate, which is central to this study, is not sufficiently discussed. What is the sensitivity of the production rate to the various parametrizations? How does the reconstruction of ^{14}C production in this study compare to other reconstructions?

As described in lines 331-334, we spliced three geomagnetic field reconstructions into one continuous record to estimate geomagnetic field intensity changes over the last 70,000 years. Obviously, other production rate reconstructions are available (for a comparison see e.g. Dinuer et al. 2020 or Köhler et al. 2022) but their differences are not resolved and it is beyond the scope of the paper to attempt to do so. Compared to previous reconstructions, our VADM-based ^{14}C -production rate has a stronger long-term trend, and hence, comes closer to the 700-800 ‰ during the Laschamps excursion but instead is 50 ‰ too high during the preindustrial (lines 221-225 in the main text).

Regarding the parametrization of the production rate calculation from VADM, the sensitivity to the applied parameters is very low (typically below 2 %). This is because the reconstructed production rates are normalized and used as a multiplier of the steady-state ^{14}C production rate in the model. Hence, choosing a different solar modulation potential, a different Local Interstellar Spectrum, or a different parametrization for the derivation of cutoff-rigidities from VADM are below 2 % (see the figure below). The only exception is during the peak of the Laschamps geomagnetic field minimum, where a systematically different solar modulation would have larger effects. However, there is no evidence for a such a difference and solar cycles occurred in a similar manner as today (Wagner et al., 2001).



We added to the section “Reconstruction of ^{14}C production rate changes” (lines 339-345):

“The used production rate changes are relatively insensitive to the exact choice of the above mentioned parameters due to their normalization prior to multiplication with $^{14}\text{P}_\text{C}^*$. Other sets of parameters yield differences of less than 2 % except during the period of the lowest VADM during the Laschamps event, in case systematically different solar modulation is assumed. However, there is no evidence for. However, there is no evidence for a such a difference and solar cycles occurred in a similar manner as today (Wagner et al., 2001).”

It is not clear whether the ^{14}C production rate at a given time is homogeneous over the globe or varies with location. Are the $^{14}\text{P}_\text{C}$ rates in the three latitude bands and in Eqs (5) to (7) identical? I suppose they are not. Could the authors illustrate the temporal evolution of $^{14}\text{P}_\text{C}$ in the three latitude bands?

The ^{14}C production rates are globally homogeneous, which is clarified in lines 306-307.

Equations (5) to (7) are erroneous. It seems there is a confusion between terms driving the radiocarbon ratio (^{14}R) and terms controlling the radiocarbon concentration (^{14}C) in the atmosphere.

We disagree, in equations (5) to (7) the total amount of ^{14}C is conserved. We demonstrate this for the mass balance of total ^{14}C the northern atmosphere zone which reads

$$(i) \quad d/dt \, ^{14}\text{M}_{\text{A},\text{N}} = ^{14}\text{P}_{\text{C},\text{N}} - \lambda \, ^{14}\text{M}_{\text{A},\text{N}} - ^{14}\text{H}_{\text{A},\text{N} \rightarrow \text{A},\text{T}} - \iint ^{14}\text{F}^*_{\text{A},\text{N} \rightarrow \text{O},\text{N}} dS_{\text{N}}$$

where

$$(ii) \quad ^{14}\text{M}_{\text{A},\text{N}} = ^{14}\text{R}_{\text{A},\text{N}} x_{\text{A}} \text{M}_{\text{A},\text{N}}$$

is the total amount of ^{14}C in the northern atmosphere (mol s^{-1}) for a CO_2 volume mixing ratio x_A and total molecular inventory $M_{A,N}$, and $^{14}\text{P}_{C,N}$ and λ $^{14}\text{M}_{A,N}$ denote production and radioactive decay of total ^{14}C (each in mol s^{-1}). The third term at the right-hand side in (i) is the hemispheric exchange rate (mol s^{-1}) between the northern (N) and tropical (T) atmosphere zones,

$$(iii) \quad {}^{14}\text{H}_{A,N \rightarrow A,T} = -\mu_{N \rightarrow T} {}^{14}\text{R}_{A,N} x_A M_{A,N} + \mu_{T \rightarrow N} {}^{14}\text{R}_{A,T} x_A M_{A,T}$$

$$(iv) \quad {}^{14}\text{H}_{A,N \rightarrow A,T} = -\mu_{N \rightarrow T} {}^{14}\text{R}_{A,N} x_A M_A / 4 + \mu_{T \rightarrow N} {}^{14}\text{R}_{A,T} x_A M_A / 2 = -\mu_{N \rightarrow T} ({}^{14}\text{R}_{A,N} - {}^{14}\text{R}_{A,T}) x_A M_A / 4$$

with

$$(v) \quad M_A = 4 M_{A,N} = 2 M_{A,T} = \text{total molecular inventory of the global atmosphere.}$$

The fourth term at the right-hand side in (i) is the air-sea exchange flux of ^{14}C integrated over the surface area dS_N (mol s^{-1}). Applying equation (2) from the paper,

$$(vi) \quad \iint {}^{14}\text{F}_{A,N \rightarrow O,N}^* dS_N = \iint {}^{14}\text{F}_{A,N \rightarrow O,N} C_{\text{DIC}_0} dS_N.$$

In the following we apply equation (v) to (ii) and to $^{14}\text{P}_{C,N} = {}^{14}\text{P}_C / 4$, where $^{14}\text{P}_C$ is the global production rate. Inserting (ii), (vi), (v), and (vi), equation (i) becomes

$$(vi) \quad \begin{aligned} & [(d/dt {}^{14}\text{R}_{A,N}) x_A + {}^{14}\text{R}_A (d/dt x_A)] M_A / 4 \\ & = {}^{14}\text{P}_C / 4 - \lambda {}^{14}\text{R}_{A,N} x_A M_A / 4 - \mu_{N \rightarrow T} ({}^{14}\text{R}_{A,N} - {}^{14}\text{R}_{A,T}) x_A M_A / 4 - \iint {}^{14}\text{F}_{A,N \rightarrow O,N} C_{\text{DIC}_0} dS_N \end{aligned}$$

or, with $(d/dt x_A) / x_A = d/dt \ln x_A$,

$$(vii) \quad \begin{aligned} & d/dt {}^{14}\text{R}_{A,N} + {}^{14}\text{R}_A (d/dt \ln x_A) \\ & = {}^{14}\text{P}_C / (x_A M_A) - \lambda {}^{14}\text{R}_{A,N} - \mu_{N \rightarrow T} ({}^{14}\text{R}_{A,N} - {}^{14}\text{R}_{A,T}) - 4 \iint {}^{14}\text{F}_{A,N \rightarrow O,N} C_{\text{DIC}_0} dS_N / (x_A M_A) \end{aligned}$$

Equation (vii) is equation (5) in the paper, except for the denominator below $^{14}\text{P}_C$ (which was originally defined differently and implicitly included $x_A \cdot M_A$; this has now been changed for clarity in lines 313-315) and for the small term $d/dt \ln x_A$ on the left-hand side. In the model this term is formally neglected but implicitly taken into account, as atmospheric CO_2 is updated at each time step. The derivation of equations (6) and (7) is analogous.

Line 39 : that we carried out... (suppress the 'had').

Changed accordingly, see line 41.

Figure 1 & line 84: With respect to the timing of Laschamps event, it would help the reader to superimpose a plot of the reconstructed $\Delta^{14}\text{C}_A$ in Figure 1.

Done, see the new Figure 1 in the revision (line 596).

Line 77: Could the authors document the difference between the atmospheric $\Delta^{14}\text{C}$ reconstructions used with CLIMBER-X and with LSG?

The LSG model simulations were forced with globally homogeneous values of atmospheric $\Delta^{14}\text{C}$ according to IntCal20 which is clarified in lines 604-605.

Line 97: Since the envelope of MRA obtained with the LSG results from the combination of three scenarios (PD, GS, and CS), it is not only the AMOC strength in scenario GS that should be considered here.

This is certainly true and has been rephrased (lines 97-102), see also our reply to the to first comment.

Line 164: there is no Section 3,3.

Corrected to “discussed further below“, line 171

Line170: Galbraith et al. (2015) is not the right reference here.

We would appreciate if Reviewer #2 could explain why this paper should not be the right reference here. Galbraith et al. (2015) present a derivation of the different air-sea equilibration time scales of ^{12}C , ^{13}C , and ^{14}C . They show that the equilibration of ^{14}C directly depends on $p\text{CO}_2$ (while the equilibration of ^{12}C depends on the CO_2 buffer capacity; cf. their equations (A1) - (A15)), and they explicitly discuss the consequences of this effect on MRAs (cf. Section 4.2 in their paper).

Lines 172 & 174: adding the evolution of atmospheric $p\text{CO}_2$ in Figs 1 and 5 would really help the reader.

Figures 1 and 5 have been enhanced accordingly, see lines 596 and 635.

Line 212: Dinauer et al. (2020) is missing in the reference list.

Amended, cf. lines 394-396.

Line 240: Nonetheless, our simulations *predict* less sea ice... (instead of ‘arrive at’).

Changed accordingly, see line 261.

Lines 249 & 318: The production rate should be in $\text{atoms cm}^{-2} \text{s}^{-1}$.

Corrected, see lines 270 and 307.

Line 269: The units of $^{14}\text{F}_\text{A}$ as given by Eq. (2) are m s^{-1} , not $\text{m}^{-2} \text{s}^{-1}$.

Corrected, see line 290.

Line 305: The age that is reported by Eq. (8) is the Marine Radiocarbon Age (MRA); hence, line 305 should read: ‘We report *marine* radiocarbon ages...’.

Changed accordingly, see line 326.

Figure 3a: the caption mentions that the observations by Rubin and Key are identified by means of triangles. However, I do not identify any triangle in Fig. 3a.

Amended, see Fig. 3a near line 614 (it is a mystery to us how the symbols could disappear from the figure).

Figure S1, caption: the range for H1 is 15-17.5, not 15-7.5.

Corrected, see line 29 in the Supplementary Information.

References

Köhler, P., Adolphi, F., Butzin, M., Muscheler, R., 2022. Toward Reconciling Radiocarbon Production Rates With Carbon Cycle Changes of the Last 55,000 Years. *Paleoceanography and Paleoclimatology* 37, e2021PA004314. <https://doi.org/10.1029/2021PA004314>

Monterey, G. I., Levitus, S., 1997. Seasonal variability of mixed layer depth for the world ocean. NOAA atlas NESDIS 14. United States. National Oceanographic Data Center (U.S.), Ocean Climate Laboratory. <https://repository.library.noaa.gov/view/noaa/49153>

Wagner, G., Beer, J., Masarik, J., Muscheler, R., Kubik, P.W., Mende, W., Laj, C., Raisbeck, G.M., Yiou, F., 2001. Presence of the solar de Vries cycle (~205 years) during the last ice age. *Geophysical Research Letters* 28, 303–306. <https://doi.org/10.1029/2000GL006116>

Reply to the Reviews

Our answers are highlighted in blue.

Reviewer #1 (Remarks to the Author): I'm satisfied with the revisions and recommend publication.

We thank Reviewer #1 for reviewing.

Reviewer #2 (Remarks to the Author):

The reference to Galbraith et al. (2015) points to a paper which is not relevant to the present work. The right reference is another work by Galbraith et al., also published in 2015. Hence, this reference

Galbraith, E.D., Dunne, J.P., Gnanadesikan, A., Slater, R.D., Sarmiento, J.L., Dufour, C.O., de Souza, 410 G.F., Bianchi, D., Claret, M., Rodgers, K.B., Marvasti, S.S., 2015. Complex functionality with 411 minimal computation: Promise and pitfalls of reduced-tracer ocean biogeochemistry 412 models. Journal of Advances in Modeling Earth Systems 7, 2012–2028. 413 <https://doi.org/10.1002/2015MS000463>

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Galbraith, E. D., E. Y. Kwon, D. Bianchi, M. P. Hain, and J. L. Sarmiento, 2015. The impact of atmospheric pCO₂ on carbon isotope ratios of the atmosphere and ocean. Global Biogeochem. Cycles, 29, 307–324. doi: 10.1002/2014GB004929.

We thank Reviewer #2 for reviewing and replaced the erroneous reference accordingly, see lines 376-378 in the revision.

Marine radiocarbon reservoir age simulations for the past 50,000 years revisited

Manuscript submitted for publication in Communications Earth & Environment, November 2025

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Abstract

Marine radiocarbon (^{14}C) archives are depleted relative to the atmospheric ^{14}C reservoir, which is commonly expressed as the Marine ^{14}C Reservoir Age (MRA). Here we apply the Earth system model of intermediate complexity CLIMBER-X equipped with ^{14}C to simulate the MRA evolution over the past 50,000 years. Globally, our simulated MRAs range from 400 to 1200 ^{14}C years and are systematically lower than previous estimates from the uncoupled LSG ocean general circulation model, which is mainly due to stronger ocean ventilation in CLIMBER-X. Short-term MRA variations are dominated by changes in cosmogenic ^{14}C production, while glacial-interglacial climate variations exert slower and damped effects. From the coupled evolution of atmospheric and marine ^{14}C in these simulations we infer that during the Laschamps geomagnetic excursion the interhemispheric ^{14}C gradient in the atmosphere was up to twice as large as previously estimated.

Introduction

The most recent radiocarbon (^{14}C) calibration curve IntCal20 (Reimer et al., 2020) has not been entirely derived from terrestrial ^{14}C archives. For times prior to the continuous tree record (~ 14 ka BP) IntCal20 has also considered absolutely dated speleothems, terrestrial macrofossils, and marine records from corals as well as foraminifera. Marine ^{14}C archives are systematically depleted relative to the atmospheric ^{14}C reservoir, which must be corrected for in ^{14}C dating. The depletion is typically expressed in terms of the Marine Reservoir Age (MRA) which is the ^{14}C age of seawater relative to the overlaying atmosphere. In the case of IntCal20, the MRA correction of the marine archives was supported by long-term simulations of ^{14}C that we had carried out with the Large-Scale Geostrophic (LSG) ocean general circulation model (Butzin et al., 2020). The LSG model was not fully coupled to the atmosphere. It considered transient input values of atmospheric ^{14}C (reconstructed from the Hulu cave record; Cheng et al., 2018) but did not incorporate transient climate forcing. Instead, the LSG model considered various characteristic scenarios of constant climate forcing intended to represent the range of past glacial to deglacial climate variability. The original idea of this approach was to vary (or mix) between the upper and lower bounds of the climate scenarios and their corresponding MRA outcomes by means of expert knowledge. However, to avoid the potential pitfalls of such subjective reasoning, IntCal20 ultimately only considered the MRA results of a single intermediate climate scenario. A further shortcoming was that the LSG model did not include a prognostic sea ice module but inherited sea ice from the atmosphere general circulation model ECHAM, which also provided the other climate forcing fields (Lohmann & Lorenz, 2000). The ECHAM model itself only included reconstructed glacial sea surface temperatures which are considered outdated now (CLIMAP Project Members, 1981; Sarnthein et al., 2003). All in all, this ^{14}C modelling approach is no longer up to date. Here, we replace the LSG model with a more comprehensive model that overcomes the aforementioned limitations.

Simulations of marine ^{14}C are computationally challenging, even if the entire ^{14}C dating range of $\sim 50,000$ years is not targeted. The deep sea contains a large pool of slowly decaying dissolved inorganic ^{14}C which is only slowly ventilated with atmospheric $^{14}\text{CO}_2$. This implies that a well-defined initial state is only reached after several thousands of years of model integration (e.g., Aumont et al., 1998; Orr et al., 2017). In addition, marine ^{14}C observations of the past are often based on corals and sediment cores from sites that would require high spatial model resolution, which would further increase the computational costs. We are not aware of a comprehensive Earth system model that could be applied to simulate ^{14}C over tens of thousands of years at such sufficiently high spatial resolution in the ocean. Here, we prioritize model speed over resolution and apply the Earth System Model of Intermediate Complexity (EMIC) CLIMBER-X (Willeit et al., 2022; 2023) that we have equipped with ^{14}C to revisit our previous MRA simulations.

In the following, we present new MRA simulation results obtained with prescribed atmospheric ^{14}C , compare them with previous LSG model results, and discuss potential implications for the

above-mentioned initial value problem as well as for the calibration of ocean models using ^{14}C . Then we will broaden the perspective and discuss ^{14}C simulations with cosmogenic ^{14}C production forcing. While these simulations primarily aim at an improved understanding of the drivers of past MRA variability, we will also address aspects of atmospheric ^{14}C variations in the past.

Results and discussion

MRA evolution in simulations with prescribed atmospheric $\Delta^{14}\text{C}$

In this scenario CLIMBER-X is driven with reconstructed values of atmospheric $\Delta^{14}\text{C}$ ($\Delta^{14}\text{C}_A$), similar (but not identical) to the simulations with the LSG model. First we consider the long-term evolution of global-mean MRAs, disregarding the polar latitudes where sea ice could cause considerable regional ^{14}C age anomalies. Our simulated global MRAs vary between ~ 400 and 1200 ^{14}C years, where the highest values are reached at 42 ka BP shortly before the geomagnetic field minimum of the Laschamps geomagnetic excursion (~ 41 ka BP, e.g. Channell and Vigliotti, 2019) when the rates of change in geomagnetic field strength, and thus, transient changes in ^{14}C production, are largest (Fig. 1). Similar to our previous simulations with the LSG model (Butzin et al., 2017; 2020), this major MRA excursion occurs prior to the reconstructed peak of $\Delta^{14}\text{C}_A$ in the atmosphere. Between 30 and 20 ka CLIMBER-X simulates a slow decline of global MRAs from about 700 to 600 ^{14}C years which is superimposed by centennial-scale MRA fluctuations of about ± 100 ^{14}C years. During the last deglaciation the global MRA declined to ~ 400 ^{14}C years in the early Holocene which was interrupted by two MRA excursions of 100 to 200 ^{14}C years near 15 and 13 ka BP, shortly before the Bølling-Allerød (BA) and the Younger Dryas (YD) periods, respectively. In the Holocene the global MRA varied between 400 and 600 ^{14}C years with a preindustrial value of 430 ^{14}C years in 1850 CE. In the 20th century the global MRA declined to 320 ^{14}C years in 1950 CE, which is slightly higher than the value of 290 ^{14}C years that can be estimated from gridded preuclear GLODAP observations (Key et al., 2004). Overall, the long-term MRA transients of CLIMBER-X and LSG tend to run parallel, but the CLIMBER-X MRA values are systematically lower. This can be explained with a stronger ocean ventilation in CLIMBER-X, which is due to a stronger meridional overturning circulation in the Atlantic (AMOC, by about 40 % compared to LSG scenario GS regarding the long-term average of the streamfunction maximum; see also Supplementary Fig. 1) as well as due to a higher gas transfer velocity for $^{14}\text{CO}_2$ (at the LGM by about 4 % when globally averaged between 50°N - 50°S; not shown).

A comparison with $\Delta^{14}\text{C}_{DIC}$ records of (sub) tropical corals between 13 and 25 ka BP yields that the CLIMBER-X values roughly align with the observations (Fig. 2a). On the other hand, the LSG results for $\Delta^{14}\text{C}_{DIC}$ are mostly lower than the observations (Fig. 2b) which translates to systematically elevated regional MRAs. This indicates that, at least between 13 and 25 ka BP and in the low-latitude regions, the revised CLIMBER-X MRAs are more realistic than the previous LSG model results.

As $\Delta^{14}\text{C}_{\text{DIC}}$ and hence MRA values can vary significantly between regions, we take a closer look at the spatial distribution of MRAs in the recent past and during the LGM. Our starting point is the 20th century as this is the period with the most comprehensive database of marine ^{14}C observations. We focus on the pre-nuclear state in 1950 CE which is more representative for the Holocene than more recent decades. Our simulation captures the large-scale pattern of observed MRAs, such as minimum values in the subtropics and maximum values in the Southern Ocean (Fig. 3a). However, the model overestimates MRAs in the tropical oceans, particularly in the eastern upwelling zones. Our simulated MRAs also tend to be too high in the northern North Atlantic while they tend to be too low in the Southern Ocean. These biases become more apparent after the bomb ^{14}C pulse in the 1990s (Fig. 3b). We note that the bomb pulse not only increases ^{14}C ages at the surface but also in the deep sea, where it also amplifies model biases such as the overly weak and shallow ventilation, visible in higher benthic-atmospheric (B-Atm) ^{14}C ages than observed (Fig. 4). We also note that another bomb pulse ^{14}C simulation that was directly started from the preindustrial steady-state conditions at the end of the initial model spin-up displays different biases (also shown in Fig. 4). This indicates that a thorough model evaluation with ^{14}C observations is misleading if a simulation does not include a sufficiently long (millennial rather than centennial) transient ^{14}C input history.

Turning to the LGM (here defined as the period 19.0 - 21.8 ka BP following Skinner et al., 2023), we find that CLIMBER-X yields MRAs ranging from 300 ^{14}C years in the subtropical oceans to about 1600 ^{14}C years in the Southern Ocean (Fig. 3c), which results in global-mean MRA values of about 600 ^{14}C years within 50°N and 50°S, and of 670 ^{14}C years including the polar oceans. Our simulation does not reproduce the high MRAs of about 2000 ^{14}C years (or more) observed at subarctic and subantarctic sites. This is probably due to the insufficient sea ice extent of CLIMBER-X in these regions compared to reconstructed values (Paul et al., 2021; see Supplementary Fig. S2), which could lead to excessive air-sea exchange of $^{14}\text{CO}_2$ there. A striking difference to the earlier LSG model simulation results (e.g., Butzin et al., 2020) is that the MRAs of this study are considerably lower not only at the sea ice margins but also in fully ice-covered areas. As the sea ice extent and thickness in the LSG model were adjusted ad hoc, we think that its polar MRA values are spuriously high. This is also indicated by the rapid and counterintuitive decline of subsurface MRAs in ice-covered regions seen in LSG model simulations (cf. Fig. 3 in Butzin et al., 2017). As for the non-polar oceans, we find that CLIMBER-X, like the LSG model, underestimates MRAs at coastal sites and in upwelling regions. Multi-scale ocean modelling has shown that local dynamics can substantially modulate MRAs there (Lohmann et al., 2020). Such effects are only partially represented in simplified models which may explain some of the remaining discrepancies between simulated and observed patterns.

A brief inspection of B-Atm ^{14}C ages in the interior of the glacial oceans yields that our results are qualitatively in line with observations (Supplementary Fig. 3). Our simulation captures the pattern of deeper ocean ventilation in the North Atlantic than in the North Pacific, but the entire

Pacific below about 1.5 km depth is too young compared to observations. This may reflect the aforementioned low-MRA biases at the sea surface in high-latitude source water regions. However, the LSG model simulation for the LGM with considerably older polar-subpolar surface waters also fails to reproduce the high B-Atm ^{14}C ages reconstructed in the Pacific (not shown). Neither simulation accounts for the injection of ^{14}C -dead geologic carbon at the sea floor, which has been proposed to explain the anomalously old glacial – deglacial high ^{14}C ages observed at some locations in the Pacific (Rafter et al., 2019; Stott et al., 2019).

MRA evolution in simulations with prescribed cosmogenic ^{14}C production

In the previous simulations the MRA evolution was subject to prescribed $\Delta^{14}\text{C}$ variations in the atmosphere. These variations reflect changes in cosmogenic ^{14}C production (i.e., changes in the source strength of the global ^{14}C cycle) as well as climate-induced changes in the partitioning of ^{14}C between the atmosphere and oceans. To disentangle both effects, we consider two sensitivity experiments (named P_const and P_VADM) in which $\Delta^{14}\text{C}_A$ is not prescribed but forced with ^{14}C production rates. This approach ensures that the global ^{14}C budget is always closed which is not necessarily the case in $\Delta^{14}\text{C}_A$ -driven scenarios. In the following we neglect the terrestrial biosphere which represents a comparatively small ^{14}C sink that can be factored into the marine ^{14}C reservoir.

In the first sensitivity experiment P_const, we apply a constant ^{14}C production flux over the entire simulation period and chose the value that was diagnosed after the spin-up integration ($^{14}P_C^* = 1.72 \text{ atoms cm}^{-2} \text{ s}^{-1}$). This scenario reveals the effects of glacial to interglacial climate variations in CLIMBER-X on the MRA history (and on $\Delta^{14}\text{C}_A$, discussed in Section 3.3). In experiment P_const, global (50°N - 50°S) MRAs vary only in a narrow range of 600 - 700 ^{14}C years prior to 14 ka BP, before they almost monotonously decline to 470 ^{14}C years in 1850 CE (0.1 ka BP) at the end of the experiment (Fig. 1). The global MRA curve is much smoother than in simulation IntSH20 and only reflects short-term variations of the AMOC in a muted way (cf. Supplementary Fig. 1). It rather reflects the long-term trend of $p\text{CO}_2$, where increasing CO_2 levels lead to decreasing MRAs (see also Galbraith et al., 2015, for a theoretical treatment). The MRA- $p\text{CO}_2$ relationship appears to be decoupled during the last deglaciation as the simulated global MRA decline lags the prescribed rise of $p\text{CO}_2$ by about 4000 years (not shown). This is due to an AMOC shutdown in CLIMBER-X during Heinrich Stadial 1 (HS1, 15.0 - 17.5 ka BP, Supplementary Fig. 1), which leads to a temporary increase in $\Delta^{14}\text{C}_A$ that counterbalances the $p\text{CO}_2$ effect (Fig. 5). The absence of other pronounced global-scale MRA variations (simulated in experiment IntSH20) suggests that these are due to cosmogenic ^{14}C production variations. This is investigated in the second sensitivity experiment P_VADM.

Sensitivity experiment P_VADM considers transient cosmogenic ^{14}C production fluxes $^{14}P_C$ between 55 and 0.1 ka BP (shown in Fig. 5). The simulation branches from experiment P_const at 55 ka BP, analogously to scenario IntSH20 in which $\Delta^{14}\text{C}_A$ changes from constant to transient values at the same time. Simulation P_VADM yields enhanced millennial-scale variations of global

MRAs with a pronounced peak of almost 1200 ^{14}C years around 41 ka BP (Fig. 1). Moreover, it yields an earlier onset of the deglacial global MRA decline than sensitivity experiment P_const. Both traits coevolve with the $^{14}\text{P}_\text{C}$ record, i.e., with a production peak around 41 ka BP and a dip of $^{14}\text{P}_\text{C}$ between 18.8 ka BP and 16.5 ka BP. Overall, this experiment captures the characteristic features of the global MRA evolution resulting from the concentration-forced simulation IntSH20. Differences such as the temporal mismatch of MRA peak values and the lower short-term MRA variability reflect the uncertainties of the underlying atmospheric ^{14}C input functions ($\Delta^{14}\text{C}_\text{A}$ and $^{14}\text{P}_\text{C}$) and their different temporal variability. Taken together, these findings prove the significant influence of cosmogenic ^{14}C production variations onto the MRA history and that MRA excursions may reflect $^{14}\text{P}_\text{C}$ changes rather than ocean ventilation changes.

Coevolution of MRAs and atmospheric $\Delta^{14}\text{C}$

While the focus of this paper is on ^{14}C in the surface ocean, we finally exploit our production-forced sensitivity runs for a brief study of the coevolving $\Delta^{14}\text{C}$ in the atmosphere. In sensitivity experiment P_const, $\Delta^{14}\text{C}_\text{A}$ varies around 0 ‰. Prior to the deglaciation, the variability is probably too low because Heinrich Stadials HS2 – HS5 are not resolved by the PaleoMIST reconstruction which uses a time step of 2500 years. During the deglaciation $\Delta^{14}\text{C}_\text{A}$ initially rises to ~ 20 ‰, which is followed by a decline to ~ -50 ‰ by the beginning of the Holocene, before $\Delta^{14}\text{C}_\text{A}$ rises again to ~ -20 ‰ in 1850 CE (Fig. 5). This is due to the above mentioned abrupt variations in ocean ventilation simulated between 17.5 and 10 ka BP. The interhemispheric gradient of atmospheric $\Delta^{14}\text{C}$ (IHG) in this simulation amounts 4.7 ‰ in the long-term average and is constant (Fig. 6).

Sensitivity experiment P_VADM shows a sharp rise of $\Delta^{14}\text{C}_\text{A}$ between 42 and 40.5 ka BP, followed by an almost linear long-term decline (Fig. 5). The simulated peak value of about 650 ‰ is close to the reconstructed amplitude of ~ 700 to 800 ‰ during the Laschamps excursion (e.g., Cheng et al., 2018; Cooper et al., 2021) but the timing does not match. The model fails to reproduce the reconstructed rebound of $\Delta^{14}\text{C}_\text{A}$ between 33 and 25 ka BP and overestimates Holocene values (~ 50 ‰ instead of 0 ‰ in 1850 CE / 0.1 ka BP). The causes of these deviations are uncertain. A lower value of $^{14}\text{P}_\text{C}^*$ would improve the Holocene bias but further reduce $\Delta^{14}\text{C}_\text{A}$ before the deglaciation. Prior to the LGM, Heinrich events as well as Dansgaard-Oeschger events are not captured. Based on HS1, their contribution likely remained below 50 ‰, insufficient to explain the persistent $\Delta^{14}\text{C}_\text{A}$ values near 500 ‰ between 33 and 25 ka BP. Similar behaviour was found by Dinauer et al. (2020), despite testing various $^{14}\text{P}_\text{C}$ reconstructions and carbon cycle feedbacks.

Despite of these global mismatches, the simulated interhemispheric gradient (IHG) remains within the reconstructed long-term range of 4.5 - 7 ‰ (Hogg et al., 2020; Reimer et al., 2020). Around the Laschamps excursion our model produces an IHG of up to 13 ‰, almost twice as high as reconstructions suggest (Fig. 6). Whether this stronger gradient is realistic remains uncertain as the IHG reconstruction prior to ~ 13 ka BP relies on statistical modelling rather than on direct observations (Hogg et al., 2020). Our hypothesis can be tested once additional ^{14}C observations from that period are available.

Methods

Model and simulation setup

We apply the climate core of the Earth System Model of Intermediate Complexity CLIMBER-X (Willeit et al., 2022; 2023), version 3, which we have equipped with ^{14}C in terms of $\Delta^{14}\text{C}$. CLIMBER-X has been developed for long-term simulations and is about 100–1000 times faster than more comprehensive Earth System Models. As the model has been described in detail elsewhere, we will only provide a brief overview here. To summarize, the climate core of CLIMBER-X consists of the 2.5-dimensional statistical-dynamical atmosphere model SESAM, the 3-dimensional frictional-geostrophic ocean model GOLDSTEIN (Edwards and Marsh, 2005), the dynamic-thermodynamic sea ice model SISIM, and the land surface-vegetation model PALADYN (Willeit & Ganopolski, 2016). The horizontal model resolution is $5^\circ \times 5^\circ$. Vertical resolution in the ocean is 23 layers of uneven thickness using z-level coordinates. CLIMBER-X takes glacial-interglacial sea level variations into account and features a flexible land-sea mask. The model is forced with insolation (Laskar et al., 2004), greenhouse gas concentrations (Köhler et al., 2017), and continental ice sheets including their meltwater discharge. Here, we apply the PaleoMIST ice sheet reconstruction (Gowan et al., 2021; implemented into CLIMBER-X by Masoum et al., 2024a, 2024b) which, different to other ice sheet reconstructions, covers the past 80 kyr and thus the entire time span accessible to ^{14}C dating. In combination with orbital and greenhouse gas forcing, the PaleoMIST ice sheet forcing leads to an ocean cooling of up to 10 K at the sea surface during the Last Glacial Maximum (LGM) which is colder than in reconstructions (Paul et al., 2021; see Supplementary Fig. 2). Nonetheless, our simulations arrive at less sea ice in the Northeast Atlantic and in the Southern Ocean than the GLOMAP reconstruction for the LGM (Paul et al., 2021).

First, the model is integrated over 20,000 years with steady climate forcing for 1850 CE and $\Delta^{14}\text{C}_A = 0\text{‰}$ to achieve a defined non drifting initial state. Then CLIMBER-X is run from 70 ka BP forward in time with transient climate forcing, either with prescribed $\Delta^{14}\text{C}_A$ or $^{14}P_C$ values. Prior to 55 ka BP (the current limit of ^{14}C reconstructions) $\Delta^{14}\text{C}_A$ and $^{14}P_C$ are kept constant. In the concentration-driven simulation we apply a homogeneous $\Delta^{14}\text{C}_A$ value of 528 ‰ which corresponds to the global-mean value for 55 ka BP according to IntCal20 (Reimer et al., 2020). In the production-driven simulation we employ a homogeneous rate $^{14}P_C^*$ corresponding to a flux of $\sim 1.72 \text{ atoms s}^{-2}$, which would maintain $\Delta^{14}\text{C}_A$ at 0 ‰ in the simulation under preindustrial steady-state conditions. This value comes close to geomagnetic model estimates of the modern ^{14}C production (e.g., Panovska et al., 2023). From 55 ka BP onward we also apply transient values of $\Delta^{14}\text{C}_A$ and $^{14}P_C$, respectively. In the concentration-driven scenario we use $\Delta^{14}\text{C}_A$ according to IntCal20 (Reimer et al., 2020) for the Northern Hemisphere north of 30°N , $\Delta^{14}\text{C}_A$ according to SHCal20 (Hogg et al., 2020) for the Southern Hemisphere south of 30°S , and consider the average of both curves in the intertropical zone between 55 ka BP and 0.1 ka BP (1850 CE). Between 1850 CE and 2015 CE we apply $\Delta^{14}\text{C}_A$ values for the same three atmosphere zones reconstructed by

Graven et al. (2017). In the production-driven case we force the model with reconstructed ^{14}C production rates, the reconstruction of which is described further below.

Implementation of radiocarbon

Radiocarbon (^{14}C) is implemented in terms of $\Delta^{14}\text{C}$ as $^{14}R = 1 + \Delta^{14}\text{C} / (1000 \text{ ‰})$, which is the normalized $^{14}\text{C} / ^{12}\text{C}$ ratio corrected for carbon-isotopic fractionation. In the ocean we only consider dissolved inorganic carbon ($\text{DI}^{12}\text{C} \approx \text{DIC}$ and $^{14}R_{\text{DIC}}$) and neglect the organic ^{14}C cycle. We also assume that DIC concentrations are constant and homogeneous and that the sea surface is in local solubility equilibrium with respect to atmospheric CO_2 (see Mouchet, 2013, for further discussion). It has been shown that these approximations increase the relative uncertainty of $^{14}R_{\text{DIC}}$ by less than 5 % compared to more accurate approaches (Butzin et al., 2024).

The computation of the evolution of marine ^{14}C then simplifies to

$$(1) \quad \partial_t ^{14}R_{\text{DIC}} \approx \mathbf{D}(^{14}R_{\text{DIC}}) - \lambda ^{14}R_{\text{DIC}}$$

and the local air-sea exchange flux of $^{14}\text{CO}_2 / \text{CO}_2$ (in $\text{m}^{-2} \text{ s}^{-1}$) $^{14}F_A$ becomes

$$(2) \quad ^{14}F_A = k_w C_{\text{sat}} / C_{\text{DIC}_0} (^{14}R_A - ^{14}R_{\text{DIC}}).$$

In Equation (1), $\mathbf{D}$ is a nonlinear differential operator which describes the transport of $^{14}R_{\text{DIC}}$, and $\lambda = 1.216 \times 10^{-4} \text{ yr}^{-1}$ is the radioactive decay constant of ^{14}C (Audi et al., 2003; Bé and Chechev, 2012; Kutschera, 2013). In Equation (2) k_w and C_{sat} are the gas transfer velocity and the saturation concentration of CO_2 , $C_{\text{DIC}_0} = 2.0 \text{ mol m}^{-3}$ is the global-mean concentration of DIC in surface water, and $^{14}R_A$ is the $^{14}\text{CO}_2 / ^{12}\text{CO}_2$ ratio in the atmosphere. The gas transfer velocity (in m s^{-1}) is parameterized following Wanninkhof (2014):

$$(3) \quad k_w = 6.9722 \times 10^{-7} (Sc / 660)^{-0.5} u^2 (1 - f_i),$$

where Sc is the Schmidt number of CO_2 , u (in m s^{-1}) the wind speed at 10 m height, and f_i the sea ice area fraction (ranging from zero to one). The saturation concentration of CO_2 (in mol m^{-3}) is calculated as

$$(4) \quad C_{\text{sat}} = \Phi_0 x_{\text{CO}_2} P_{\text{SL}} / P_0,$$

where Φ_0 is the solubility function of CO_2 depending on temperature and salinity (Weiss and Price, 1980), x_{CO_2} is the mole fraction of atmospheric CO_2 , P_{SL} is the atmospheric in-situ pressure at sea level, and $P_0 = 1.01325 \times 10^5 \text{ Pa}$ is the atmospheric standard pressure.

Atmospheric radiocarbon values $^{14}R_A$ are either directly prescribed or computed from cosmogenic ^{14}C production rates $^{14}P_C$ using a simple atmospheric ^{14}C model. In both cases we consider three latitude zones (N: 90°N — 30°N , T: 30°N — 30°S , and S: 30°S — 90°S) with homogeneous atmospheric ^{14}C concentrations within each zone. In the production-driven model we allow for meridional ^{14}C exchange and assume that the meridional exchange rates μ across

the subtropical boundaries are identical and constant. The temporal evolution of atmospheric ^{14}C in the three latitude zones then becomes

$$(5) \quad {}^{14}\dot{R}_{A,N} = {}^{14}P_C - \lambda {}^{14}R_{A,N} - \mu ({}^{14}R_{A,N} - {}^{14}R_{A,T}) - 4 \iint {}^{14}F_A C_{DIC_0} dx_N dy_N / (x_{\text{CO}_2} M_A)$$

$$(6) \quad {}^{14}\dot{R}_{A,T} = {}^{14}P_C - \lambda {}^{14}R_{A,T} - \mu (2 {}^{14}R_{A,T} - {}^{14}R_{A,N} - {}^{14}R_{A,S}) - 2 \iint {}^{14}F_A C_{DIC_0} dx_T dy_T / (x_{\text{CO}_2} M_A)$$

$$(7) \quad {}^{14}\dot{R}_{A,S} = {}^{14}P_C - \lambda {}^{14}R_{A,S} - \mu ({}^{14}R_{A,S} - {}^{14}R_{A,T}) - 4 \iint {}^{14}F_A C_{DIC_0} dx_S dy_S / (x_{\text{CO}_2} M_A)$$

The terms at the far right side of Equations (5)–(7) describe the integrated air-sea exchange of $^{14}\text{CO}_2$ / $^{12}\text{CO}_2$ in a latitude zone, where $M_A = 1.7726 \times 10^{20}$ mol is the molecular inventory of the global dry atmosphere for a total mass of 5.1352×10^{18} kg (Trenberth and Smith, 2005) and a molar mass of dry air of $0.02897 \text{ kg mol}^{-1}$.

We chose a meridional exchange rate of $\mu = 1.667 \text{ yr}^{-1}$ which leads to an interhemispheric $\Delta^{14}\text{C}$ gradient (defined as $\Delta^{14}\text{C}_{A,N} - \Delta^{14}\text{C}_{A,S}$) of 4.6 ‰ during the Holocene in line with observations (Hogg et al., 2020; Reimer et al., 2020). This value of μ roughly corresponds to an inter-hemispheric exchange time of about 1.2 yr which is slightly lower than indicated by bomb ^{14}C pulse records (Levin et al., 2021) but still in the range of studies considering other transient tracers (e.g., Geller et al., 1997; Maiss et al., 1996).

We report radiocarbon ages ${}^{14}\tau$ in ^{14}C years (i.e., using the conventional half-life of 5568 years) according to

$$(8) \quad {}^{14}\tau = -8033 \ln ({}^{14}R_{DIC} / {}^{14}R_A).$$

Numerical ^{14}C age values in the text are rounded to full decades.

Reconstruction of ^{14}C production rate changes

To infer ^{14}C production rate changes, we combined the Virtual Axial Dipole Moments (VADM) of the models pfm9k1a (0 - 9 ka BP; Nilsson et al., 2024), GGF100k (9 - 14 ka BP; Panovska et al., 2019), and GGFSS70 (14 - 70 ka BP; Panovska et al., 2021) into one continuous record of geomagnetic field strength covering the last 70,000 years. Radiocarbon production rates were then calculated by converting VADM into cutoff-rigidities following Lifton et al. (2014) and using the Local Interstellar Spectrum by Herbst et al. (2017), a constant solar modulation potential of 500 MV, and the production yield functions by Poluianov et al. (2016). The resulting production rates have been normalized and converted to absolute ^{14}C production fluxes using the numerical value of ${}^{14}P_C^*$ ($\sim 1.72 \text{ atoms s}^{-2}$).

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Author contributions

M. B. implemented radiocarbon into the model, carried out the simulations, and wrote the
manuscript with contributions from all authors. F. A. reconstructed the radiocarbon production
rates. G. L. initiated and conceptualised the study. All authors analysed and discussed the results.

Competing interests

The authors declare that there is no conflict of interest.

Data availability

Model results are available at <https://nextcloud.awi.de/s/QmoqZojipQ3TWQ3>
but will be transferred to another public repository after the publication of the paper.

Code availability

The model code is available at <https://nextcloud.awi.de/s/tSSA2LaNY6NGFB5>
and will be transferred to another public repository after the publication of the paper.

Figures

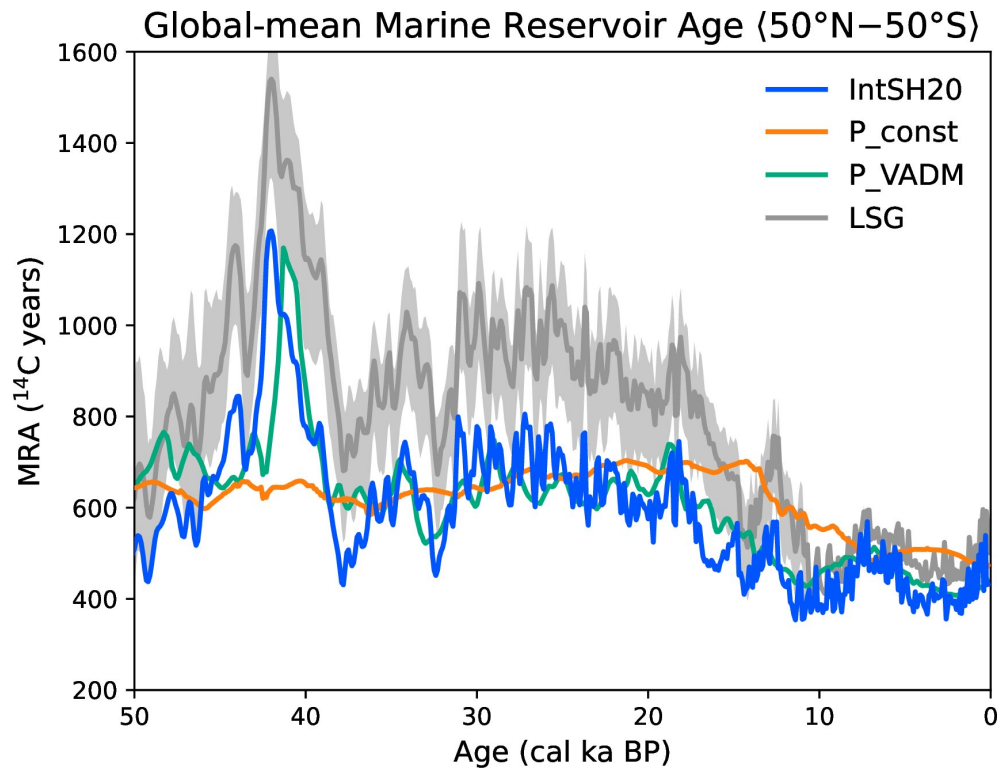


Figure 1. Long-term evolution of simulated Marine Reservoir Ages (MRA). Shown are model results globally averaged between 50°N and 50°S. The blue curve (scenario IntSH20) is obtained with atmospheric $\Delta^{14}\text{C}$ according to IntCal20 (Reimer et al., 2020) combined with SHCal20 (Hogg et al., 2020). Orange (P_const) and green (P_VADM) curves are results for constant and transient rates of cosmogenic ^{14}C production, respectively. Previous simulation results obtained with the LSG model using IntCal20 that were carried out for Marine20 (Heaton et al., 2020) are shown in gray, with their uncertainty range estimated by alternative ocean climate scenarios.

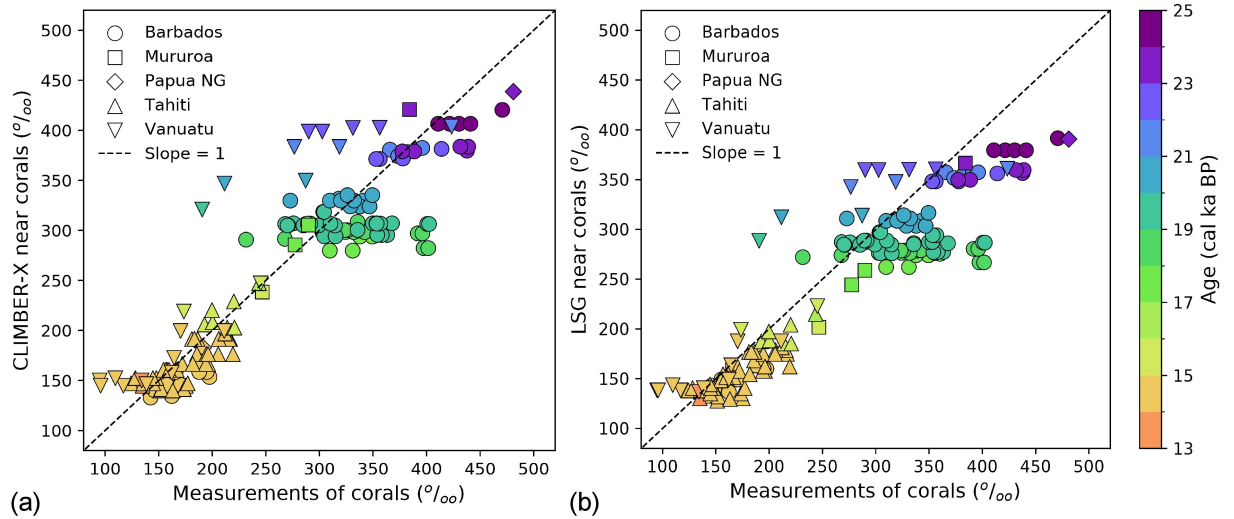


Figure 2. Comparison of observed and simulated $\Delta^{14}\text{C}$ in surface water in the past. Scatter plots comparing marine $\Delta^{14}\text{C}$ of the past (13 - 25 cal ka BP) measured in corals contributing to IntCal20 (Bard et al., 1990, 1998, 2004; Cutler et al., 2004; Durand et al., 2013; Fairbanks et al., 2005) with model results. (a): CLIMBER-X (this study). (b): Previous results obtained with the LSG model for Marine20 (Heaton et al., 2020).

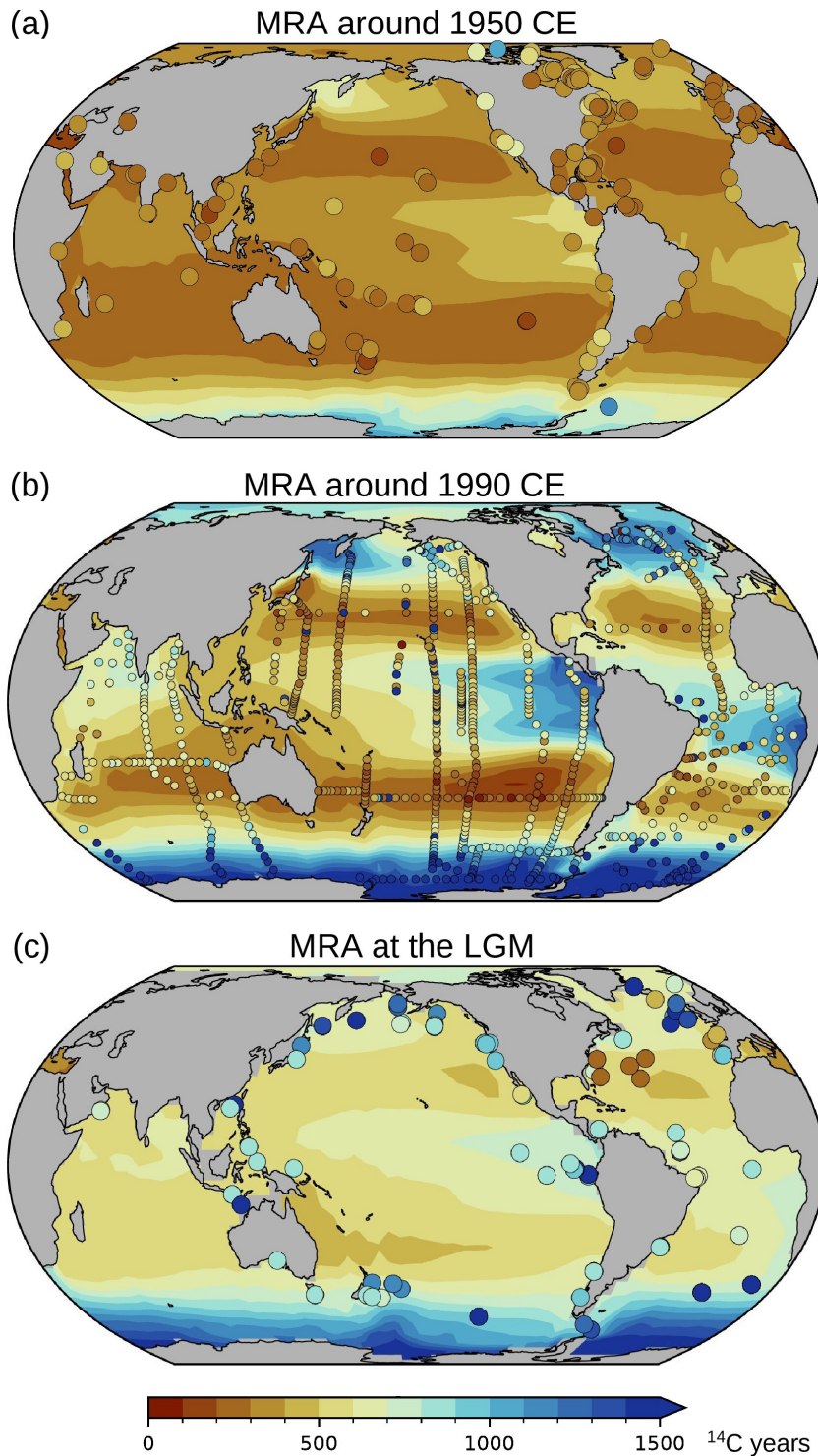


Figure 3. Comparison of observed and simulated marine reservoir ages (MRA). Shaded areas are model results (this study). (a): MRA around 1950 CE, triangles are observations compiled by Rubin and Key (2002; see further references therein) and dots are observations between 1945 and 1954 CE collected in the Marine20 database (Reimer and Reimer, 2001; cf. <https://calib.org/marine> for further references). (b): MRA around 1990 CE, dots are observations between 1985 CE and 1995 CE (compiled by Lauvset et al., 2023; data retrieved using webODV Explore, Schlitzer and Mieruch-Schnülle, 2025) (c): MRA at the Last Glacial Maximum, model results are averaged over 19.0 - 21.8 ka BP and dots are observations between 19.0 and 21.8 ka BP (compiled by Skinner et al., 2023; see further references therein).

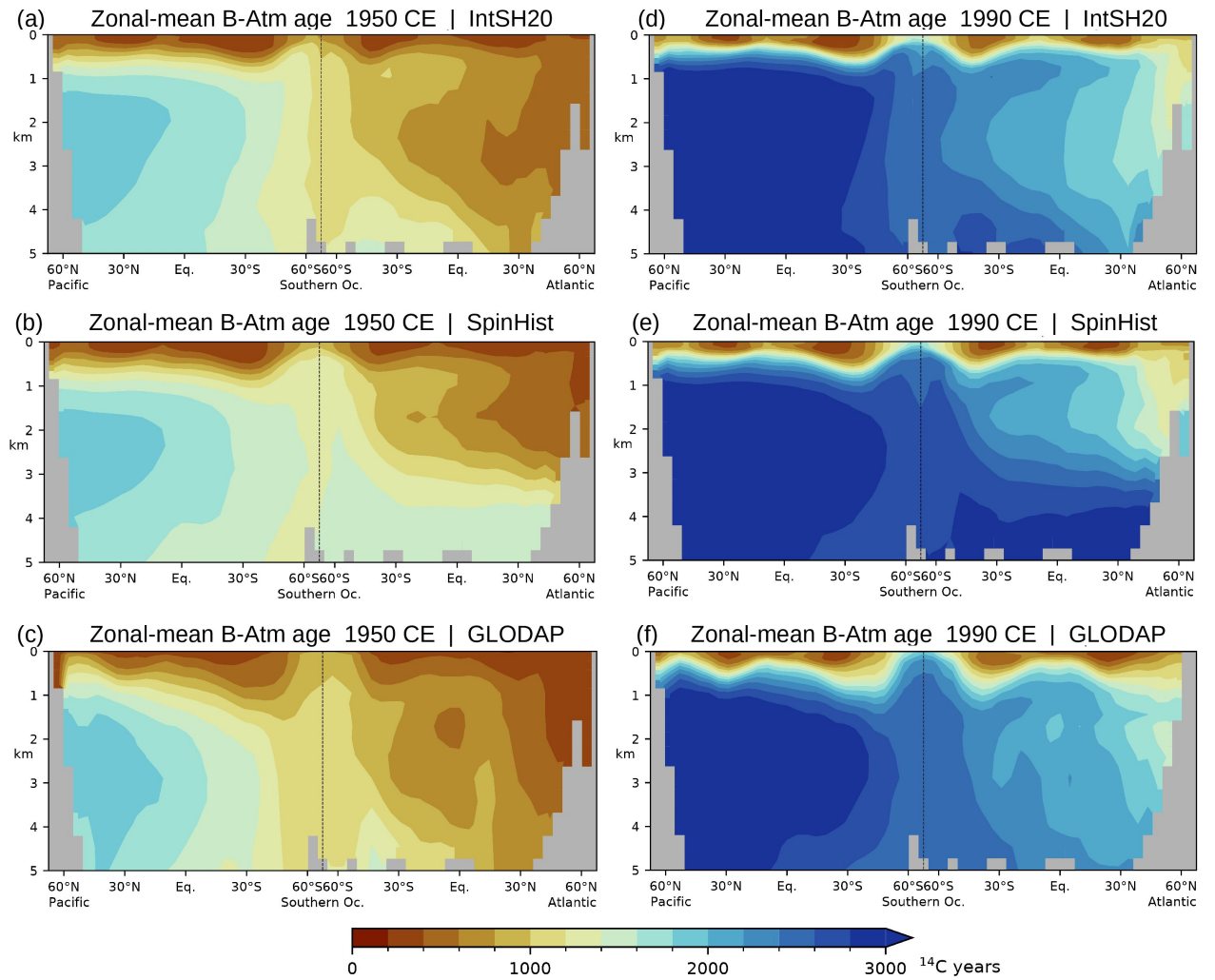
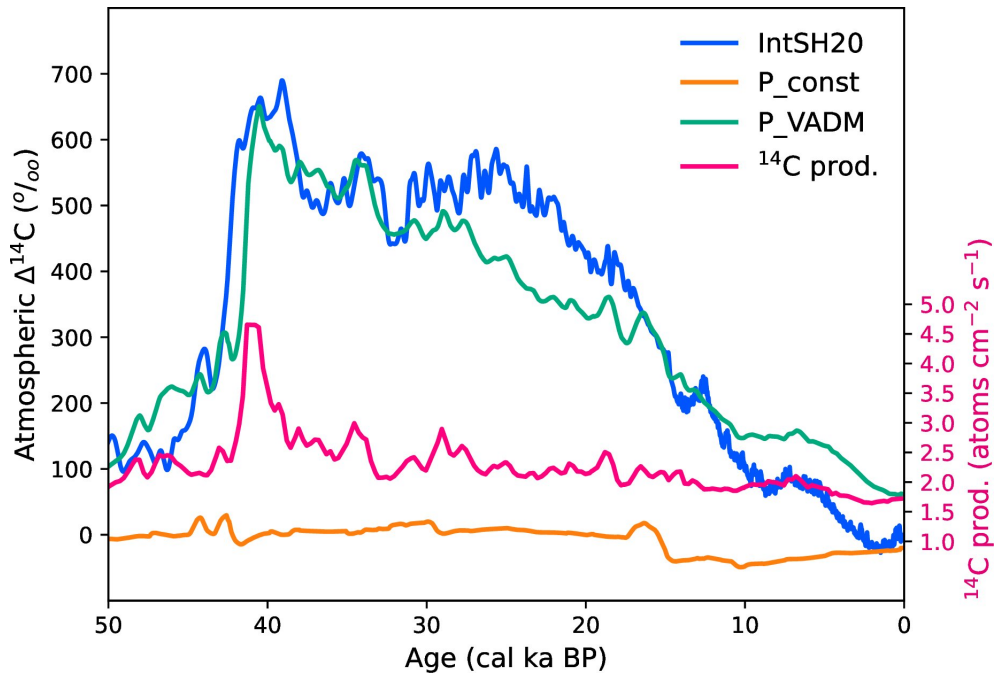


Figure 4. Comparison of simulated and observed benthic-atmospheric (B-Atm) ^{14}C ages. Shown are zonal-mean values in the Atlantic and Pacific. (a): Long-term simulation results for 1950 CE (this study, extension of simulation IntSH20). (b): Short-term simulation results for 1950 CE (this study, extension of the spin-up equilibrium simulation). (c): Reconstruction of pre-nuclear values (Key et al., 2004). (d): Long-term simulation results for 1990 CE (this study, extension of simulation IntSH20). (e): Short-term simulation results for 1990 CE (this study, extension of the spin-up equilibrium simulation). (f) Gridded observations around 1990 CE (Key et al., 2004). Gridded observations were remapped to the spatial resolution of CLIMBER-X.



584

585 **Figure 5. Atmospheric $\Delta^{14}\text{C}$ according to simulations and reconstructions.** The blue curve
 586 (IntSH20) is the average of the IntCal20 and SHCal20 reconstructions. The orange curve (P_const)
 587 shows $\Delta^{14}\text{C}_A$ simulated for a constant cosmogenic ^{14}C production flux of $^{14}P_C^* = 1.72 \text{ atoms cm}^{-2}$
 588 s^{-1} that would keep the global ^{14}C inventory constant under preindustrial steady-state climate
 589 conditions. The green curve (P_VADM) is obtained if ^{14}C in CLIMBER-X is forced with
 590 reconstructed ^{14}C production rates of the past $^{14}P_C$, which are shown in red.

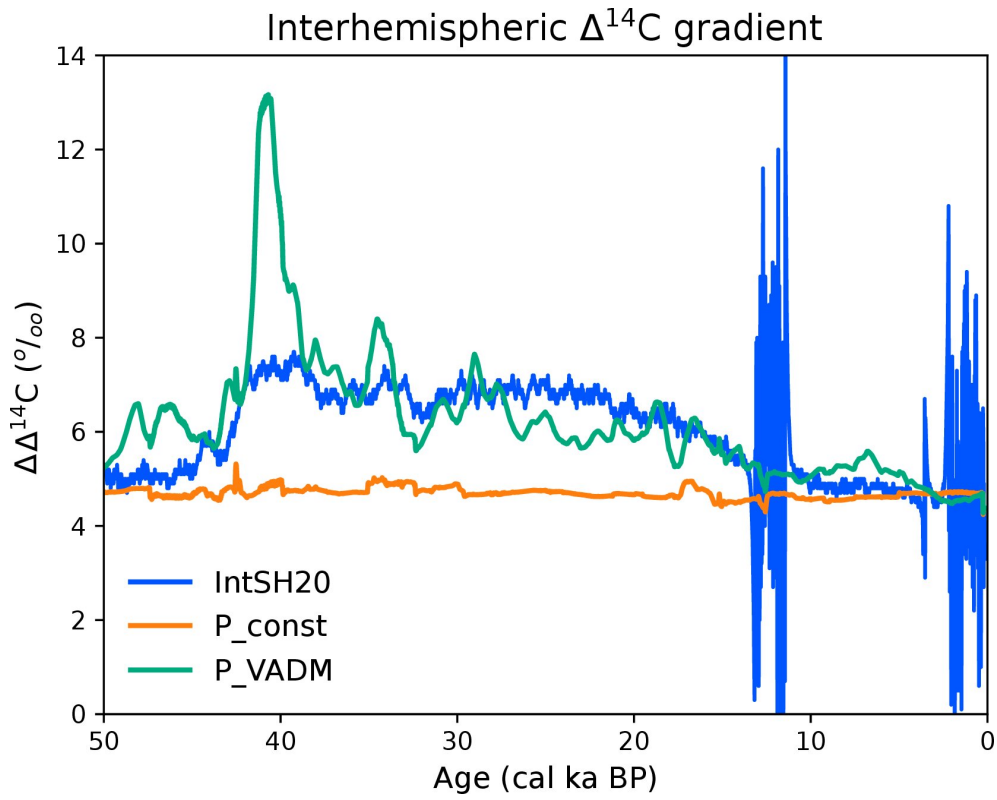


Figure 6. Interhemispheric gradient of $\Delta^{14}\text{C}$ in the atmosphere. The blue curve displays the difference between IntCal20 (Reimer et al., 2020) and SHCal20 (Hogg et al., 2020) values. Orange and green curves result from simulations forced with constant and transient ^{14}C production rates (sensitivity experiments P_const and P_VADM).

Comments on the manuscript by Butzin et al.

"Marine radiocarbon reservoir age simulations for the past 50,000 years revisited" (COMMSENV-25-5679)

In this work the authors study the evolution of Marine Reservoir Age (MRA) over the past fifty thousand years with the help of an Earth system model of intermediate complexity.

There are two distinct parts in their work.

First they constrain the atmospheric $\Delta^{14}\text{C}$ to follow the Intcal20 and SHcal20 reconstructions. Under that scenario, they compare the MRAs with those previously obtained with an ocean global circulation model under different circulation scenarios (hence, the mention 'revisited' in the title of the paper). In this first part they also examine the impact of initial conditions on the final radiocarbon distribution in the ocean.

In the second part, the model is forced with cosmogenic radiocarbon production in the atmosphere. Two sensitivity experiments were performed, one with a reconstructed rate, and one with a constant rate, allowing for disentangling the influence of ocean circulation changes and changes in the production rates on MRA. A very short discussion on the interhemispheric $\Delta^{14}\text{C}$ gradient then follows. The authors then conclude that the interhemispheric $\Delta^{14}\text{C}$ gradient was much larger during the Laschamps event (~ 41 ka BP) than predicted by the reconstructions.

The manuscript is well structured and clear to read. However, there are several points of concern.

I am not convinced by the explanation of the difference between the MRA predicted by the LSG and by CLIMBER-X (lines 95-99). Given that the air-sea exchange of radiocarbon is very slow, a more vigorous ocean ventilation would result in an increase of the difference between the radiocarbon content in the surface layer and in the atmosphere, since the faster renewal of surface waters would allow for less time to equilibrate.

I suspect that the larger MRA obtained with the LSG are due to a more vigorous vertical mixing in that model. Did the authors examine the difference in background vertical mixing (explicit and numerical) and mixed layer depth between the LSG and CLIMBER-X?

The transient climate simulation should be discussed further. Over the deglaciation, the results presented here differ in several aspects from previously published ones under the same ice-sheet scenario (Masoum et al., 2024). For example, in this study, the changes in AMOC at the end of Heinrich 1 event are much larger than obtained by Masoum et al. (2024) in the PaleoMist scenario.

How well is the present-day ocean state represented at the end of the long transient simulation? I agree that $\Delta^{14}\text{C}$ in 1990 is reasonably reproduced (Fig. 4), but how do temperature and salinity distributions compare to observations?

The differences observed in 1950 and 1990 between experiments IntSH20 and SpinHist (Fig. 4) are not solely explained by the longer or shorter transient ^{14}C input history as claimed on lines 120-122. In 1950 or 1990, the ocean circulation obtained with IntSH20 is most probably very different from that obtained with SpinHist. Indeed, in IntSH20, the ocean has been adjusting to a varying climate forcing over 70 kyr in addition to the transient ^{14}C input. The SpinHist experiment which starts in 1850 from a long spin-up with steady climate forcing is not influenced by the long term climate trend. Differences in the radiocarbon distribution between IntSH20 and SpinHist also result from the different climate scenarios and are not only due to the duration of the transient atmospheric radiocarbon forcing.

The reconstruction of the radiocarbon production rate, which is central to this study, is not sufficiently discussed. What is the sensitivity of the production rate to the various parametrizations?

How does the reconstruction of ^{14}C production in this study compare to other reconstructions?

It is not clear whether the ^{14}C production rate at a given time is homogeneous over the globe or varies with location. Are the $^{14}\text{P}_\text{C}$ rates in the three latitude bands and in Eqs (5) to (7) identical? I suppose they are not. Could the authors illustrate the temporal evolution of $^{14}\text{P}_\text{C}$ in the three latitude bands?

Equations (5) to (7) are erroneous. It seems there is a confusion between terms driving the radiocarbon ratio (^{14}R) and terms controlling the radiocarbon concentration (^{14}C) in the atmosphere.

Line 39 : that we carried out... (suppress the 'had').

Figure 1 & line 84: With respect to the timing of Laschamps event, it would help the reader to superimpose a plot of the reconstructed $\Delta^{14}\text{C}_\text{A}$ in Figure 1.

Line 77: Could the authors document the difference between the atmospheric $\Delta^{14}\text{C}$ reconstructions used with CLIMBER-X and with LSG?

Line 97: Since the envelope of MRA obtained with the LSG results from the combination of three scenarios (PD, GS, and CS), it is not only the AMOC strength in scenario GS that should be considered here.

Line 164: there is no Section 3,3.

Line 170: Galbraith et al. (2015) is not the right reference here.

Lines 172 & 174: adding the evolution of atmospheric pCO_2 in Figs 1 and 5 would really help the reader.

Line 212: Dinaiuer et al. (2020) is missing in the reference list.

Line 240: Nonetheless, our simulations *predict* less sea ice... (instead of 'arrive at').

Lines 249 & 318: The production rate should be in $\text{atoms cm}^{-2} \text{s}^{-1}$.

Line 269: The units of $^{14}\text{F}_\text{A}$ as given by Eq. (2) are m s^{-1} , not $\text{m}^{-2} \text{s}^{-1}$.

Line 305: The age that is reported by Eq. (8) is the Marine Radiocarbon Age (MRA); hence, line 305 should read: 'We report *marine* radiocarbon ages...'.
Figure 3a: the caption mentions that the observations by Rubin and Key are identified by means of triangles. However, I do not identify any triangle in Fig. 3a.

Figure S1, caption: the range for H1 is 15-17.5, not 15-7.5.

References

Masoum, A., Nerger, L., Willeit, M., Ganopolski, A., Lohmann, G. (2024). Paleoclimate data assimilation with CLIMBER-X: An ensemble Kalman filter for the last deglaciation. PLOS ONE 19, e0300138. <https://doi.org/10.1371/journal.pone.0300138>