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MARINE20—THE MARINE RADIOCARBON AGE CALIBRATION CURVE (0–55,000 CAL BP)

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ABSTRACT. The concentration of radiocarbon (¹⁴C) differs between ocean and atmosphere. Radiocarbon determinations from samples which obtained their ¹⁴C in the marine environment therefore need a marine-specific calibration curve and cannot be calibrated directly against the atmospheric-based IntCal20 curve. This paper presents Marine20, an update to the internationally agreed marine radiocarbon age calibration curve that provides a non-polar global-average marine record of radiocarbon from 0–55 cal kBP and serves as a baseline for regional oceanic variation. Marine20 is intended for calibration of marine radiocarbon samples from non-polar regions; it is not suitable for calibration in polar regions where variability in sea ice extent, ocean upwelling and air-sea gas exchange may have caused larger changes to concentrations of marine radiocarbon. The Marine20 curve is based upon 500 simulations with an ocean/atmosphere/biosphere box-model of the global carbon cycle that has been forced by posterior realizations of our Northern Hemispheric atmospheric IntCal20 ¹⁴C curve and reconstructed changes in CO₂ obtained from ice core data. These forcings enable us to incorporate carbon cycle dynamics and temporal changes in the atmospheric ¹⁴C level. The box-model simulations of the global-average marine radiocarbon reservoir age are similar to those of a more complex three-dimensional ocean general circulation model. However, simplicity and speed of the box model allow us to use a Monte Carlo approach to rigorously propagate the uncertainty in both the historic concentration of atmospheric ¹⁴C and other key parameters of the carbon cycle through to our final Marine20 calibration curve. This robust propagation of uncertainty is fundamental to providing reliable precision for the radiocarbon age calibration of marine based samples. We make a first step towards deconvolving the contributions of different processes to the total uncertainty; discuss the main differences of Marine20 from the previous age calibration curve Marine13; and identify the limitations of our

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approach together with key areas for further work. The updated values for ΔR , the regional marine radiocarbon reservoir age corrections required to calibrate against Marine20, can be found at the data base <http://calib.org/marine/>.

KEYWORDS: Bayesian modeling, calibration, carbon cycle, computer model, marine environment.

1. INTRODUCTION

Marine Specific Calibration

In order to convert the radiocarbon (^{14}C) date of a sample into a calendar age we need to perform calibration. This depends upon creating an accurate estimate of ^{14}C concentration over time for the specific local environment in which the sample was exchanging carbon. From this, we can generate a calibration curve which provides estimates of the radiocarbon date at any particular calendar age. Comparison of the radiocarbon date for a sample of interest against this calibration curve then enables us to infer the potential calendar ages at which it stopped exchanging with its local environment.

The reservoirs of ^{14}C within surface-ocean environments are systematically depleted compared to the atmosphere, and with higher frequency components in the variation predominantly damped. As a consequence, one cannot use the atmospheric calibration curve for the calibration of marine radiocarbon dates. The reasons for the depletion, and associated damping, are primarily the time it takes for atmospheric ^{14}C to exchange with the surface ocean and that the ocean interior stores large amounts of old carbon which slowly circulates up to the surface. To further complicate marine calibration, the level of depletion varies according to both location and time due to spatio-temporal differences in ocean mixing, various factors affecting the rates of air-sea exchange and wider changes to the carbon cycle, notably the atmospheric CO_2 mixing ratio (Bard 1988). To calibrate marine radiocarbon samples accurately, this varying and localised depletion must be accounted for.

The level of surface-ocean ^{14}C depletion, for a particular point in calendar time and a particular location, is typically expressed in terms of the marine radiocarbon reservoir age (hereafter **Marine Reservoir Age or MRA**)¹. We denote the MRA, for a specified location and calendar age θ cal kBP, by $R^{\text{Location}}(\theta)$. The MRA defines the difference, at calendar age θ cal kBP, between $h^{\text{MarineLocation}}(\theta)$, the radiocarbon age of dissolved inorganic carbon (DIC) in the mixed ocean surface layer at that location, and $h^{\text{NHAtmosphere}}(\theta)$, the radiocarbon age of CO_2 in the Northern Hemispheric (NH) atmosphere, i.e.

$$h^{\text{MarineLocation}}(\theta) = R^{\text{Location}}(\theta) + h^{\text{NHAtmosphere}}(\theta).$$

Note that $h^{\text{NHAtmosphere}}(\theta)$ is the function we estimate as the IntCal20 curve (Reimer et al. 2020 in this issue).

Present day MRA values range from about 400 ^{14}C yr in subtropical oceans to over 1000 ^{14}C yr near the poles (Bard 1988; Reimer and Reimer 2001; Key et al. 2004). To describe the variation in MRA between regions, we generally consider how the MRA at a particular location varies from $R^{\text{GlobalAv}}(\theta)$, the globally averaged mixed-layer reservoir age. These regional corrections of the MRA are reported as $\Delta R(\theta)$ values, i.e.

¹Some authors (e.g. Skinner et al. 2019) refer to the MRA as an R-age.

$$R^{Location}(\theta) = R^{GlobalAv}(\theta) + \Delta R^{Location}(\theta).$$

Pre-bomb estimates for $\Delta R(\theta)$ covering a wide range of locations are provided by Reimer and Reimer (2001) via the maintained [online database <http://calib.org/marine/>](http://calib.org/marine/).

How to Calibrate a Marine Radiocarbon Sample

In order to calibrate a radiocarbon age X_i from a marine sample, one needs to calibrate against the suitable calibration curve for that particular location, i.e. $h^{MarineLocation}(\theta)$. While in principle a location-specific marine calibration curve could be found by adding the MRA for that location to the NH atmospheric calibration curve, without highly detailed time-varying regional MRA estimates this does not account for the attenuation in the level of ^{14}C in the surface ocean. The more accepted approach is instead to use the globally averaged marine calibration curve, which we call Marine20,

$$h^{MarineAv}(\theta) = R^{GlobalAv}(\theta) + h^{NHAtmosphere}(\theta),$$

with an adjustment for the regional $\Delta R(\theta)$. This $h^{MarineAv}(\theta)$ provides the globally averaged radiocarbon age of DIC in the mixed surface layer.

While changes over time in $R^{GlobalAv}(\theta)$, the globally averaged MRA, are expected to be significant, temporal changes in the regional corrections $\Delta R(\theta)$ are generally believed to be smaller. Consequently, we typically make a simplifying assumption that, for any individual location, $\Delta R(\theta)$ is constant over time, i.e. $\Delta R^{Location}(\theta) \equiv \Delta R^{Location}$. Temporal changes in MRA are restricted to the global-average $R^{GlobalAv}(\theta)$ and automatically incorporated into $h^{MarineAv}(\theta)$, the Marine20 curve.

Specifically, suppose one wishes to calibrate a marine ^{14}C determination X_i with a laboratory-reported radiocarbon age uncertainty of σ_i . Using the appropriate regional values found at <http://calib.org/marine/>, it is first required to subtract the appropriate regional $\Delta R^{Location}$ correction from X_i and adjust, in quadrature, the radiocarbon age uncertainty,

$$X_i^{Adj} = X_i - \Delta R^{Location},$$

$$\sigma_i^{Adj} = \sqrt{\sigma_i^2 + \tau_{\Delta R}^2}.$$

Here $\tau_{\Delta R}$ is the standard deviation, also provided by the calib.org database, corresponding to the uncertainty in the regional $\Delta R^{Location}$ correction. The user then calibrates this region-adjusted value X_i^{Adj} , using the adjusted uncertainty σ_i^{Adj} , against the global-average Marine20 curve $h^{MarineAv}(\theta)$. Note however that several calibration software packages perform these adjustments, and subsequent calibration, automatically if given the appropriate regional $\Delta R^{Location}$ and $\tau_{\Delta R}$ values. The specific formatting requirements of their particular software should therefore be checked by any user before performing calibration of marine samples.

The Marine20 Calibration Curve

In this paper, we provide an estimate for $\Delta_{MarineAv}^{14}\text{C}(\theta)$, the “globally averaged” mixed-layer ^{14}C concentration from 0–55 cal kBP; and hence $h^{MarineAv}(\theta)$, the corresponding global-average mixed-layer marine calibration curve Marine20. Simultaneously, we provide the

corresponding estimate of $R^{GlobalAv}(\theta)$, the globally averaged mixed-layer MRA, which is of scientific interest in its own right. For this work, we define the “global-average” as being over the non-polar ocean, which lies between 40°S and either 50°N (Atlantic) or 40°N (Pacific), since local sea ice might significantly affect MRAs in higher latitudes (Butzin et al. 2017).

Our Marine20 curve is constructed using the BICYCLE global carbon cycle model, a Box-model of the Isotopic Carbon cYCLE (Köhler and Fischer 2004, 2006; Köhler et al. 2005, 2006). This model incorporates a globally averaged atmospheric box and modules of the terrestrial (7 boxes) and oceanic (10 boxes) components of the carbon cycle, including deep ocean-sediment fluxes of DIC and alkalinity that mimic the process of carbonate compensation. It is driven by temporal changes in the boundary conditions (changing climate) and simulates changes in the carbon cycle, including ^{13}C and ^{14}C . Here, we revise BICYCLE to allow the atmospheric CO_2 and $\Delta^{14}\text{C}$ to be specified externally by an ice-core based reconstruction and the IntCal20 curve respectively, rather than using the values that BICYCLE calculates internally. In overriding BICYCLE’s internally calculated CO_2 and $\Delta^{14}\text{C}$ with these two external, independently obtained, reconstructions arising from observed data, we aim to correct for any potential model limitations and ensure our simulated changes in the carbon cycle remain as close as possible to the observed ice core and ^{14}C data.

We take a Monte-Carlo approach whereby we generate an ensemble of 500 BICYCLE simulations, each driven by a different potential reconstruction of atmospheric $\Delta^{14}\text{C}$ and CO_2 as well as other key carbon cycle parameters. Our CO_2 reconstructions are based upon a stack of multiple ice core records (Köhler et al. 2017), while our estimates of atmospheric $\Delta^{14}\text{C}$ from 0–55 cal kBP are taken from individual posterior realizations of the IntCal20 curve (Reimer et al. 2020 in this issue). By considering the variability in the resultant ensemble of BICYCLE simulations, we are able to propagate uncertainty about past carbon cycle dynamics through to our final Marine20 curve. Figure 1 illustrates the various steps, data sets and methods necessary for the construction of both IntCal20 and Marine20.

The proposed modeling approach offers a significant improvement over previous marine calibration curves, e.g. Marine04 (Hughen et al. 2004), Marine09 (Reimer et al. 2009) and Marine13 (Reimer et al. 2013). In these previous versions, the curves were generated in distinct sections and spliced together. For Marine13, the curve from 0–10.5 cal kBP was created using a time-invariant ocean-atmosphere box diffusion model (Oeschger et al. 1975) driven by the corresponding atmospheric IntCal13 pointwise means and a range of piston velocities and eddy diffusivities as described in Hughen et al. (2004). From 10.5–13.9 cal kBP, it was constructed based upon marine observations from a range of locations. Finally, beyond 13.9 cal kBP, Marine13 was identical to the atmospheric IntCal13 curve with a constant MRA offset of 405 ^{14}C yr.

Conversely, for Marine20, the use of BICYCLE and of time-dependent forcings permit more complex and realistic inclusion of key carbon cycle changes extending back to 55 cal kBP. Additionally, improvements in the Bayesian modeling of IntCal20 (Heaton et al. 2020 in this issue) allow us for the first time to rigorously incorporate the time-dependent uncertainty of atmospheric $\Delta^{14}\text{C}$ into the marine age calibration curve through our Monte-Carlo approach. Together, these features enable more accurate estimation of both the

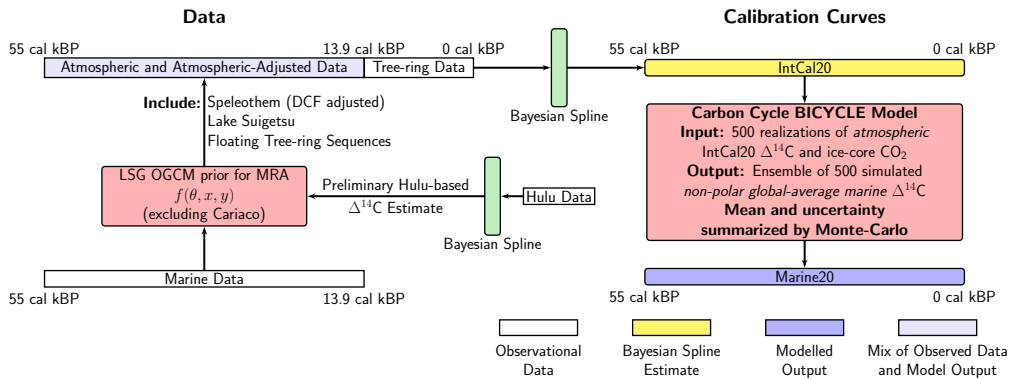


Figure 1 Schematic diagram of IntCal20 and Marine20 age calibration curve construction. IntCal20 is based solely upon tree-ring measurements from 0 to ca. 13.9 cal kBP. Beyond this a variety of data are used, including marine records which require an initial transformation to an atmospheric equivalent. To achieve this, a preliminary $\Delta^{14}\text{C}$ history is estimated based upon Hulu Cave speleothems (Southon et al. 2012; Cheng et al. 2018). This preliminary $\Delta^{14}\text{C}$ curve is used to drive the LSG OGCM and provide prior, first-order, estimates of regional MRAs (Butzin et al. 2020 in this issue) for each IntCal20 marine dataset, with the exception of the Cariaco Basin which is adjusted adaptively during curve construction (Hughen and Heaton 2020 in this issue). The adjusted marine data are then combined with speleothems, macrofossils from Lake Suigetsu, and floating tree-ring sequences to constitute a mixed, atmospheric and atmospheric-adjusted, dataset extending from ca. 13.9–55 cal kBP. A Bayesian spline is jointly fitted to both the tree-ring samples (from 0–13.9 cal kBP) and this mixed data (beyond 13.9 cal kBP) to create the IntCal20 curve (Heaton et al. 2020 in this issue). To generate Marine20, 500 posterior atmospheric $\Delta^{14}\text{C}$ realizations are taken from the IntCal20 Bayesian spline and propagated through the BICYCLE carbon cycle model alongside reconstructions of atmospheric CO_2 obtained from ice core records. The ensemble of 500 simulated outputs are then summarized to create the Marine20 curve.

global marine calibration curve and the globally averaged MRA, especially beyond the Holocene. We show that the results BICYCLE provides for individual runs are almost indistinguishable from the three dimensional Large Scale Geostrophic Ocean General Circulation Model (LSG OGCM) (Butzin et al. 2017, 2020 in this issue) and that BICYCLE's model output agrees well with pre-bomb data based on seawater (Bard 1988) and marine shells (Reimer and Reimer 2001; Toggweiler et al. 2019).

While more complex ocean models such as the LSG OGCM exist, we deliberately retain the simpler BICYCLE box model since it is more appropriate for our specific, global-average, calibration needs. For practical radiocarbon calibration, we require not only a curve consisting of a single point estimate for the global marine $\Delta^{14}\text{C}$ at any calendar age but also a reliable measure of the uncertainty on that estimate. The use of the computationally fast BICYCLE box model allows us to create a large ensemble of simulations and hence, via Monte-Carlo, rigorously quantify the uncertainty in the modeled global-average marine $\Delta^{14}\text{C}$ level which is a necessary consequence of the uncertainty in our various model/climate inputs.

The paper is set out as follows. In Section 2 we describe the requirements for a computer model in order to be suitable for use in construction of a radiocarbon age calibration curve; how uncertainty in the various driving inputs can be propagated through such a computer model; and provide a short tutorial on how the resultant output uncertainty can be quantified using Monte-Carlo techniques as well as why these Monte-Carlo techniques are needed. In Section 3, we justify our recommended approach to marine calibration of using local ΔR adjustments, based upon observations, to a global-average curve rather than

directly using regional output from more complex three-dimensional models. In Section 4 we present a short summary of BICYCLE and describe in detail the key processes/forcings and their uncertainties which we propagate through to our final calibration curve. We generate our ensemble of 500 BICYCLE simulations across the plausible ranges of its key inputs and present the resultant Marine20 curve in Section 5. Here we also discuss differences with the previous Marine13 curve. In particular, we note that Marine20 estimates a significant, and consistent, increase in the globally averaged MRA when compared to Marine13. With a sensitivity analysis we also document how much the different processes contribute to the overall uncertainty in Marine20. Section 6 provides a comparison against simulations with the more complex LSG OGCM and against both current day (pre-bomb) global marine data and older marine data from the IntCal20 database. Here we show that the increase, when compared to Marine13, in the global-average MRA in Marine20 is in agreement with average pre-bomb observations and the LSG OGCM. Finally, in Section 7 we discuss the limitations of our approach alongside areas needed for future work and development. For those seeking further details on the model output, including the final Marine20 radiocarbon age calibration curve, the spatially resolved regional open-ocean MRA estimates from the LSG OGCM, as well as the 500 NH atmospheric $\Delta^{14}\text{C}$ realizations used as inputs for our Monte-Carlo ensemble, see PANGAEA (<https://doi.org/10.1594/PANGAEA.914500>).

For a calibration end-user, it is key to note that the update from Marine13 to Marine20 is accompanied by updates to the regional ΔR corrections which can be found at <http://calib.org/marine/>. This is particularly critical since Marine20 estimates a significantly increased globally averaged MRA. Calibration of local reservoirs using values of ΔR based on Marine13 will give incorrect calendar age estimates.

Calibration in Polar Regions

Our Marine20 curve is intended for calibration of marine ^{14}C samples arising from non-polar locations, nominally/approximately ranging from 40°S–40/50°N. In higher-latitude polar regions, MRAs and critically their possible fluctuations over time are expected to be larger due to significantly increased variability in ocean ventilation and air-sea gas exchange mostly arising from changes in sea ice extent and differences in wind strength (Butzin et al. 2005). This is particularly likely to be the case during glacial periods. LSG OGCM estimates for the MRA in a particular calendar year within polar regions can vary by over 1000 ^{14}C yr between a simulation assuming a climate scenario suitable for the present day (its PD scenario which assumes little sea ice, Butzin et al. 2005, 2020 in this issue) and runs in climate scenarios appropriate for more extreme glacial periods (e.g. its GS/CS scenarios where the extent of sea ice is much greater in addition to significant changes to ocean ventilation and wind stress). This difference indicates that uncertainty in past polar climate conditions can have a significant effect on calibration in these regions. Consequently, for those polar regions affected by such additional variability, an assumption of a constant ΔR adjustment from our equatorial Marine20 global-average is not suitable. The current Marine20 curve is not therefore suitable for calibration in these polar regions.

Current proxy records for such ocean ventilation, sea ice extent and wind strength are not sufficient to reliably reconstruct these variables to permit accurate and precise modeling of polar calibration curves. Construction of such polar curves would be a valuable area for future study which we discuss in more detail within Section 7. For current users wishing to

calibrate marine ^{14}C samples from polar locations where variability in sea ice and ocean circulation may have significantly affected MRAs, we direct them to the regional LSG OGCM output as discussed in Section 3 and available on PANGAEA. Such users should however be aware that each individual LSG OGCM simulation only considers a single climate scenario. Due to the very considerable uncertainty on the appropriate historic polar climatic conditions, and since the impact of different climate scenarios on MRA in these regions is very large, resultant polar calibration age based upon such fixed scenarios should be treated with extreme caution. The calibrated age estimates obtained under any single fixed scenario are likely to be considerably over-precise.

2. THE MONTE-CARLO PRINCIPLE OF MARINE20 CURVE CONSTRUCTION

A Computer Model

The use of models for the study and prediction of MRA extends back to Craig (1957), Arnold (1957), Revelle and Suess (1957) and was updated later by Oeschger et al. (1975) with the introduction of a diffusion coefficient to simulate mixing in the deep ocean box (eddy diffusivity). This so-called box-diffusion model has been fundamental to the construction of the marine radiocarbon age calibration curve since 1986 (Stuiver et al. 1986). Over that time, the complexity of these models has increased as we have gained more knowledge of both the Earth systems and past climate constraints. Recent approaches include more advanced carbon cycle box models which allow transient forward simulations such as BICYCLE (Köhler et al. 2006) through to full three-dimensional ocean general circulation models such as the LSG OGCM (Butzin et al. 2005, 2012, 2017, 2020 in this issue). For all of these models, the basic principle is the same: given a specific user-defined set of inputs/parameters, they provide an estimate/prediction of the variable of interest, in our case the non-polar global-average marine $\Delta^{14}\text{C}$, where this average in the following is restricted to the area between 40°S and either 50°N (Atlantic) or 40°N (Pacific) to avoid the potential complications of including regions covered with sea ice (e.g. Butzin et al. 2017).

The most useful computer model for a particular situation will depend upon the specific question one wishes to answer. For Marine20, we wish to not only obtain a reliable estimate of the global-average marine ^{14}C concentration over time incorporating known changes in both the carbon cycle and atmospheric ^{14}C production rate, but also to be able to quantify our uncertainty on that estimate. Such quantification of uncertainty is fundamental to radiocarbon calibration. When calibrating a radiocarbon date, we need not only a best “point estimate” for its calendar age but also a plausible range, i.e. to understand the uncertainty on our calibrated calendar age. These two competing aims of reliability and uncertainty quantification typically require a trade-off in model selection. We require a computer model that can incorporate desired physical processes and changes to the carbon cycle that proxy data tell us occurred; and yet also allows us to understand and propagate model uncertainty, specifically in the values of the input parameters, through to the final marine $\Delta^{14}\text{C}$ estimate. In the case of our chosen BICYCLE computer model, the various inputs are illustrated in Figure 2.

A Brief Introduction to Quantifying Model Uncertainty

The majority of computer models simulating changes in the carbon cycle are deterministic. This means that they do not internally introduce randomness—if one drives the model with the same inputs one will always get identical outputs. Let us assume that such a computer

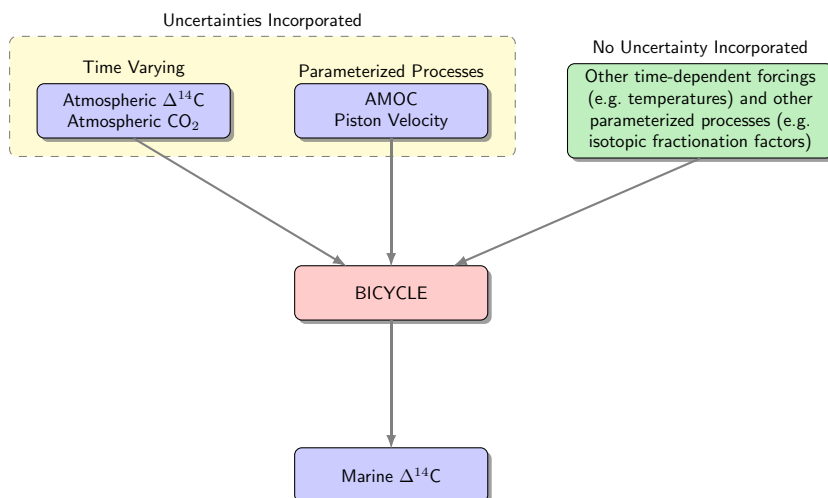


Figure 2 Propagating uncertainty through the BICYCLE model. For each simulation, we generate different inputs for BICYCLE by drawing from the variables on which we consider uncertainty (shown in the yellow box). AMOC denotes the Atlantic meridional overturning circulation and piston velocity the rate of air-sea gas exchange, see Section 4 for further details. These random inputs are combined with the inputs for which no uncertainties are incorporated (shown in green) and entered into the BICYCLE model. Each simulation of BICYCLE therefore provides a different, deterministic historic global-average estimate of marine surface $\Delta^{14}\text{C}$ from 0–55 cal kBP. From the resulting ensemble we generate a Monte Carlo estimate of the mean and variance at any calendar age which is then used for the Marine20 radiocarbon age calibration curve.

model takes as inputs a range/vector of parameters ϕ which we specify. In our context, these model parameters are any input needed by the model—they could be a forcing, a user chosen setting or a tuned variable. In the case of marine ^{14}C reservoir age simulations, these input parameters typically include the historic levels of atmospheric $\Delta^{14}\text{C}$ and CO_2 concentration; ocean circulation; and air-sea gas exchange amongst others. Having specified values for these inputs, the model prediction at any calendar age θ is then represented by the deterministic function $f(\theta; \phi)$.

In the case of a deterministic computer model, output uncertainty arises from two sources (see Section 2 of Kennedy and O’Hagan 2001). The first is model parameter uncertainty, i.e. that, typically, we do not know the true value of the model’s input parameters. We specify our uncertainty on our computer model inputs in terms of a distribution $\phi \sim \pi(\phi)$ that encapsulates our current understanding about their values. The second source of uncertainty is model discrepancy, i.e. that any computer model is not an entirely accurate representation of reality and will not capture the complete carbon cycle of the Earth. This type of uncertainty is much harder to evaluate but one might expect a more complex and detailed model to typically have a smaller model discrepancy. We do not attempt to formally incorporate model discrepancy into Marine20 but we are able to incorporate model parameter uncertainty.

For calibration purposes, we wish to estimate, for any calendar age θ , both the expectation $\mu_{Mar}(\theta)$ and the variance $\sigma_{Mar}^2(\theta)$ in the global-average level of marine $\Delta_{MarineAv}^{14}\text{C}(\theta)$. To propagate our model parameter uncertainty, we therefore need to evaluate both:

$$\mu_{Mar}(\theta) = E[\Delta_{MarineAv}^{14}C(\theta)] = \int f(\theta; \boldsymbol{\phi})\pi(\boldsymbol{\phi})d\boldsymbol{\phi},$$

and

$$\sigma_{Mar}^2(\theta) = Var[\Delta_{MarineAv}^{14}C(\theta)] = \left\{ \int f^2(\theta; \boldsymbol{\phi})\pi(\boldsymbol{\phi})d\boldsymbol{\phi} \right\} - \mu_{Mar}^2(\theta),$$

Where $\pi(\boldsymbol{\phi})$ is the density of the model's input parameters that specifies our prior beliefs/uncertainties about their true values, and $f(\theta; \boldsymbol{\phi})$ is the model output at calendar year θ cal kBP when it has been run with specific inputs $\boldsymbol{\phi}$.

Monte-Carlo Estimation

Due the complexity of the computer model, it is not possible to calculate either of these integrals explicitly, but we can estimate them via a Monte-Carlo approach. To do so, we sample a range of N plausible inputs $\boldsymbol{\phi}_1, \dots, \boldsymbol{\phi}_N$ from our density $\pi(\boldsymbol{\phi})$. The model is then run with each of these inputs giving us a large ensemble of N outputs $f(\theta; \boldsymbol{\phi}_1), \dots, f(\theta; \boldsymbol{\phi}_N)$. The sample mean and variance of this ensemble then provide estimates for $\mu_{Mar}(\theta)$ and $\sigma_{Mar}^2(\theta)$ respectively, i.e. we estimate

$$\hat{\mu}_{Mar}(\theta) = \frac{1}{N} \sum_{i=1}^N f(\theta; \boldsymbol{\phi}_i) \text{ and}$$

$$\hat{\sigma}_{Mar}^2(\theta) = \frac{1}{N-1} \sum_{i=1}^N (f(\theta; \boldsymbol{\phi}_i) - \hat{\mu}_{Mar}(\theta))^2, \text{ where each } \boldsymbol{\phi}_i \sim \pi(\boldsymbol{\phi}).$$

Importantly, as most computer models are highly non-linear, we stress that only running them at extreme “end-member” inputs is not sufficient to reliably represent output uncertainty. Equally, running a computer model with the value of an input parameter set at its mean is not sufficient to estimate the mean of the output when input parameter uncertainty is properly incorporated. Both these points are illustrated with a simple example in the next subsection.

To perform accurate Monte-Carlo estimation, it is necessary to generate a sufficiently large ensemble N of outputs in order that they provide reliable estimates. Here, we selected $N=500$ based on a smaller pilot study which indicated such a sample size would provide Monte-Carlo estimates that were within $\pm 1\%$ of the true model mean, and $\pm 5\%$ of the true model variance. This requires repeated model simulations and so it is necessary to have a model that is sufficiently fast to permit this. BICYCLE provides an appropriate compromise between reducing model discrepancy while still running with sufficient computational speed to enable creation of a large ensemble of outputs and hence accurately propagate model parameter uncertainty. As we show later in Figure 4 and Section 6, there are only small differences between the global-average MRA obtained from individual BICYCLE simulations and from the more complex 3D LSG OGCM alternative, yet each BICYCLE simulation can be performed in seconds rather than the weeks required for the LSG OGCM.

Finally, we note that, as described in Heaton et al. (2020 in this issue) laying out the statistical methodology for the atmospheric calibration curve, one could calibrate directly against the ensemble of N model outputs if covariance information on the marine calibration curve is

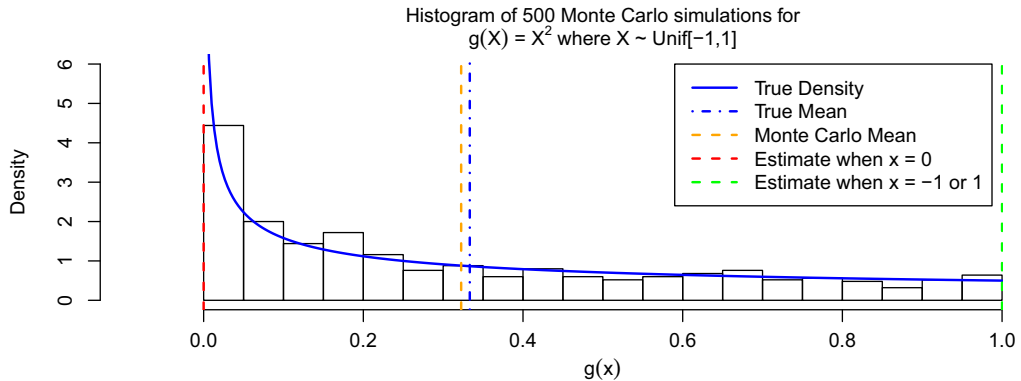


Figure 3 Toy example to illustrate the need for the Monte Carlo approach to accurately estimate both the mean and variability in the output of a computer model.

desired. However, since current calibration tools use pointwise calibration curve means and variances, we do not discuss this further here.

A Toy Monte-Carlo Example

To illustrate the need for the Monte-Carlo approach, if we wish to accurately estimate the mean output of a computer model and capture the model uncertainty, we consider a simple example. While somewhat pathological it does illustrate the key points that running a computer model at the extreme values of its inputs does not ensure we have captured the full uncertainty in the model output; and also that setting the inputs at their mean values does not result in the mean model output.

Specifically, let us suppose we have a computer model that takes a single input x and returns the value $g(x) = x^2$. Further, assume that we do not know x but can quantify our prior belief that it lies uniformly between -1 and 1 , i.e. $X \sim Unif[-1, 1]$. With such a simple model we can work out exactly what the distribution of the outputs should be. This is shown by the solid blue line in Figure 3. All values in the range $[0, 1]$ are clearly possible outputs. Further, we can calculate the true mean output $E[g(X)] = \frac{1}{3}$ (dot-dashed blue line).

However, if we run the model at the extreme ends of its possible inputs (either $x=1$ or -1), in both cases we observe the output $g(x) = 1$ (green dashed line). If we were to assume that these end-member outputs, generated by the extreme inputs, informed us about the full range of potential outputs then we would falsely assume that the output of the model had no uncertainty. Similarly, to find the true mean of the model output we cannot simply run the model with the input x set to its mean value, i.e. $E[X] = 0$. Doing so would provide the output $g(0) = 0$ (red dashed line) rather than the correct mean of $\frac{1}{3}$.

On the other hand, if we create a Monte Carlo sample of 500 independent $X_i \sim Unif[-1, 1]$ values and propagate each of these through the model we obtain the ensemble of outputs shown by the histogram. Comparing this against the true output density in solid blue, we can see that our random ensemble is able to capture the correct uncertainty in the model's output. Further, the mean of these 500 outputs (orange dashed line) shows that, by Monte-Carlo, we are able to accurately estimate the true mean output.

3. WHY DO WE NEED A MONTE CARLO APPROACH AND WHY ARE WE NOT USING REGIONAL 4D OGCM OUTPUT DIRECTLY?

Why Do We Need Monte-Carlo for Marine20?

To be of use for calibration, it is essential to accurately represent the full range of uncertainty in the level of marine radiocarbon. When calibrating a radiocarbon determination, we need to discover all calendar ages which are potentially consistent with that level of ^{14}C . As such we are required to understand the variability in the output of our, complex and non-linear, carbon cycle computer model. This is somewhat different from much modeling which aims to test specific scenarios. As described in Section 2, in order to capture the full variability in the computer model output we must explore the full range of input scenarios. Understanding this uncertainty in computer model output requires us to consider not just the extremes but also those intermediate scenarios, i.e. a Monte Carlo approach.

Limitations of a Constant ΔR Approach

It is expected that the most significant temporal changes in MRA will be shared and hence occur on a global scale rather than in individual regions, i.e. changes in values over time will predominantly be seen in $R^{\text{GlobalAv}}(\theta)$ as opposed to the regional $\Delta R(\theta)$ adjustments (Stuiver et al. 1986). These global-scale MRA changes are incorporated, by construction, in our global-average Marine20 curve and hence automatically taken into account for all calibration users wishing to use the Marine20 curve.

However, in assuming a constant ΔR we do not permit for potential smaller region-specific temporal changes. This assumption of temporally invariant ΔR values should be adopted with caution in the context of a changing climate and a changing marine carbon cycle. Any global marine radiocarbon calibration product that makes use of modern ΔR values should therefore be used with similar caution.

How Were Marine Data Incorporated into IntCal20?

For time periods where sufficient tree-ring ^{14}C determinations exist, i.e. continuously from 0–13,913 cal BP, no marine data were used in the creation of the IntCal20 curve. Further back in time however, marine observations did contribute alongside speleothems, floating tree rings and macrofossils from Lake Suigetsu (see Figure 1 and Reimer et al. 2020 in this issue). To incorporate this marine data into the atmospheric IntCal20, estimates of the MRA for each constituent datum were required. With the exception of Cariaco Basin (Hughen and Heaton 2020 in this issue), for which a different approach was taken due to its unusual topography, to obtain these MRA estimates the following approach was taken.

To begin, an interim atmospheric ^{14}C curve based upon only the Hulu Cave speleothems (Southon et al. 2012; Cheng et al. 2018) was constructed using the same Bayesian spline with errors-in-variables methodology as the final IntCal20 curve (Heaton et al. 2020 in this issue). This interim Hulu-based ^{14}C history was then entered as a forcing to the LSG OGCM to get the coarse regional MRA estimates under its GS scenario (Butzin et al. 2020 in this issue) shown in Figure 3 of Reimer et al. (2020 in this issue).

Since LSG OGCM estimates are intended for the open-ocean as opposed to the more coastal locations of the marine IntCal20 data, a further constant coastal radiocarbon reservoir age adjustment/shift was however required. For each marine dataset, an independent prior was

placed on this required coastal shift based upon the observed offset between the ^{14}C ages of the more recent observations from that specific marine dataset and IntCal20's atmospheric dendrodated trees (which extend back to ca. 14,190 cal BP²).

Intuitively, this coastal shift approach used the relative changes (i.e. the shape) in the regional LSG OGCM open-ocean MRA estimates but not their absolute values. The internally estimated coastal shift is equivalent to estimating a temporally invariant pseudo- ΔR for the specific coastal location relative to the nearby open-ocean. These coastal pseudo- ΔR shifts from the GS scenario of LSG OGCM ranged from approximately 12 ^{14}C yr in Vanuatu up to over 280 ^{14}C yr in Kiritimati indicating the difficulty in using open-ocean OGCM estimates directly for coastal data.

These preliminary Hulu-Cave based MRA estimates are expected to be overly smooth (since they are based upon the relatively smooth Hulu Cave record) however they should provide robust first-order approximations that enable marine data to contribute to IntCal20. The region-specific prior MRA estimates were then used as inputs for the Bayesian approach to IntCal20's curve construction allowing for some further fine-scale MRA variability not captured by the coarse preliminary Hulu-based estimates. In this curve construction process, the coastal shift priors were further updated to resolve differences between the diverse pre-Holocene data sets, see Heaton et al. (2020 in this issue) for complete details.

Importantly, as shown in Figure 3 of Reimer et al. (2020 in this issue), for locations of the marine data used for making IntCal20, the relative differences in the shapes of the Hulu-based regional open-ocean MRA estimates provided by the LSG OGCM were small (in the order of $\pm 5\%$ of the absolute value of MRA). Furthermore, these regional open-ocean MRA estimates had similar shapes to the non-polar global-average MRA. Hence, in terms of the final MRAs used for the marine data contributing to IntCal20, there would be little difference in having applied individual temporally invariant coastal shifts to the non-polar global-average MRA (i.e. the approach we recommend for Marine20 to restrict temporal changes in MRA to the global-average) of the LSG OGCM rather than the region-specific output. Our proposed global-average Marine20 approach to calibration is therefore consistent with that taken to integrate the marine data into IntCal20.

Why Not Use OGCM Output Directly to Provide Regional Calibration Curves?

Much of the marine data which users wish to calibrate come from coastal regions. With our current state of knowledge, most global OGCMs do not have the resolution to simulate these coastal regions accurately and their results may be flawed at such sites. This includes the LSG OGCM. We are therefore still recommending, for the current update, the use of a global-average radiocarbon age calibration curve with the adjustment of a constant ΔR estimated independently based upon the offset between local observations and the Marine20 curve.

In Figure 4, we plot the pre-Holocene LSG OGCM regional open-ocean MRA estimates obtained when forced by the pointwise mean of the final IntCal20 curve (Reimer et al. 2020 in this issue) under the GS scenario. These estimates are the closest LSG OGCM

²Within the IntCal20 database, NH tree-ring ^{14}C determinations extend back to ca. 14,190 cal BP. However, beyond approximately 13,910 cal BP these tree-ring measurements alone are not of sufficient density to estimate the IntCal20 calibration curve precisely. Further back in time than 13,913 cal BP other types of ^{14}C material including macrofossils, speleothems and marine-based samples are therefore also used to construct the curve. See Heaton et al. (2020 in this issue) for more details.

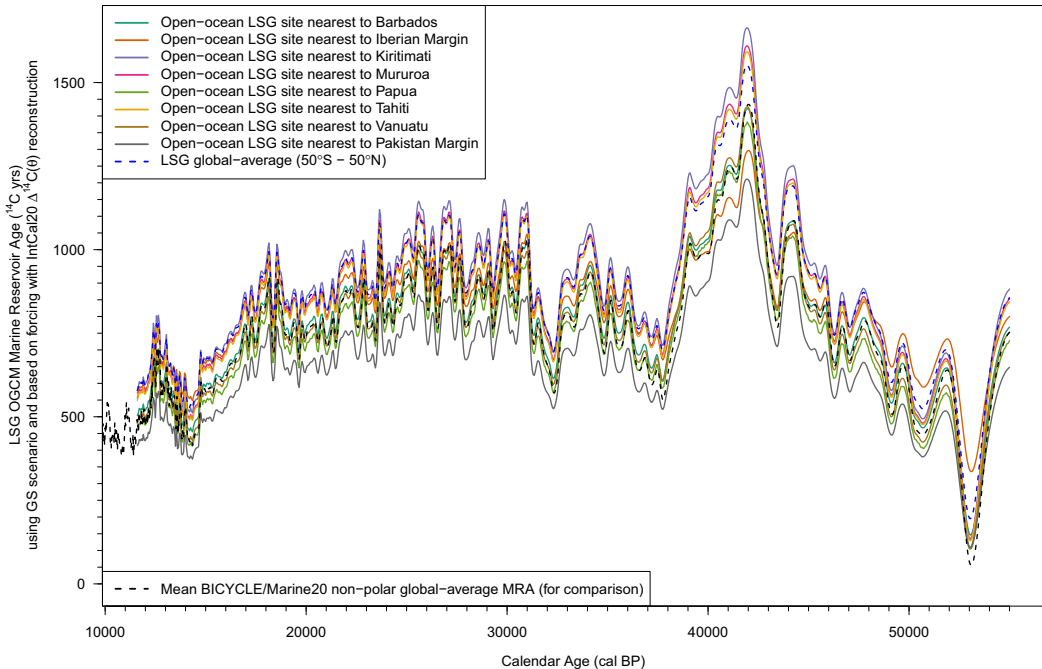


Figure 4 Regional open-ocean MRA estimates generated by the LSG OGCM when forced by the mean of the IntCal20 (Reimer et al. 2020 in this issue) reconstruction of NH atmospheric $\Delta^{14}\text{C}$. These estimates are obtained under LSG's GS scenario (Butzin et al. 2020 in this issue) and correspond to the open-ocean sites nearest to the locations of the IntCal20 marine data. Also shown is the LSG (50°S–50°N) global-average MRA estimate in the GS scenario; and also the mean of BICYCLE's global-average MRA for comparison.

analogue to our Marine20 approach but do not incorporate any uncertainty propagation as Monte-Carlo is infeasible due to the LSG OGCM's long run time. Shown are the LSG OGCM's MRA estimates for the open-ocean sites lying closest to each of the marine locations used in IntCal20. Also shown are the LSG OGCM (50°S–50°N) globally averaged estimate with the same IntCal20 pointwise mean forcing and GS scenario, and also the Monte-Carlo mean of our BICYCLE/Marine20 globally averaged MRA discussed in detail later.

A potential approach to adapt this open-ocean LSG OGCM output for calibration in wider and/or coastal locations would be that taken for inclusion of marine data in IntCal20. One would initially identify the regional open-ocean LSG OGCM site that lies closest to the particular location of interest, before then applying a constant, baseline-specific, pseudo- ΔR shift from this specific regional open-ocean LSG OGCM estimate. However, extending this approach to Marine20 would be complex and reliant upon users identifying both the correct LSG OGCM regional baseline and the matching pseudo- ΔR shift.

Furthermore, for the non-polar sites illustrated in Figure 4, the relative shapes of the regional open-ocean MRAs estimated by the LSG OGCM are highly similar to both the LSG global-average estimate and also the mean BICYCLE global-average estimate. Changing to two alternative scenarios (CS or PD, Butzin et al. 2020 in this issue) within the LSG OGCM also has little effect on the relative shapes of the resultant regional open-ocean MRA

estimates—the primary effect of such scenario changes are simply constant shifts, up and down respectively, in each regional open-ocean MRA estimate albeit with the size of these shifts potentially varying between regions.

Consequently, the difference in using constant pseudo- ΔR -type shifts from regional LSG OGCM curves, as opposed to constant ΔR shifts from the global-average curves, would be small for these sites. As a result, the proposed approach to radiocarbon age calibration described here, of using a global-average Marine20 radiocarbon age calibration curve based on BICYCLE with application of constant ΔR shifts, is consistent with an approach using regional LSG OGCM curves while also being much simpler and more efficient.

As discussed in more detail in Section 6, the output from the LSG OGCM matches that from BICYCLE at the global-average level suggesting that BICYCLE is a satisfactory model at this broader scale. However, by running much more quickly than the full LSG OGCM, BICYCLE provides greatly more flexibility which provides several key benefits for our calibration curve needs. In particular, LSG (or any other current OGCM) is too computationally expensive to carry out a rigorous exploration of the uncertainty in its output. The use of BICYCLE enables us to explore a much wider range of potential carbon cycle scenarios. Further, the LSG OGCM simulations were kept in a single scenario for the entire period from 0–55 cal kBP. The effect of a transient switch from a glacial scenario to an interglacial, more suitable for the Holocene, within a simulation was therefore unknown. With BICYCLE we can more easily investigate the effect of such scenario changes.

Our BICYCLE approach attempts to minimise potential inaccuracies in marine radiocarbon variability that stem from strictly unknown changes in the marine carbon cycle by exploring a wide range of possible carbon cycle states, providing a trade-off between expected accuracy and precision in resulting calibrated ages. As we describe later in Section 7, key future work lies in developing 3D OGCM models which can provide region-specific output further and understanding their inherent uncertainties. Work towards this goal would be invaluable in improving both marine calibration and our insight into the global carbon cycle.

Users calibrating open-ocean samples who wish to use individual regional MRA estimates provided by the LSG OGCM can find them on PANGAEA (<https://doi.org/10.1594/PANGAEA.914500>). This may also be relevant for users wishing to calibrate data outside Marine20's intended range, i.e. where sea ice may have been present (north of 40/50°N or south of 40°S), especially if the LSG OGCM estimates exhibit significantly different relative changes. One approach to incorporate LSG OGCM MRA estimates, for example into OxCal, can be found via Alves et al. (2019).

4. BICYCLE—THE BOX MODEL OF THE ISOTOPIC CARBON CYCLE

BICYCLE is a global carbon cycle box model that includes terrestrial, atmospheric and oceanic modules (Köhler and Fischer 2004; Köhler et al. 2005). The terrestrial biosphere is modeled with seven different compartments distinguishing C_3 and C_4 photosynthesis, and soils with different turnover times; the atmosphere as a single globally averaged box; and the ocean as a set of ten boxes covering different spatial regions (both equatorial and high latitude) and ocean depths including deep ocean-sediment exchange fluxes. Prognostic variables solved by the model are C (as DIC in the ocean, as CO_2 in the atmosphere), ^{13}C , and ^{14}C in all boxes, and additionally alkalinity, O_2 , and PO_4 in the ocean. The marine carbonate system is fully defined by the two variables DIC and alkalinity. The model

design has been chosen to allow long-term (thousands of years) transient simulations, but still capture the essential processes important for the global carbon cycle and carbon isotopes. Temporal changes in the physical boundary conditions (temperature, ocean circulation, sea ice extent, sea level, aeolian dust/iron input, ^{14}C production rate) are prescribed, and thus changes in the simulated subsystem of the global carbon cycle are externally forced. In past studies covering Termination I, the last glacial cycle, or the past 740 kyr (Köhler et al. 2005, 2006, 2010, 2014; Köhler and Fischer 2006) BICYCLE has been used to reproduce (simulate) reconstructed variables of the carbon cycle, such as atmospheric CO_2 , ^{13}C , and ^{14}C .

The version of BICYCLE used here is based on Köhler et al. (2006), in which a previous attempt to simulate the changes in atmospheric $\Delta^{14}\text{C}$ over the last 50 cal kBP documented in IntCal04 has been published. For our current Marine20 application, the following model improvements with respect to that earlier application have been implemented:

- In Köhler et al. (2006) sediment-deep ocean fluxes of DIC and alkalinity were prescribed by temporal change in the depth of the lysocline. The representation of sediment-deep ocean fluxes was revised in Köhler and Fischer (2006) to a scheme that mimics the process of carbonate compensation (Broecker and Peng 1987) in a time-delayed response function to changes in deep ocean carbonate ion concentration. This response function approach has been used in all later applications of BICYCLE and is also applied here.
- Atmospheric CO_2 in the standard BICYCLE setup is a prognostic variable, i.e. is internally calculated rather than externally forced/prescribed. BICYCLE's simulated CO_2 time series has been in reasonable agreement with ice core data, but never matched them perfectly. However, for our current application we have decided to override BICYCLE's internal CO_2 estimates and instead prescribe CO_2 from ice core data (Figure 5, based on Köhler et al. 2017). This decision was made since the air-sea gas exchange, one of the key processes determining the simulated surface ocean $\Delta^{14}\text{C}$, is significantly dependent on the atmospheric concentration of CO_2 . In making this change, the internally calculated carbon cycle is partly corrected by an additional source/sink term in the atmosphere to arrive at the prescribed ice core reconstructed CO_2 value. This is a common approach in carbon cycle model applications for future scenarios, in which not CO_2 emissions, but CO_2 concentrations are prescribed, and has recently been implemented in BICYCLE for its contribution to a carbon cycle model-intercomparison (Keller et al. 2018).
- Atmospheric $\Delta^{14}\text{C}$, which was also a prognostic variable in previous versions of BICYCLE, is now also prescribed from data by individual realizations of NH atmospheric $\Delta^{14}\text{C}$ taken from the Bayesian posterior of IntCal20 (Reimer et al. 2020 in this issue). Here, ^{14}C production rates are simply adapted to match prescribed $\Delta^{14}\text{C}$ values. Thus, no additional source term (as was necessary for CO_2) has to be introduced.

Our model improvements of forcing atmospheric CO_2 from external ice core data and atmospheric $\Delta^{14}\text{C}$ by realizations of the IntCal20 curve aim to correct for any potential limitations in the BICYCLE model. These changes should bring the carbon cycle of the surface ocean closer to its true state than previous BICYCLE simulations where CO_2 and atmospheric $\Delta^{14}\text{C}$ were fully prognostic variables.

These new approaches of externally prescribing atmospheric CO_2 and $\Delta^{14}\text{C}$ manage to bring these variables in the model very close to (<1 ppm for CO_2 and <1 ‰ for $\Delta^{14}\text{C}$) but not in full agreement with the prescribed time series. These minor differences remain because the

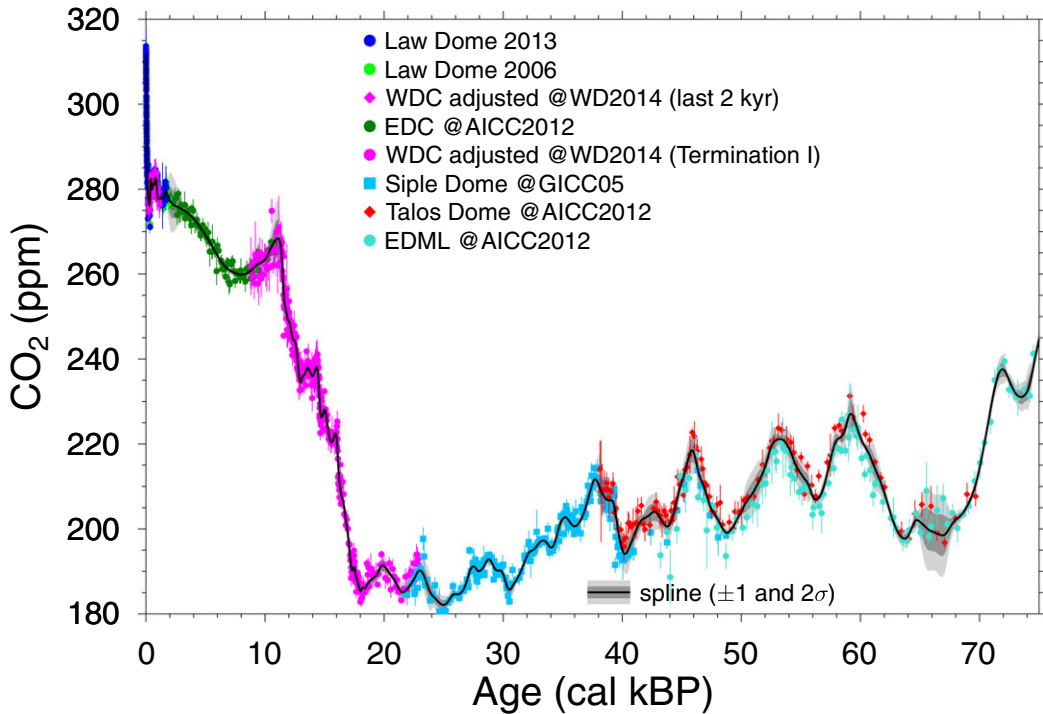


Figure 5 Atmospheric CO₂ concentration used to force the BICYCLE model. Underlying ice core data and corresponding fitted spline are plotted for 0–75 cal kBP (modified from Köhler et al. 2017).

prognostic variables in BICYCLE are determined through ordinary differential equations which are numerically solved with a 4th order Runge-Kutta method (Press et al. 1992). Such differences also prevent us from directly proposing how the ¹⁴C production rate should vary over time to be in agreement with IntCal20. Nevertheless, with more efforts in this direction it might, in the future, be possible to further constrain ¹⁴C production rate with atmospheric Δ¹⁴C by this approach.

Our simulations here are run over the last 75 cal kBP. Between 75 and 55 cal kBP, atmospheric Δ¹⁴C is kept fixed at its value for 55 cal kBP but CO₂ and all other time-dependent forcings vary according to the data. These 20 kyr prior to 55 cal kBP give the radiocarbon cycle in BICYCLE reasonable time to adjust to the boundary conditions (for further details see Köhler et al. 2006).

For Marine20, we consider the DIC-weighted mean of the simulated Δ¹⁴C of the equatorial surface ocean boxes as the non-polar global-average. These boxes are 100 m deep and range from 40°S to 50°N (Atlantic) or 40°N (Pacific).

Propagating Model Parameter Uncertainty through BICYCLE

While BICYCLE is a box model, it still contains a large number of parameters. To keep our approach practical, we focus our analysis of propagated model uncertainties on key processes and do not consider the uncertainty in all parameters. From sensitivity tests we identified that two processes are most important for simulated surface ocean Δ¹⁴C: piston velocity; and

Atlantic meridional overturning circulation (AMOC), which is directly connected to the strength of North Atlantic Deep Water (NADW) formation.

Piston velocity is the rate of air-sea gas exchange and depends on near-surface wind speed. Thus, this parameter is important for the oceanic uptake of ^{14}C from the atmosphere. In all previous applications, it was parametrized to 2.5×10^{-5} and $7.5 \times 10^{-5} \text{ m s}^{-1}$, for the equatorial and high latitude surface boxes respectively (Heimann and Monfray 1989). These values agree reasonably well with more recent estimates (Wanninkhof 2014). To create Marine20, we maintain these values as the mean piston velocities but introduce uncertainty by placing a prior on each informed by a meta-analysis as described below.

The strength of the AMOC in the model is important for the transport of tracers from the surface to the deep ocean via the physical carbon pump. In our Marine20 setup, the AMOC, represented by strength of the NADW formation, is considered to have two states with some uncertainty on the value in each state. In interglacials, it is centred around 16 Sv (1 Sverdrup = $10^6 \text{ m}^3 \text{ s}^{-1}$, e.g. Talley et al. 2011) while during glacials it is centred at 10 Sv. See below for more details.

In addition to uncertainties in piston velocity and AMOC, we aim to propagate uncertainty on the past levels of atmospheric $\Delta^{14}\text{C}$ and CO_2 . Below, we present a summary of how these various inputs were selected together with a description of their uncertainties. Those model parameters for which we consider uncertainty, and specify distributions on their values, are referred to as having *uncertainty incorporated*; while those for which we do not consider uncertainty are denoted as having *uncertainty not incorporated*. An overview of how these enter BICYCLE is given in Figure 2.

TIME DEPENDENT MODEL PARAMETERS FOR WHICH UNCERTAINTY IS INCORPORATED

Atmospheric $\Delta^{14}\text{C}$: for each BICYCLE simulation, we draw a distinct posterior realization of NH atmospheric $\Delta^{14}\text{C}$ from the Bayesian spline of IntCal20 (Reimer et al. 2020 in this issue). Each of these realizations represent a different plausible past atmospheric history. To input into BICYCLE, they are sampled on a 10-calendar-year input grid and interpolated in-between. Using realizations in this way, as opposed to the pointwise IntCal20 curve summaries, allows us to incorporate not only our inherent uncertainty on the level of atmospheric $\Delta^{14}\text{C}$ at any calendar age but also its smooth dependence over time. See Heaton et al. (2020 in this issue) for some illustrative atmospheric realizations over the Bølling-Allerød and further explanation on how these individual posterior realizations form the summarized IntCal20.

Atmospheric CO_2 : our record of past atmospheric CO_2 (Figure 5) is taken from a spline through ice core data as described in Köhler et al. (2017). The spline estimate combines raw data from several different ice cores (Talos Dome, Siple Dome, WAIS Divide (WDC), EPICA Dome C (EDC), EPICA Dronning Maud Land (EDML) and Law Dome) each on the most recent age scale, e.g. AICC2012, GICC05, WD2014 (Ahn and Brook 2014; Ahn et al. 2012; Bauska et al. 2015; Bereiter et al. 2012; Lüthi et al. 2010; MacFarling-Meure et al. 2006; Marcott et al. 2014; Monnin et al. 2001, 2004; Rubino et al. 2013). This fitted spline is accompanied by uncertainties that consider data resolution errors, Monte Carlo errors, and error due to a chosen cutoff period during smoothing leading to combined uncertainties in the spline, $\sigma_{\text{CO}_2}(\theta)$. For each of the 500 realizations used as model inputs, we draw a Gaussian random variable $x_i \sim N(0, 1^2)$ that defines the specific CO_2 time series used as a forcing on

simulation i ,

$$CO_2^i(\theta) = \mu_{CO_2}(\theta) + x_i \sigma_{CO_2}(\theta),$$

where $\mu_{CO_2}(\theta)$ is the mean of the ice core spline at calendar age θ . This ensures that each individual CO_2 forcing varies smoothly over time.

OTHER KEY PROCESS MODEL PARAMETERS FOR WHICH UNCERTAINTY IS INCORPORATED

As mentioned above, the two most influential processes for surface ocean $\Delta^{14}C$ are piston velocity and AMOC (the strength of the NADW formation). Standard values have been taken for their means as in previous studies (e.g. Köhler et al. 2005, 2006) in which they have been justified. We specify uncertainties around these means, providing our reasoning below.

Piston Velocity:

Piston velocity is a key factor regulating the uptake of ^{14}C by the ocean and hence largely determines simulated MRAs. Piston velocity depends on near-surface wind speed for which our knowledge is limited, in particular regarding values of the past. Therefore, piston velocities are perhaps the least well-constrained parameters in our model.

As stated above, BICYCLE operates with two piston velocities—dependent upon whether the surface ocean box falls in low or high-latitudes, which typically have either low or high wind speeds, respectively. Since BICYCLE has been tuned previously, and there is no common agreement on their absolute values, we do not change the mean values of these parameters as to do so would lead to a different carbon cycle baseline, potentially requiring further adjustments/analysis. We do however incorporate uncertainties around these means.

In specifying piston velocities and their uncertainties, we are not concerned with short-term, annual-scale, fluctuations. Short-term changes do not have a significant effect on the marine ^{14}C reservoir since the slow air/ocean mixing process occurs over much longer time scales and so smooths such fine variations out. Instead, our interest is in quantifying the uncertainty in the long-term mean values. We model the piston velocities as constant over time and specify a $\pm 15\%$ (1σ) uncertainty on both the high-latitude and equatorial values:

$$\text{Equatorial Boxes } k_{Eq} \sim N(2.5 \times 10^{-5}, (0.375 \times 10^{-5})^2) \text{ m s}^{-1},$$

$$\text{High – Latitude Boxes } k_{HI} \sim N(7.5 \times 10^{-5}, (1.125 \times 10^{-5})^2) \text{ m s}^{-1}.$$

Justification for Selection of Piston Velocity Uncertainties

To quantify historic piston velocity uncertainty, we first consider present day mean piston velocity with present day winds/climate. Naegler (2009) provides five current day piston velocity estimates (with individual uncertainties) on comparable scales. We combine these values using a statistically rigorous meta-analysis (Figure 6) to give an estimate for the average present day piston velocity of $\bar{k} \sim N(17, 1.4^2)$ cm hr^{-1} (or equivalently $N(4.7, 0.39^2)$ 10^{-5} m s^{-1}), i.e. the 1σ uncertainty on the level of piston velocity with present day climate is approximately $\pm 10\%$.

Further uncertainty arises as a result of changing past climate, in particular different wind speeds. From the LSG OGCM (Butzin et al. 2017, 2020 in this issue) we obtain (Table 1)

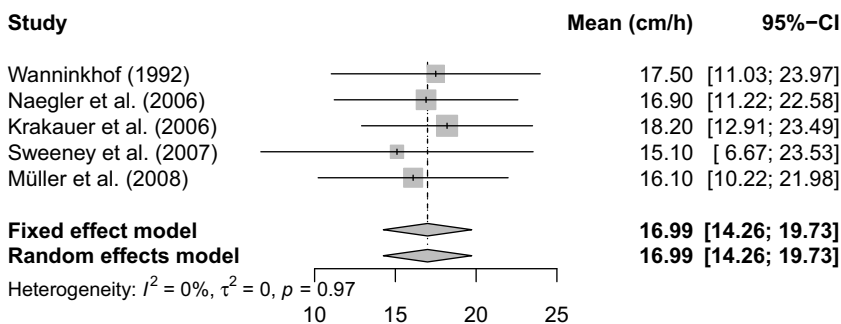


Figure 6 Meta-analysis of present-day estimates of piston velocity from Naegler (2009). Shown are, post-Naegler's adjustments, the piston velocities (cm hr⁻¹) from 5 different studies together with their 95% confidence intervals (95%-CI). The grey diamonds represent the combined estimate, again with 95% intervals. Heterogeneity, as quantified by the measures I^2 , τ^2 and p (Cochran's Q-statistic), assesses the extent to which the reported velocities differ between the different studies, see e.g. Borenstein et al. (2011) for more details. Here, the five reported velocities are seen to be consistent with a single shared underlying velocity.

gas exchange rates based on 10-year averaged monthly model results for three different climate scenarios (PD, GS and CS, Butzin et al. 2005).

We use these values to indicate the potential additional variability introduced by different past climate scenarios. These estimates suggest that, due to changes in climate, the mean piston velocity might change, within any particular basin, from its present day value by approximately another $\pm 10\%$ (at 1σ based upon the difference in the LSG OGCM estimates between the interglacial PD scenario and the glacial scenario CS). Finally, we combine the uncertainty on the present day value ($\pm 10\%$) with the LSG-based intra-basin climate variation ($\pm 10\%$) to get an approximate combined uncertainty of $\sqrt{0.1^2 + 0.1^2} \simeq 0.15$, i.e. $\sigma = 15\%$. This uncertainty is comparable with the $\pm 20\%$ proposed by Wanninkhof (2014) for the mean value of the present ocean.

Ocean Circulation

The strength of the AMOC, expressed by NADW formation, presumably varied between a weak glacial and a strong interglacial mode. In our model, the switch between both is a function of North Atlantic sea surface temperature and has been chosen such that the interglacial mode is reached at the onset of the Holocene. In each mode, we assume a constant, but unknown NADW value drawn from a suitable prior representing the two different strengths. Millennial-scale variations in the Atlantic meridional overturning circulation connected with the bipolar seesaw (e.g. Barker et al. 2009) have been neglected for this update although are a topic identified for future work (see Section 7):

$$NADW(\theta) = \begin{cases} \alpha_{Glacial} & \text{if } t \in \text{Glacial mode,} \\ \alpha_{Inter} & \text{if } t \in \text{Interglacial mode,} \end{cases}$$

where:

$$\begin{aligned} \alpha_{Glacial} &\sim N(10, 2^2), \\ \alpha_{Inter} &\sim N(16, 1^2). \end{aligned}$$

Table 1 Piston velocities for $^{14}\text{CO}_2$ employed by the LSG OGCM according to various climate forcing scenarios, results are annual-mean values averaged over ocean areas considering the same latitude range as used by BICYCLE. The scenarios considered are PD (a present-day climate scenario which may also be considered as a surrogate for interstadials), and two glacial scenarios—CS based on the CLIMAP reconstruction (McIntyre et al. 1981) and GS based on the GLAMAP climate reconstruction (Sarnthein et al. 2003). The glacial climate scenarios involve some adjustment of atmospheric forcing fields in the tropics (CS) and in the Southern Ocean (both CS and GS; see Butzin et al. 2005 for further details).

Climate scenario	Ocean LSG OGCM gas exchange rate estimate (cm hr^{-1})			
	Global	Atlantic	Indo-Pacific	Southern Ocean
PD	13.3	13.7	12.8	15.5
GS	13.5	14.2	12.5	15.8
CS	12.0	12.7	11.1	14.4

DEEP OCEAN MODEL VALIDATION

We provide, in addition to the comparison of BICYCLE's non-polar global-average (i.e. equatorial box) surface ocean output against observational data and the LSG OGCM estimates of MRA provided in Section 6, a further model validation from deep ocean ^{14}C data. Specifically, we consider the level of ^{14}C depletion, in terms of radiocarbon years, between the atmosphere and BICYCLE's three deep-ocean boxes which cover the global ocean deeper than 1000 m. The weighted average from all 500 BICYCLE simulations of these deep ocean boxes indicate that their ^{14}C depletion decreases from the LGM (23–18 cal kBP) to pre-bomb values (0 cal kBP) by 605 ± 103 ^{14}C yr. This is slightly less, but agrees within the uncertainties, with the 689 ± 53 ^{14}C yr age difference of the volume-weighted global mean ocean between those two time windows obtained from a model-based interpolation of 256 deep ocean ^{14}C samples (Skinner et al. 2017) and illustrates that the changes in ocean circulation assumed in BICYCLE are within reasonable bounds.

5. THE MARINE20 CURVE

In Figure 7A we present, in the $\Delta^{14}\text{C}$ domain, the resultant ensemble of 500 BICYCLE estimates of non-polar global-average surface ocean $\Delta^{14}\text{C}$ which constitute the Marine20 curve. Shown are the estimated pointwise means and 95% probability intervals. For comparison, we also include the IntCal20 atmospheric $\Delta^{14}\text{C}$ curve (Reimer et al. 2020 in this issue), which was used as input for our estimate of Marine20, and the previous Marine13 marine radiocarbon calibration curve based on IntCal13 (Reimer et al. 2013). Figure 7B shows the ensemble of corresponding estimates of non-polar global-average MRA, i.e. the depletion, in terms of radiocarbon years, between the atmosphere and the non-polar global-average marine surface ocean over time. Note that each MRA estimate in the ensemble of BICYCLE outputs is coupled to a specific realization of atmospheric ^{14}C .

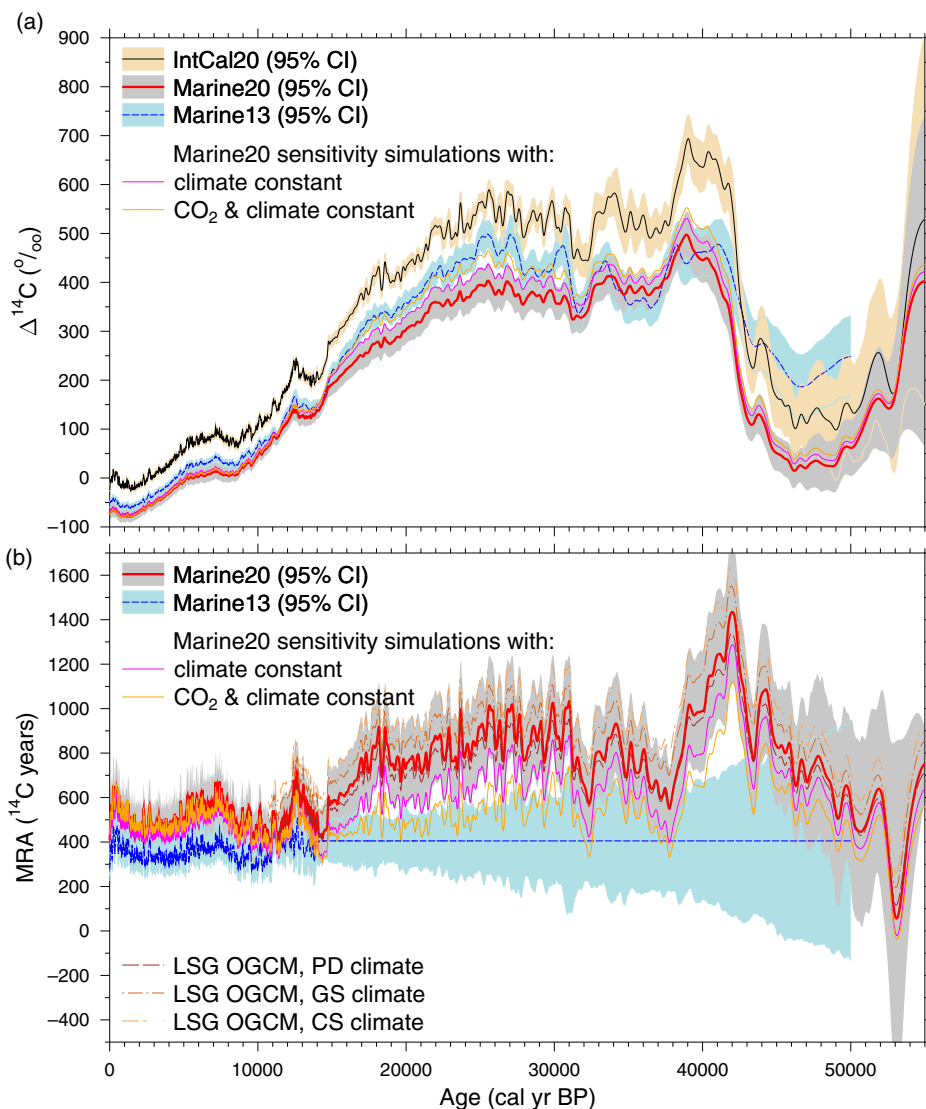


Figure 7 Panel A: The Marine20 curve (obtained via Monte-Carlo statistics of the ensemble of 500 BICYCLE simulations) in $\Delta^{14}\text{C}$ compared to Marine13 and the atmospheric IntCal20 curve. We show the mean and 95% probability intervals for the curves; IntCal20 is shown here with its published 95% predictive interval incorporating the over-dispersion seen in NH atmospheric ^{14}C tree-ring measurements. Additionally, two sensitivity simulations based only on mean values are shown, the first in which in BICYCLE climate; and the second in which both climate and CO₂ have been kept constant. Panel B: The non-polar global-average MRA corresponding to Marine20 (estimated by BICYCLE) compared to three scenarios of LSG OGCM and the global-average MRA previously assumed in Marine13.

We also plot, in Figure 10, the complete Marine20 curve in the radiocarbon age domain against unadjusted ^{14}C observations from corals and forams obtained from the marine environment (some of which were used to create IntCal20 and hence also influence Marine20). Here, we also show the atmospheric IntCal20 curve. This is discussed further in Section 6.

Marine20 (Figure 7A) shows a similar shape to Marine13 but with a significantly lower $\Delta^{14}\text{C}$ between ~ 15 and ~ 32 cal kBP, and beyond 42 cal kBP, indicating a larger global-average MRA in Marine20 than in Marine13. During the Holocene, i.e. from 0–11.6 cal kBP, Marine20 shows an approximate constant $\Delta^{14}\text{C}$ deficit to the atmosphere with a global-average MRA of around 500 ^{14}C yr although, as expected, sharp changes in atmospheric $\Delta^{14}\text{C}$ are smoothed in the marine environment. Here, the carbon cycle model elements within BICYCLE, as indicated by the atmospheric CO_2 concentrations (Figure 5), remain fairly constant. For most of this 0–11.6 cal kBP time period, Marine13 is based on a time-constant box diffusion model (Hughen et al. 2004), albeit with a different parameter choice, and so similar longitudinal changes in MRA might be expected.

Beyond 11.6 cal kBP, we see the effect of the carbon cycle changes within Marine20. The marine ^{14}C deficit decreases as the atmospheric CO_2 concentration and temperature rose from their lower glacial into their higher Holocene values and atmospheric $\Delta^{14}\text{C}$ decreased. From 20–30 cal kBP, the global-average MRA was 750–1000 ^{14}C yr (Figure 7B). These higher global-average MRA values are accompanied by larger temporal variations, likely a consequence of the increased variation in atmospheric $\Delta^{14}\text{C}$. Further back in time than this, the global-average MRA shows two local minima of 600 ^{14}C yr around 33 and 38 cal kBP, into which global-average MRA decreased from its maximum value of 1400 ^{14}C yr connected with the Laschamp geomagnetic excursion (approximately 41–42 cal kBP, Lascu et al. 2016). These high MRA values are likely predominantly driven by the corresponding large increase in atmospheric $\Delta^{14}\text{C}$ seen in IntCal20. Note however, that this peak in MRA around the Laschamp geomagnetic excursion coincides with the inflection point in atmospheric $\Delta^{14}\text{C}$, predating the actual $\Delta^{14}\text{C}$ peak by about 3000 years. This feature is also discussed in more detail in Butzin et al. (2020 in this issue).

It is in this older period, beyond 13.9 cal kBP, where the improvement in methodology leading to different results between Marine20 and Marine13 is most evident. Beyond 13.9 cal kBP, Marine13 assumed a constant reservoir age of 405 ^{14}C yr since, for the creation of that curve, the ability to incorporate carbon cycle changes was limited. The use of BICYCLE for the construction of Marine20 addresses this and permits us more realistic MRA estimation extending into the glacial time period.

A more detailed understanding on the importance of the different processes contained in Marine20 can be gained from sensitivity simulations where we select a subset of the inputs to BICYCLE to vary, while holding others fixed. Initially, we consider two simulations which give an indication as to the effect that past changes to climate and CO_2 concentration have on our global-average ocean mixed-layer ^{14}C estimates. In both these simulations, for simplicity, we only drive BICYCLE by the pointwise-mean atmospheric $\Delta^{14}\text{C}$ values as opposed to ranging over individual realizations. In the first such simulation, all time-dependent boundary conditions constraining climate (temperature, sea level, ocean circulation) are held constant at their level used for 0 cal BP; in the second simulation, in addition to climate, the CO_2 concentration is also kept constant (Figure 7A). When keeping both constant, BICYCLE's results between 20 and 30 cal kBP agree more closely with Marine13. This is to be expected since, for the Marine13 curve, these two aspects were not considered to have varied. Roughly speaking, both the changing climate and the changing CO_2 concentration appear to contribute about the same to Marine20, while the dominant contribution of changes in MRA are caused by variable atmospheric $\Delta^{14}\text{C}$.

In a further set of sensitivity studies, we provide an insight into how each individual process (with its specified prior) contributes to the overall uncertainty in Marine20 by comparing the varying level of uncertainty in BICYCLE's model output when we only propagate uncertainty in selected model inputs. This aims to identify, should we wish to reduce the uncertainty in model output for future work, which are the key processes and parameters on which we must increase our knowledge, and equally which are less critical. Here, we do not consider potential interactions between input variables (Oakley and O'Hagan 2004) in contributing to overall model uncertainty but only the main effects. We performed four additional sets of 500 simulations. In our base scenario (simulation S0) only the uncertainty in atmospheric $\Delta^{14}\text{C}$ was propagated through BICYCLE, achieved by ranging over the 500 posterior $\Delta^{14}\text{C}$ realizations taken from IntCal20, with all other model inputs, i.e. atmospheric CO_2 , piston velocity and AMOC, held fixed at their mean, but still time-dependent, values. The other three scenarios build on this base case and consider the further effect on model output when the additional, previously described, uncertainties in each of the further BICYCLE inputs: atmospheric CO_2 (S1), piston velocity (S2) and AMOC (S3) have also been propagated alongside the uncertainty in atmospheric $\Delta^{14}\text{C}$. By subtracting the uncertainty in BICYCLE's global-average marine $\Delta^{14}\text{C}$ output in S0 from those obtained in S1-S3 we obtained an estimate of the additional contributions of these processes to the overall uncertainty in Marine20.

From these sensitivity tests (Figure 8A) we learn that, for the last 31 cal kBP, the uncertainty in Marine20 is largely dominated by the uncertainty in piston velocity, with initially only a small contribution from the uncertainty in atmospheric $\Delta^{14}\text{C}$. Beyond 31 cal kBP however, the uncertainty in atmospheric $\Delta^{14}\text{C}$ becomes the dominant contributor to Marine20's uncertainty. The effect of uncertainty in atmospheric CO_2 is negligible and that of AMOC also very minor ($<1.5\%$ in terms of $\Delta^{14}\text{C}$) although, as explained earlier, we do not incorporate the potential full range of millennial-scale AMOC variability such as its possible shutdown during Heinrich events. For details on the effect of completely shutting down or switching off the various components of the time-dependent climate forcing in BICYCLE, see Figure 5 in Köhler et al. (2006). Large changes in the precision of our atmospheric $\Delta^{14}\text{C}$ IntCal20 estimate, ranging from the Holocene where it is known very precisely through to the older glacial period where it is much more uncertain, do not penetrate through to Marine20, but are smoothed out by the carbon cycle. The individual contributions nearly add up to the entire uncertainty in Marine20, but between 13–47 cal kBP there is a small positive difference between the combined uncertainty seen in Marine20 and the sum of individual contributions, especially around the Laschamp geomagnetic excursion. Interestingly, the mean of Marine20 is seen to some degree to depend on the considered uncertainties: it differs from the mean in S0 by up to 4‰ in $\Delta^{14}\text{C}$ (Figure 8B), illustrating nicely the non-linear nature of the error propagation (cf. Section 2).

6. COMPARISONS WITH A MORE COMPLEX MODEL, RECENT PRE-BOMB OBSERVATIONS AND OLDER CORALS AND FORAMS

To provide further support for Marine20's robustness and suitability we compare it to the output of a set of simulations provided by the LSG OGCM (Butzin et al. 2017, 2020 in this issue) and against a collection of pre-bomb marine observations which are primarily from surface water, coastal samples (Reimer and Reimer 2001), as well as the marine observations from corals and forams within the IntCal20 database (<http://intcal.org>) extending back to 55 cal kBP.

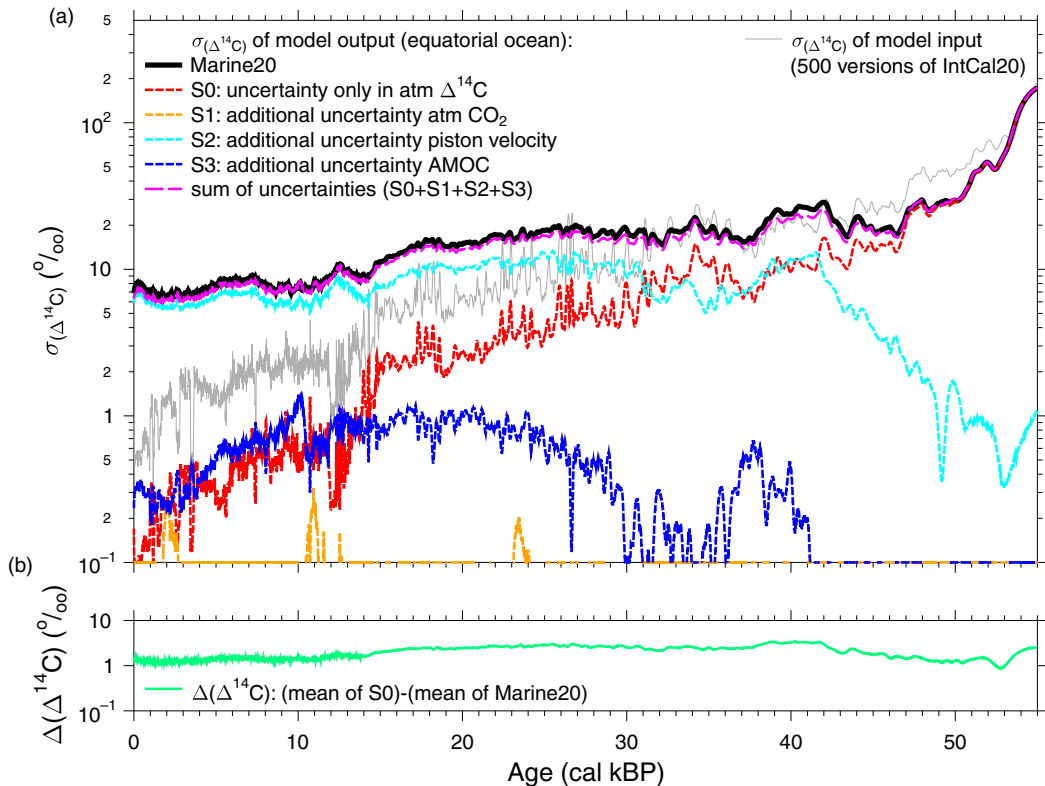


Figure 8 A closer look at the processes contributing to the uncertainties in Marine20. Panel A presents (black line) the pointwise 1σ uncertainty in Marine20's estimate for global-average marine $\Delta^{14}\text{C}$, i.e. the variability in BICYCLE's model output taking into account the selected prior uncertainties in all our model inputs (atmospheric $\Delta^{14}\text{C}$ and CO_2 , piston velocity, AMOC). Also shown (short dotted lines) are the uncertainties observed in BICYCLE's $\Delta^{14}\text{C}$ output when only propagating uncertainty in individual/selected inputs through the model, leaving other inputs fixed. The sum of the output uncertainties resulting from the individual input components (long dashed line) are also plotted, together with the pointwise 1σ uncertainty on the 500 atmospheric $\Delta^{14}\text{C}$ IntCal20 realizations used as inputs to BICYCLE (grey solid line) for reference. Panel B illustrates the difference between the mean global-average marine $\Delta^{14}\text{C}$ estimate obtained by BICYCLE when propagating all chosen input uncertainties (i.e. Marine20), and when only considering uncertainty in atmospheric $\Delta^{14}\text{C}$ and holding all other inputs fixed (S0).

Comparison with the LSG OGCM

The LSG model is a three-dimensional ocean general circulation model incorporating more processes (ocean physics) than a box model. Furthermore, it has the ability to provide location-specific MRAs. Consequently, one might expect the LSG OGCM to provide more detailed and accurate modeling of the global-average marine ^{14}C reservoir than BICYCLE, albeit without the current ability to propagate model parameter uncertainty via Monte-Carlo.

As discussed in Section 3, the LSG OGCM was applied in three different climate scenarios: PD, CS and GS. Details are found in Table 1 and elsewhere (Butzin et al. 2005, 2012, 2017, 2020 in this issue). Similarly to BICYCLE, the model was forced with the temporal changes in the atmospheric carbon records of CO_2 (Köhler et al. 2017) and IntCal20's reconstruction of

$\Delta^{14}\text{C}$ (Reimer et al. 2020 in this issue), although using the pointwise means rather than individual realizations.

For each scenario, the LSG OGCM provided estimates of surface MRAs averaged over 50°N–50°S for the ocean surface above 50 m depth. These are shown in Figure 7B alongside the BICYCLE estimates with their 95% probability intervals. As can be seen, despite BICYCLE being a much simpler model, the BICYCLE estimates retain almost all of the temporal details present within the spatially averaged results from the LSG OGCM. BICYCLE's probability interval also covers the range of LSG scenarios. Users should however be aware that the LSG scenario which has the highest agreement with the mean output from BICYCLE is that with present day climate (PD). This emphasises, that although comparable, climate change over our time window of interest leads to different implications in the two models when considered in detail. Nevertheless, the overall model comparison suggests that, despite its simplicity, BICYCLE is able to provide a robust and accurate global estimate.

Comparison with Present-Day Marine Observations

We also compare Marine20 against pre-bomb (0–200 cal BP) marine observations (Reimer and Reimer 2001). In this comparison, it should be noted that these observations include regions of coastal upwelling and further, due to sampling biases, much of this data arise from the Western US coast, so may not be representative of the open ocean. Both data and model have not been corrected for fossil fuel contamination (Suess effect), thus should be comparable. Note that BICYCLE has previously been shown to track the Suess effect sufficiently accurately, even when applied without forced atmospheric $\Delta^{14}\text{C}$ and CO_2 (Köhler et al. 2014; Köhler 2016). We restrict our comparison to marine data from 50°N–50°S, to remove potential effects due to sea ice cover, and show the comparison in both $\Delta^{14}\text{C}$ and radiocarbon age space (Figure 9, panels A and B). The previous Marine13 estimates are also shown. Figure 9C shows a zoomed-in plot of the estimated global-average MRA over this period.

As we see, the modeled global-average MRA for Marine20 is significantly higher than in Marine13 (Figure 9C). This leads us to a Marine20 curve which consistently lies above Marine13 in terms of the radiocarbon age, or equivalently gives lower $\Delta^{14}\text{C}$ values, for a particular calendar age θ .

In the $\Delta^{14}\text{C}$ space, the weighted mean of the sampled marine observations covering 0–200 cal BP is $-61 \pm 16\text{‰}$, agreeing with a further subset of samples published recently (Toggweiler et al. 2019) for 1940–1954 CE (i.e. 10 cal BP to –4 cal BP).

Although an over-time decreasing trend in $\Delta^{14}\text{C}$ due to the Suess effect seems visible in the observed marine data from ca. 1900 CE onward (i.e. 50–0 cal BP) it is statistically non-significant (p -value 0.35). Results are nearly identical when restricted to 40°N–40°S in the data selection, although this leads to a weighted $\Delta^{14}\text{C}$ mean of only -58‰ . In the pre-bomb time window, the global-average MRA in Marine20 drops from slightly above 550 ^{14}C yr at 200 cal BP to ~ 410 ^{14}C at 0 cal BP, mostly due to the Suess effect seen in the atmospheric IntCal20 $\Delta^{14}\text{C}$. This most-recent trend is also very closely reproduced in the simulations with the LSG OGCM (Figure 9C).

The increase in the $\Delta^{14}\text{C}$ deficit between the atmospheric and global-average surface ocean that we see with BICYCLE and Marine20, compared to the simpler box model of Marine13, results

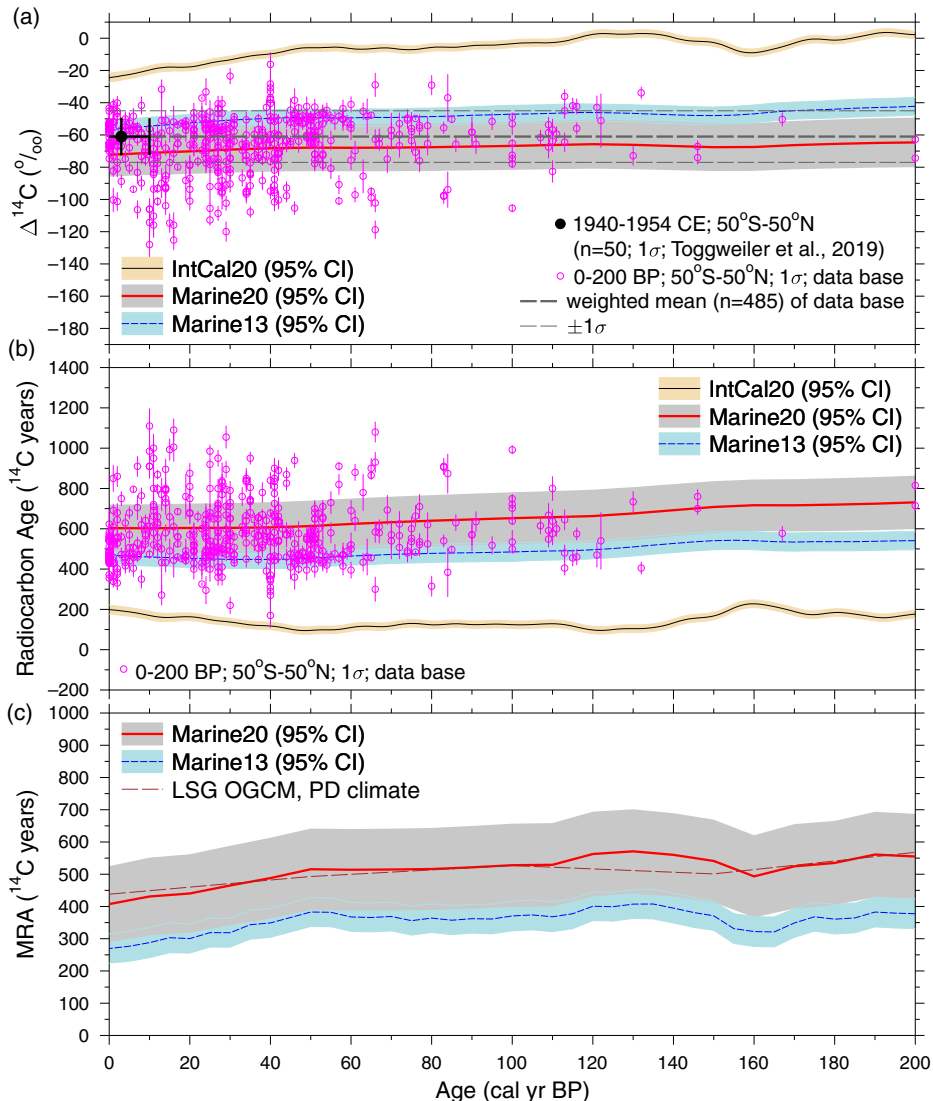


Figure 9 Zoom-in on the pre-bomb time window (0–200 cal BP). IntCal20, Marine20, Marine13 and ^{14}C determinations of observed marine samples (restricted to 50°N – 50°S) in (A) $\Delta^{14}\text{C}$ space, (B) Radiocarbon age space, and (C) illustrating the estimated global-average MRA. In (A, B) ^{14}C determinations from observed marine samples are taken from the data base (<http://calib.org/marine/>), in (A) a subset of additional $\Delta^{14}\text{C}$ data is also shown (Toggweiler et al. 2019). In (C) the global-average MRA from the LSG OGCM output (restricted to 50°N – 50°S) for the scenario PD is also given for comparison. IntCal20 is shown with its 95% predictive intervals, while Marine13 and Marine20 their 95% probability intervals.

in a calibration curve which provides better agreement with our observed present-day observations. As seen in Figure 9A and 9B, the majority of these marine observations have an older radiocarbon age (i.e. their $\Delta^{14}\text{C}$ is lower) than Marine13 and lie closer to Marine20. A further analysis suggested that the optimal offset to Marine20, in terms of maximising the statistical likelihood of the observed present-day data shown in Figure 9, was only 30 ^{14}C yr. This offset was of sufficiently small size to lie within sampling

uncertainty and so no adjustments of the BICYCLE simulations have been applied which might have led to a complete vanishing of the model/data offset.

Comparison with Corals and Foraminifera in IntCal20 Database

Figure 10 plots the complete Marine20 curve, in radiocarbon age space and from 0–55 cal kBP, against all the raw coral and foraminifera data available in the IntCal20 database, <http://intcal.org>. As described in the IntCal20 paper (Reimer et al. 2020 in this issue), corals which are older than 25 cal kBP have been excluded in view of potential diagenesis. Further, we do not include observations from the Cariaco Basin for comparison here since this basin appears to be a unique marine environment (Hughen and Heaton 2020 in this issue) which no model is currently able to resolve, see Section 7 describing the limitations of our model.

We plot here the raw ^{14}C ages of the observations, with no MRA or ΔR correction, along with their 95% probability intervals in both radiocarbon and calendar age. Since we have made no marine radiocarbon age correction for the observations in the plot, we would not expect the Marine20 curve to necessarily fit the raw data themselves. Rather our interest is in whether the offset from the curve (i.e. ΔR) is consistent for each dataset or location. Since marine data from older than 13.91 cal kBP does influence the construction of IntCal20, and hence also Marine20, the Marine20 curve is not entirely independent of the plotted data for this time period—although the large volume of other data informing IntCal20 still mean it is significantly so. From 0–13.91 cal kBP, the Marine20 curve is almost entirely independent of the plotted data since no marine data informed IntCal20 in this more recent period.

As with the present-day observations discussed in the previous section, most of this marine data is coastal and hence may not be representative of open-ocean. Most of the corals are seen to lie below the Marine20 curve suggesting that the locations of these data (Tahiti, Barbados, Vanuatu, Papua and Kiribati) have a rather smaller region-specific MRA than the Marine20 global-average estimates, i.e. have a negative ΔR . For the period between 7 and 12 cal kBP (i.e. the early Holocene) we also see that the MRA for Tahiti is consistently smaller than for Papua New Guinea and Vanuatu, which is compatible with previous work (e.g. Table 1 in Reimer et al. 2009). Beyond 12 cal kBP systematic differences are more difficult to detect because the analytical precision is lower.

Interestingly, between 15 and 17 cal kBP, the Iberian Margin data seem several centuries too old in ^{14}C yr compared with the Marine20 curve. This age range corresponds to the Heinrich 1 stadial and is potentially compatible with the near collapse of AMOC during Heinrich events which are not incorporated into BICYCLE and therefore missing in Marine20. It is harder to detect if this effect is also present in older Heinrich events since the precision of ^{14}C ages decreases through time, and some of these older events may have also had reduced intensity.

7. LIMITATIONS AND FUTURE WORK

While the Marine20 radiocarbon calibration curve already offers many improvements over previous marine calibration curves there are still a number of areas for future refinements.

Perhaps the clearest limitation in our current work is our inability to accurately reproduce the apparent MRA within the Cariaco Basin. This lack of reproduction is however not only an issue for BICYCLE and the LSG OGCM but also, to our knowledge, any current carbon cycle model. Based upon comparison with other IntCal20 data, the Cariaco Basin (Hughen

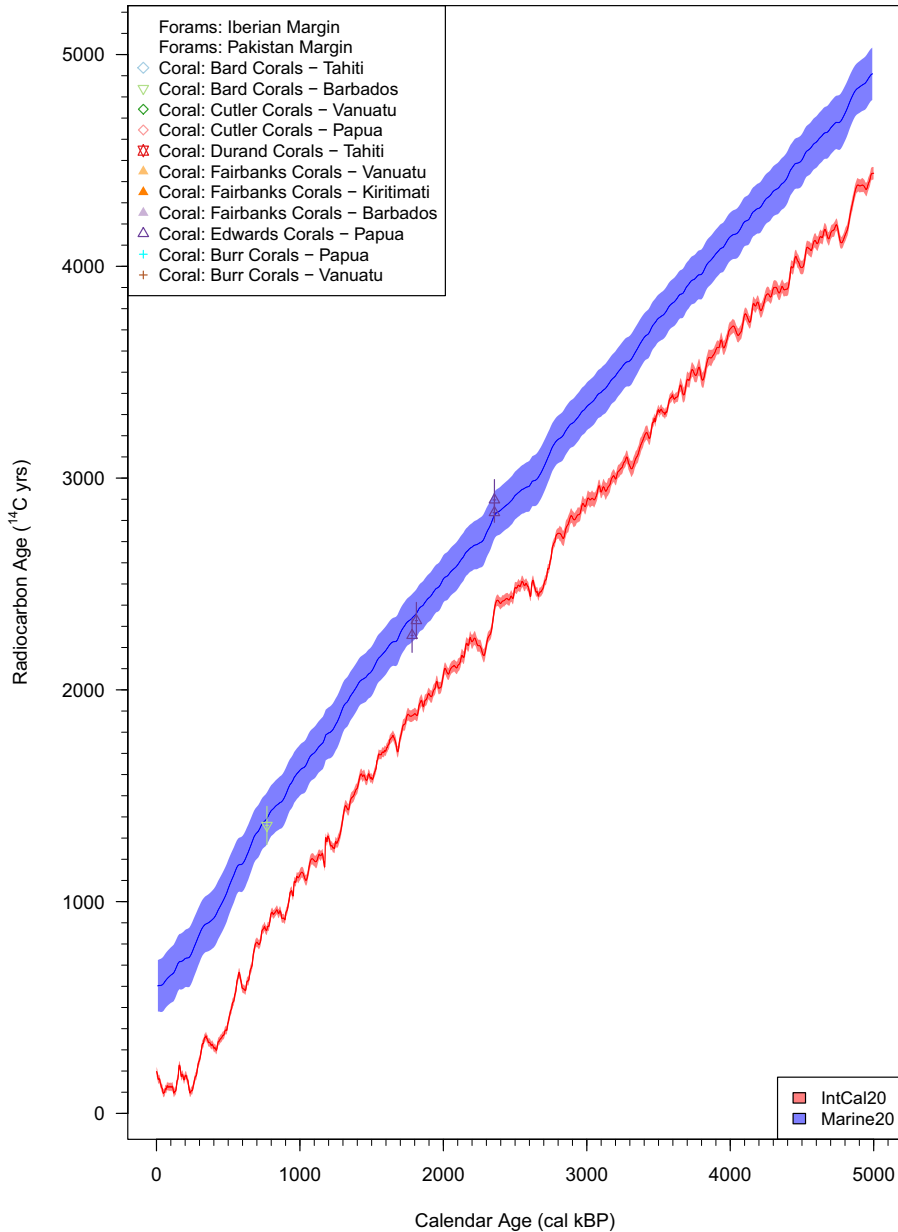


Figure 10 Plot of observed marine ^{14}C ages (with no MRA or ΔR correction) against the Marine20 and IntCal20 curves 0–55 cal kBP. The datasets shown consist of: foraminifera (forams) from the Iberian and Pakistan Margins (Bard et al. 2013); corals from Tahiti and Barbados (Bard et al. 1990, 1998, 2004); corals from Vanuatu and Papua New Guinea (Cutler et al. 2004); corals from Tahiti (Durand et al. 2013); corals from Vanuatu, Kiritimati and Barbados (Fairbanks et al. 2005); corals from Papua New Guinea (Edwards et al. 1993); and corals from Papua New Guinea and Vanuatu (Burr et al. 1998, 2004). Plotted are the 95% probability intervals for both the calendar age and radiocarbon age of each observation; together with the 95% probability/predictive intervals for Marine20 and IntCal20, respectively.

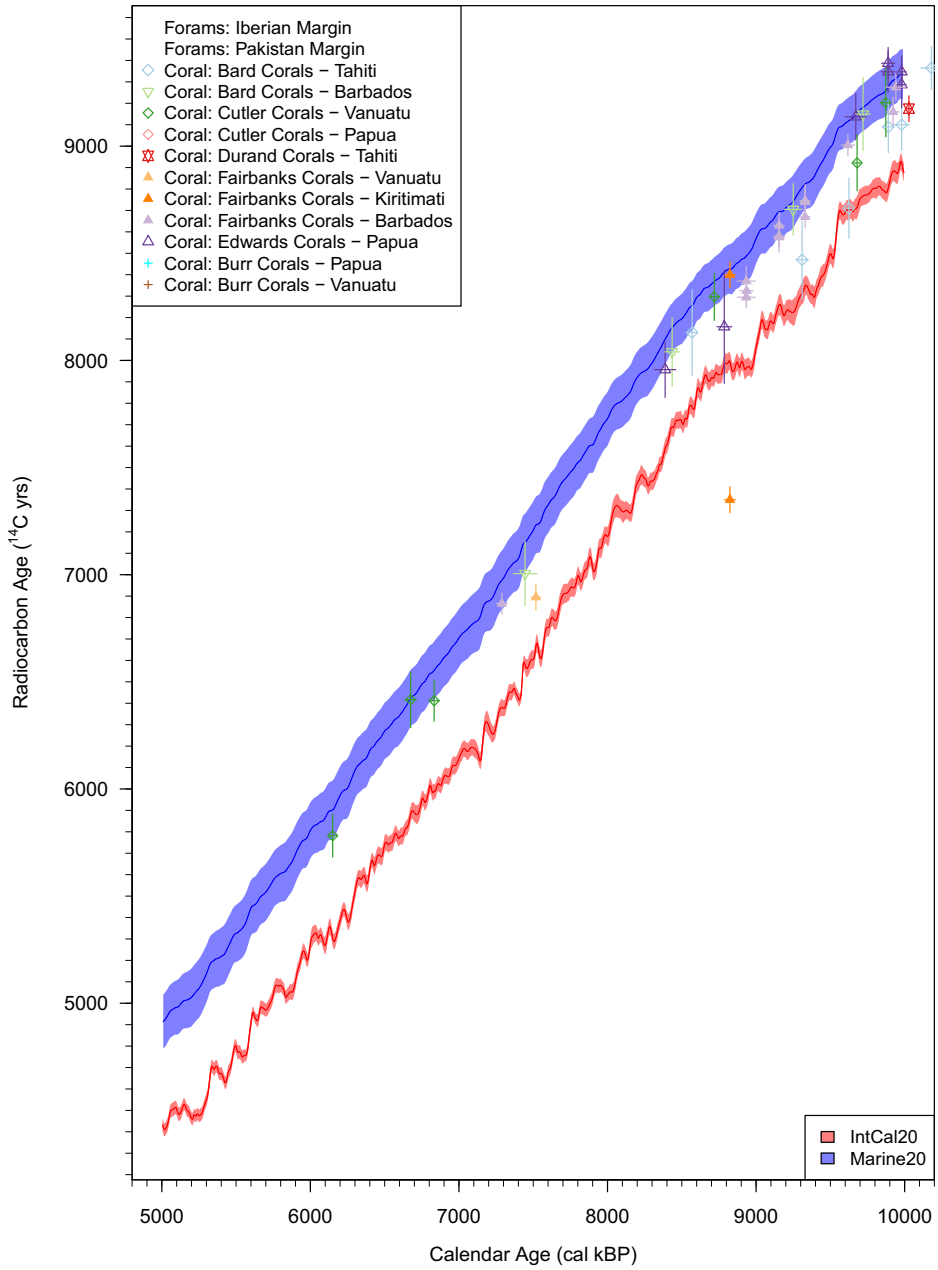


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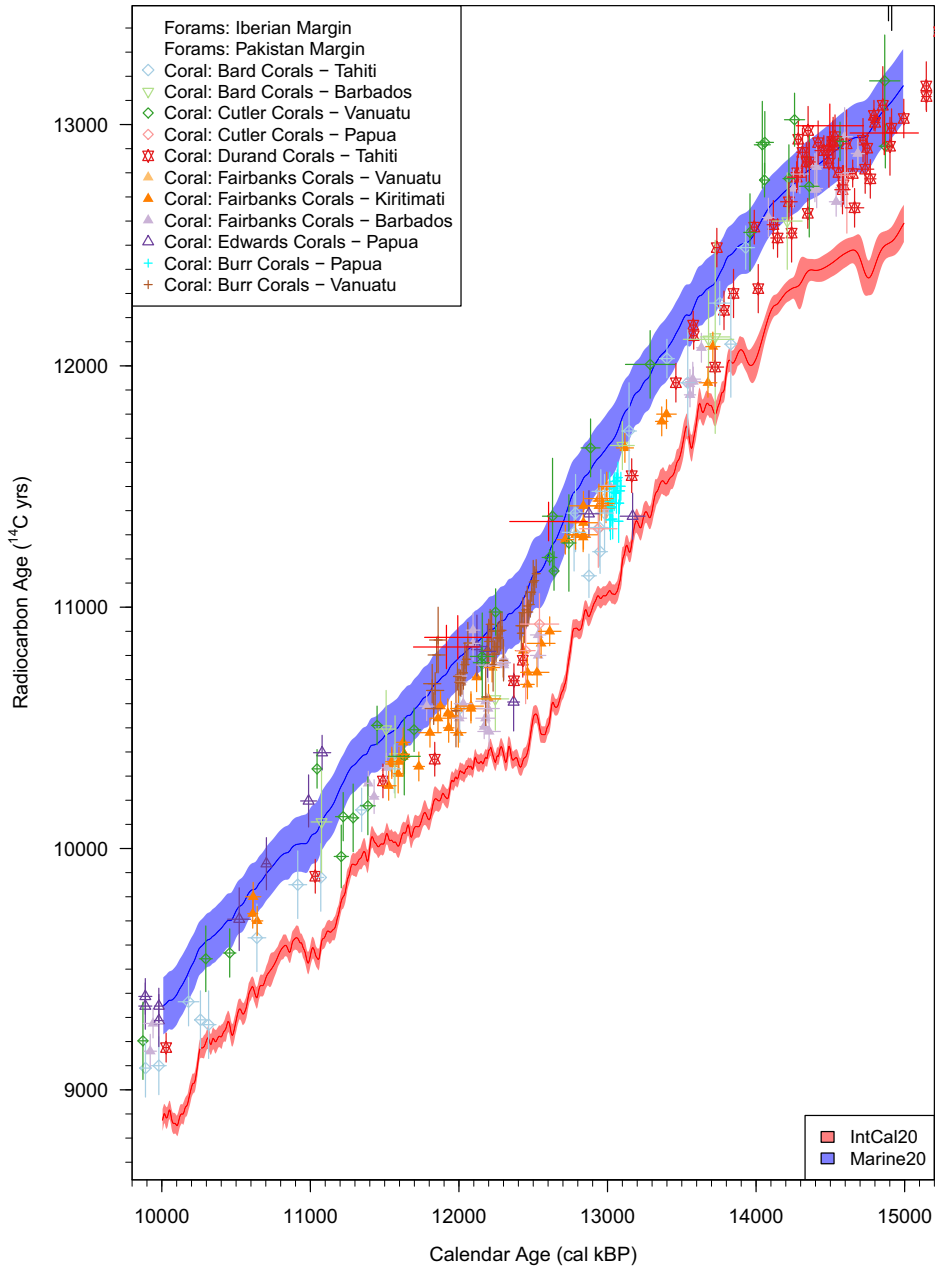


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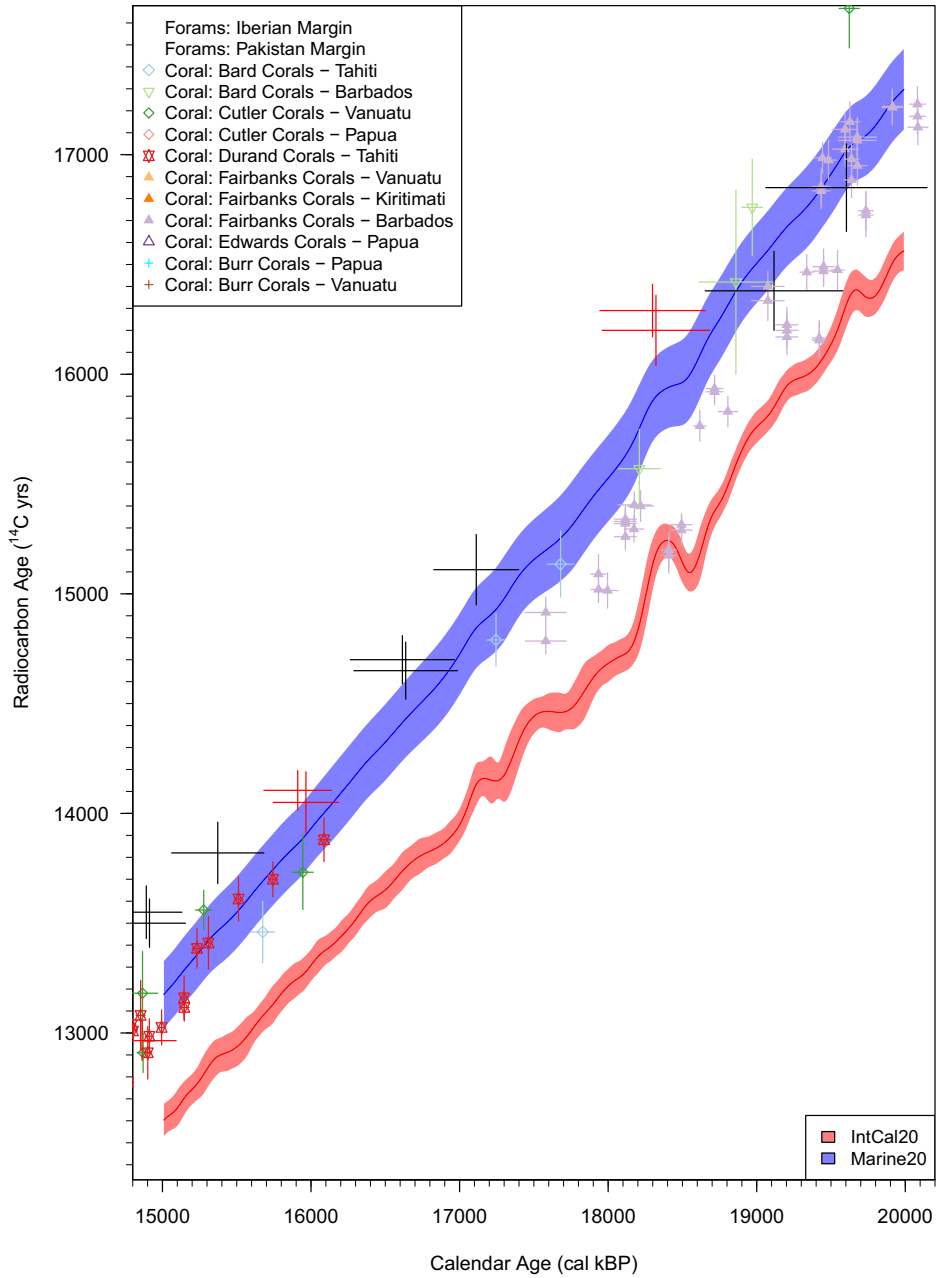


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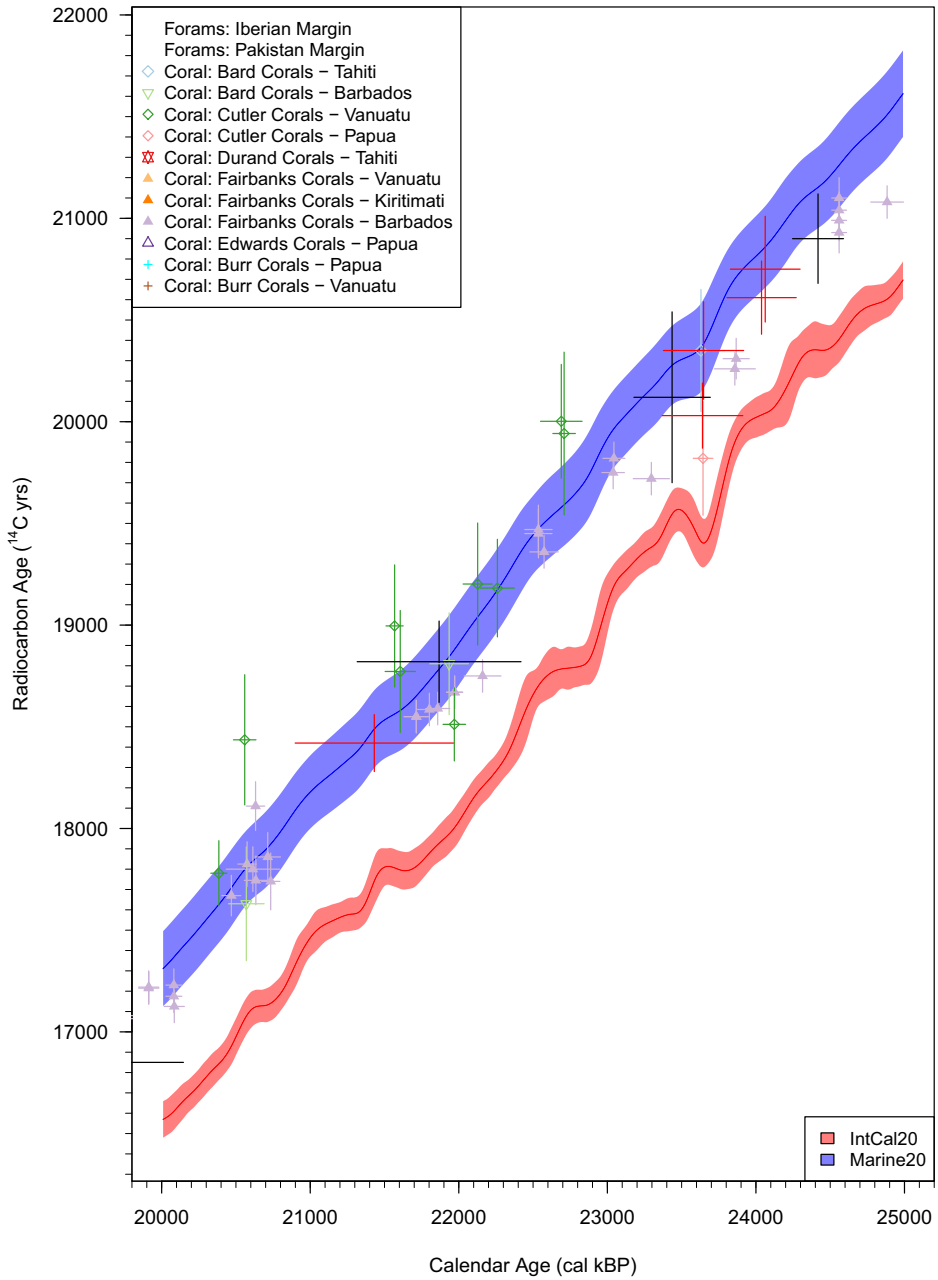


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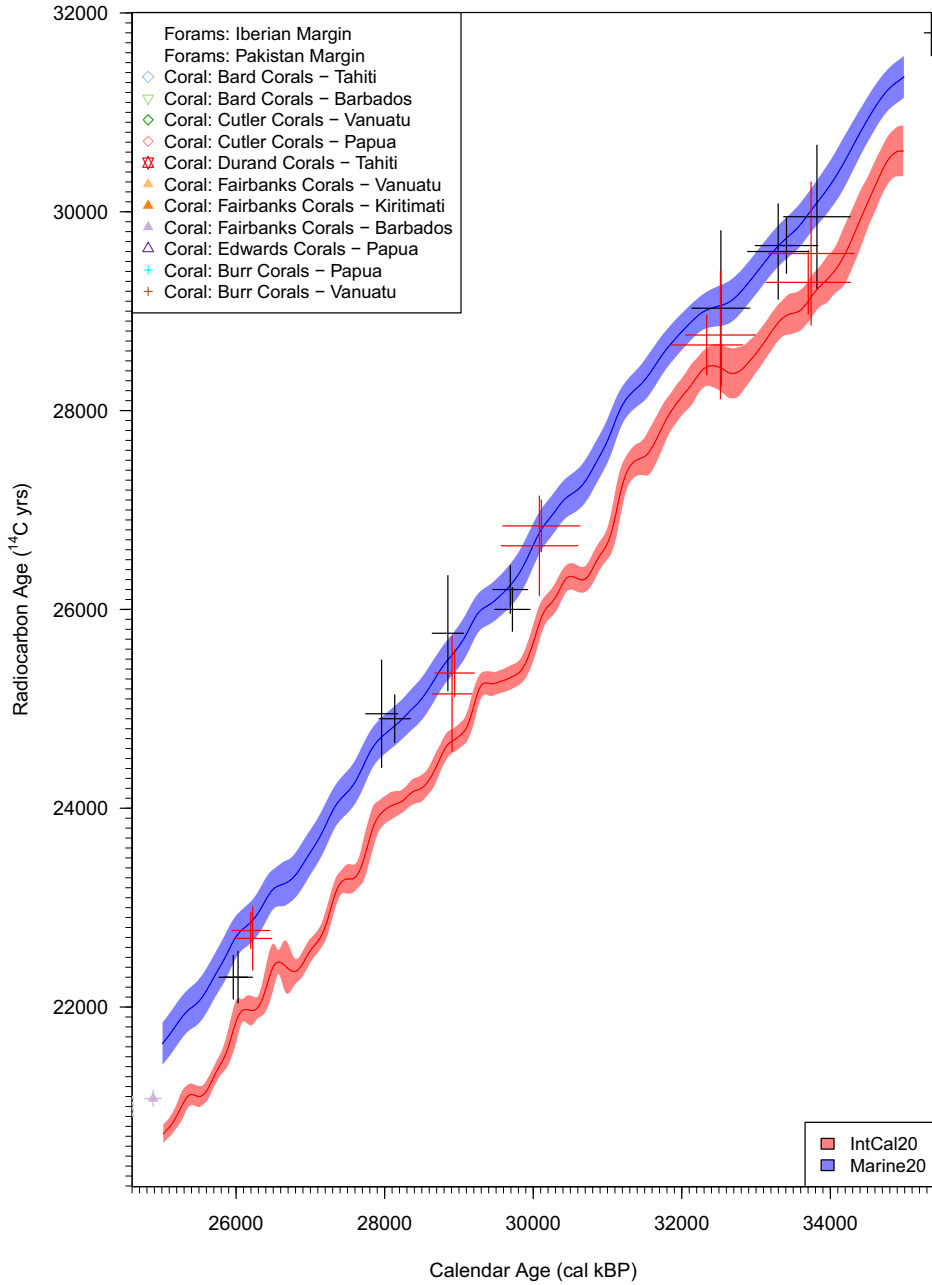


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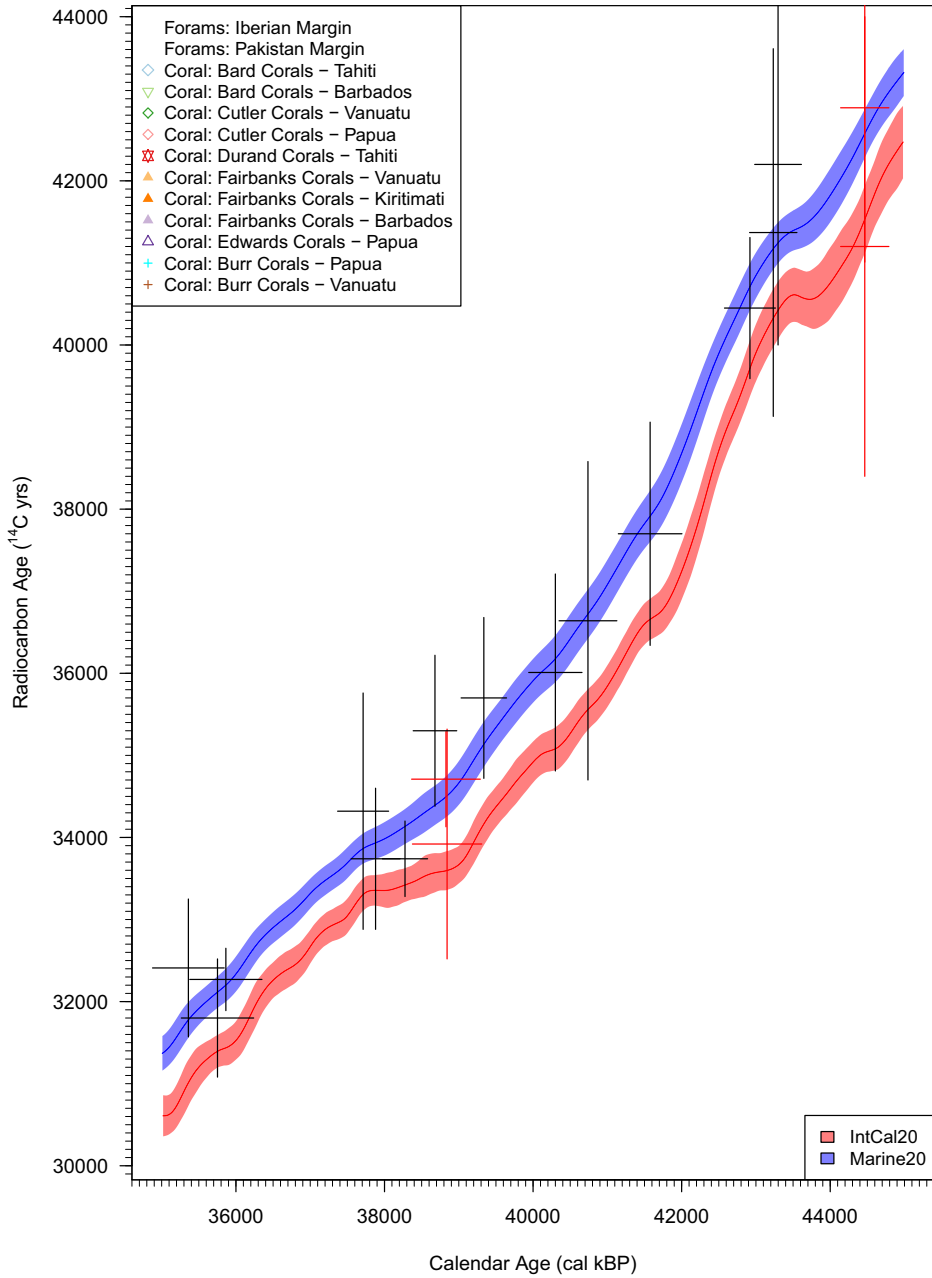


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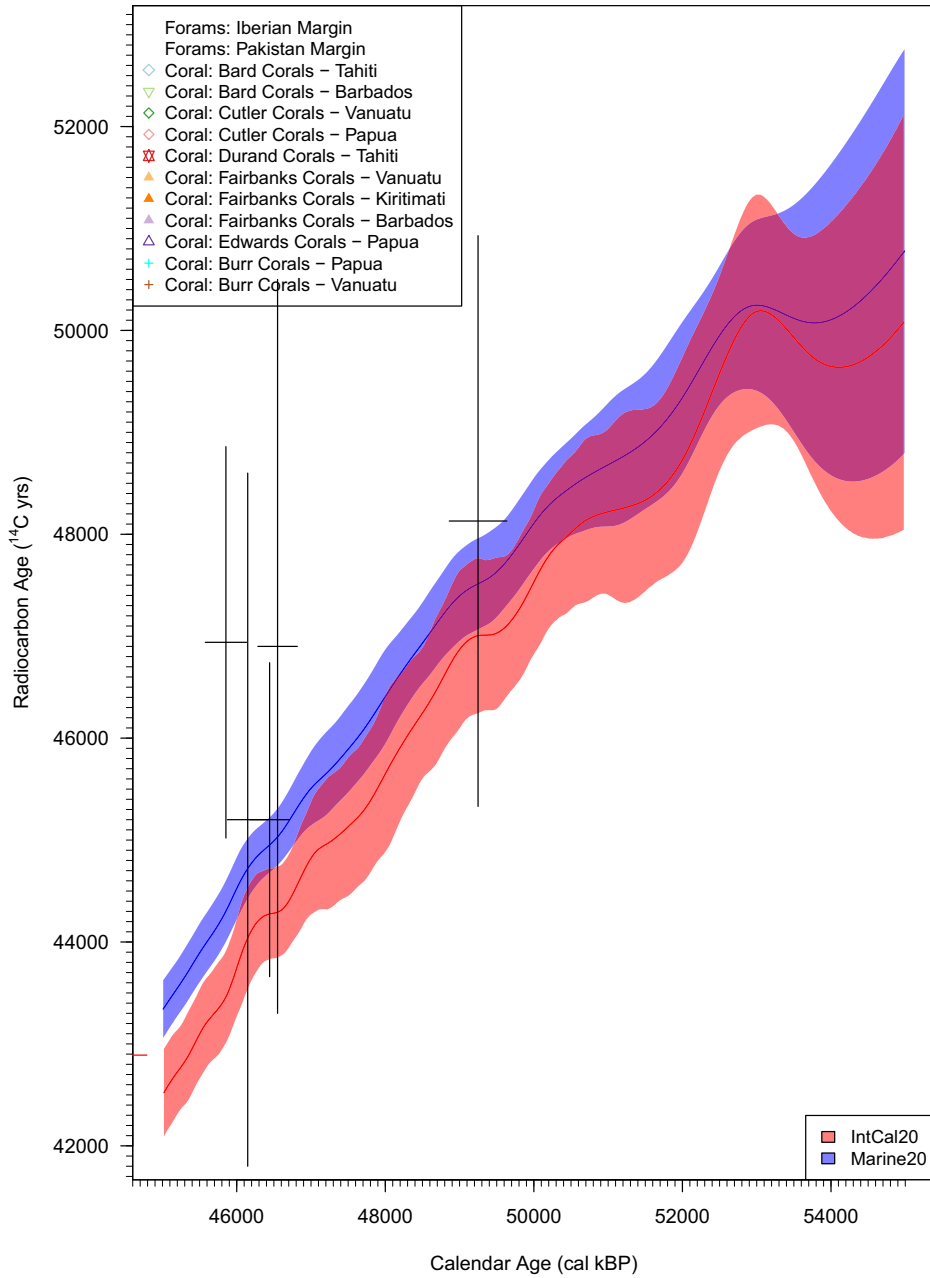


Figure 10 (continued)

and Heaton 2020 in this issue) appears to show reductions in MRA to values close to 0 ^{14}C yr during four distinct periods (ca. 12.8, 17, 23, and 31 cal kBP).

The most recent of these drops in MRA within the Cariaco Basin, at approximately 12.8 cal kBP, is evidenced by a reduction in the observed offset between ^{14}C samples of varved basin foraminifera and NH atmospheric ^{14}C tree-ring determinations. This drop occurs over a relatively short period of 200–300 calendar years. The older three drops (ca. 17, 23, and 31 cal kBP) potentially appear on longer, multi-millennia, scales. Due to the inability of our current carbon-cycle models to reproduce Cariaco's MRA, in order to include the Cariaco data within IntCal20, the basin's MRA beyond ca. 13.91 cal kBP was modeled as an additional flexible Bayesian spline simultaneously to the construction of the IntCal20 calibration curve (Heaton et al. 2020 in this issue; Hughen and Heaton 2020 in this issue). Intuitively with this approach, deviations in ^{14}C levels within Cariaco that were also seen in other atmosphere-adjusted IntCal20 records would likely be considered genuine atmospheric signal. Conversely, deviations seen in Cariaco ^{14}C levels alone would likely be modeled as changes in the basin's MRA. The consequent Cariaco MRA estimate from 13.91–55 cal kBP is therefore predominantly influenced by the offset between the observed Cariaco ^{14}C data and the Hulu Cave speleothems, for which we assumed that the dead carbon fraction (DCF) remains approximately constant.

The mechanism responsible for such MRA reductions is unknown and not seen in any of our model reproductions. Work to understand, and resolve, the differences between the estimate one obtains for the Cariaco Basin's MRA based upon observed data and that provided by carbon-cycle models would therefore be very valuable.

Regarding future potential areas of development, most directly, new records at key locations in the oceans would help to improve the resolution and accuracy of existing marine records. This might be further refined by integrating marine records for the sea surface, intermediate and deep-water masses through parallel studies of planktonic and benthic foraminifera, warm and cold water corals, studying and correcting the influence of signal perturbations (e.g. deep-sea sediment mixing, diagenesis of corals). Future work will benefit from recent analytical improvements such as the capacity to measure ^{14}C in individual shells of foraminifera (Lougheed et al. 2018; Fagault et al. 2019). Prescribing time varying piston velocity and AMOC values estimated from paleoceanographic and paleoclimatic records, in particular over the glacial-interglacial/stadial-interstadial climate system fluctuations, instead of using the present two-level representation may provide further insights.

Additionally, surface water ^{14}C reconstructions typically originate from continental margins, marginal seas, or tropical lagoons and are highly disperse dependent upon location. This is a consequence of regional variations in ocean circulation and atmospheric exchange. Ultimately, one would wish to develop location-specific (regional) marine age calibration curves. In particular we would wish to construct high-latitude marine calibration curves to cover the polar regions for which the current Marine20 is not suited. There are two potential approaches one could take to achieve regional calibration curves:

Firstly, one could aim to obtain sufficient data from individual locations in order to construct regional empirical curves. This would require considerable data, and it may be difficult to achieve sufficiently global coverage, but such work has begun in the Atlantic (Muschitiello et al. 2019; Brendryen et al. 2019; Skinner et al. 2019; Waelbroeck et al. 2019; Burke et al. [submitted](#)) and globally (e.g. Sarnthein et al. 2015).

Alternatively, the next generation of ^{14}C -equipped OGCMs might be used to provide location-specific reservoir ages, either through nesting approaches or by means of global multi-resolution models with unstructured meshes. First unpublished MRA simulations utilizing the OGCM FESOM2 with enhanced resolution down to ~ 20 km in marginal seas (Danilov et al. 2017) show promising improvements for the Cariaco region. However, for the time being, long-term simulations, which would be necessary for a potential application of FESOM2 within the radiocarbon age calibration context presented here, are not yet available.

To use such OGCMs appropriately, more detailed investigations of their properties and inherent uncertainties would be equally valuable, perhaps through the creation of statistical emulators (Sacks et al. 1989). This is complicated by the high-dimensional functional output of these complex OGCMs. Two potential approaches in such situations are the use of basis functions (Higdon et al. 2008) to emulate the complete OGCM output; or perhaps more practically to emulate the deviations of the complex OGCM from a simpler, approximate, but fast model such as BICYCLE (Kennedy and O'Hagan 2000).

Such emulators would enable us to identify the key parameters which influence model output, and equally which are less critical. Work could then be directed at narrowing down our uncertainties on these specific key values. While we performed a first step towards identifying these processes, we did not consider the contribution of potential interactions between inputs. In the case of scalar model output, Oakley and O'Hagan (2004) provide a method to do so although this would need modification to account for our functional and high-dimensional output. Further, such emulators may aid in tuning (also known as calibration within the statistical community) our OGCMs to synthesize their predictions with our observations (Kennedy and O'Hagan 2001).

In the case of the production of high-latitude marine calibration curves, one may consider production of separate curves for each of the Southern Ocean, North-Atlantic and North Pacific. By combining paleoceanographic proxy information from deep sea core records located within these ocean basins with suitable computer modeling, one could potentially not only obtain such curves but also tune and improve our carbon cycle models. Such work is however likely to require improvements in our understanding of such proxies and their links with winds, sea-ice and high-latitude upwelling.

If the next generation of marine radiocarbon age calibration curve still relies on the application of the BICYCLE carbon cycle box model some emphasis might be necessary on the inclusion of uncertainties in all time-dependent forcing, including a revision of the forcing itself. In particular, further constraints on air-sea gas exchange (piston velocity), which has a large effect on the marine ^{14}C reservoir age, will be essential. In a similar vein, the current models neglect millennial-scale changes in the Atlantic meridional circulation. While this could in principle be incorporated to address more accurately reconstructed changes in the ocean circulation, sensitivity tests have shown that the simulated carbon cycle response to such rather abrupt changes is very model dependent (Köhler et al. 2006).

8. CONCLUSIONS

In this paper we have presented an overview of the construction of the Marine20 global-average marine radiocarbon age calibration curve from 0–55 cal kBP. In this respect “global” is restricted to non-polar latitudes ranging approximately from 40/50°N to 40°S, i.e. excluding areas with sea ice. The curve is constructed using the BICYCLE carbon cycle

box model driven by estimates of atmospheric $\Delta^{14}\text{C}$ provided by IntCal20, by atmospheric CO_2 from ice cores, and by various other records changing the climatic boundary conditions. Together, these forcing time series enable us to incorporate past changes in both the carbon cycle and ^{14}C production rates into our model-based reconstruction of Marine20. Uncertainties in model parameters are propagated through the BICYCLE model by creating an ensemble of 500 simulations which are summarized by Monte-Carlo statistics. The new curve is seen to fit well with observed data for the pre-bomb period and agrees with individual simulations of a more complex model (LSG OGCM). For a calibration user, it is key to use the accompanying updated local-reservoir ΔR adjustments, found at <http://calib.org/marine/>, before using the new Marine20 curve.

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AUTHOR CONTRIBUTIONS

TJH led this study and performed the Bayesian modeling. The general setup of Marine20 has been designed through discussions of the IntCal Marine Focus group led by EB. PK provided simulation results using the BICYCLE model, MB simulation results from the LSG OGCM. The manuscript has been written by TJH with large contributions from PK. RWR performed initial modeling tests incorporating pCO_2 data and recalculated ΔR values in the marine reservoir correction database with Marine20. All other co-authors provided expert knowledge and contributions to both discussions and the writing of the manuscript.

REFERENCES

- Ahn J, Brook EJ. 2014. Siple Dome ice reveals two modes of millennial CO_2 change during the last ice age. *Nature Communications* 5:3723. doi: [10.1038/ncomms4723](https://doi.org/10.1038/ncomms4723).
- Ahn J, Brook EJ, Mitchell L, Rosen J, McConnell JR, Taylor K, Etheridge D, Rubino M. 2012. Atmospheric CO_2 over the last 1000 years: A high-resolution record from the West Antarctic Ice Sheet (WAIS) Divide ice core. *Global Biogeochemical Cycles* 26:GB2027. doi: [10.1029/2011GB004247](https://doi.org/10.1029/2011GB004247).
- Alves EQ, Macario KD, Urrutia FP, Cardoso RP, Bronk Ramsey C. 2019. Accounting for the marine reservoir effect in radiocarbon calibration. *Quaternary Science Reviews* 209: 129–138. doi: [10.1016/j.quascirev.2019.02.013](https://doi.org/10.1016/j.quascirev.2019.02.013).
- Arnold JR. 1957. The distribution of carbon-14 in nature. *Tellus* 9:28–32.
- Bard E. 1988. Correction of accelerator mass spectrometry ^{14}C ages measured in planktonic foraminifera: Paleoceanographic implications. *Paleoceanography* 3(6):635–645. doi: [10.1029/PA003i006p00635](https://doi.org/10.1029/PA003i006p00635).
- Bard E, Hamelin B, Fairbanks RG, Zindler A. 1990. Calibration of the ^{14}C timescale over the past 30,000 years using mass spectrometric U-Th ages from Barbados corals. *Nature* 345:405–410. doi: [10.1038/345405a0](https://doi.org/10.1038/345405a0).
- Bard E, Arnold M, Hamelin B, Tisnerat-Laborde N, Cabioch G. 1998. Radiocarbon calibration by means of mass spectrometric $^{230}\text{Th}/^{234}\text{U}$ and ^{14}C ages of corals. An updated data base including

- samples from Barbados, Mururoa and Tahiti. *Radiocarbon* 40(3):1085–1092. doi: [10.1017/S0033822200019135](https://doi.org/10.1017/S0033822200019135).
- Bard E, Ménot-Combes G, Rostek F. 2004. Present status of radiocarbon calibration and comparison records based on Polynesian corals and Iberian Margin sediments. *Radiocarbon* 46(3):1189–1202. doi: [10.1017/S0033822200033087](https://doi.org/10.1017/S0033822200033087).
- Bard E, Ménot G, Rostek F, Licari L, Böning P, Edwards RL, Cheng H, Wang Y, Heaton TJ. 2013. Radiocarbon calibration/comparison records based on marine sediments from the Pakistan and Iberian margins. *Radiocarbon* 55(4):1999–2019. doi: [10.2458/azu_js_rc.55.17114](https://doi.org/10.2458/azu_js_rc.55.17114).
- Barker S, Diz, P, Vantravers MJ, Pike J, Knorr G, Hall IR, Broecker WS. 2009. Interhemispheric Atlantic seesaw response during the last deglaciation. *Nature*, 457:1097–1102. doi: [10.1038/nature07770](https://doi.org/10.1038/nature07770).
- Bauska TK, Joos F, Mix AC, Roth R, Ahn J, Brook EJ. 2015. Links between atmospheric carbon dioxide, the land carbon reservoir and climate over the past millennium. *Nature Geosci*, 8:383–387. doi: [10.1038/ngeo2422](https://doi.org/10.1038/ngeo2422).
- Bereiter B, Lüthi D, Siegrist M, Schüpbach S, Stocker TF, Fischer H. 2012. Mode change of millennial CO₂ variability during the last glacial cycle associated with a bipolar marine carbon seesaw. *Proceedings of the National Academy of Sciences*, 109(25):9755–9760. doi: [10.1073/pnas.1204069109](https://doi.org/10.1073/pnas.1204069109).
- Borenstein M, Hedges LV, Higgins JPT, Rothstein HR. 2011. *Introduction to meta-analysis*. John Wiley & Sons.
- Brendryen J, Hafliðason H, Yokoyama Y, Haaga KA, Hannisdal B. 2019. Collapse of Eurasian ice sheets 14,600 years ago was a major source of global Meltwater Pulse 1a. *EarthArXiv*. 21 August 2019. doi: [10.31222/osf.io/7g5bn](https://doi.org/10.31222/osf.io/7g5bn).
- Broecker WS, Peng TH. 1987. The role of CaCO₃ compensation in the glacial to interglacial atmospheric CO₂ change. *Global Biogeochemical Cycles* 1(1):15–29. doi: [10.1029/GB001i001p00015](https://doi.org/10.1029/GB001i001p00015).
- Burke A, Rae J, Greenop R, Reimer PJ, Heaton TJ. (submitted). North Atlantic radiocarbon constraints on ocean circulation over the last deglaciation.
- Burr GS, Beck JW, Taylor FW, Recy J, Edwards RL, Cabioch G, Corregge T, Donahue DJ, and O'Malley JM. 1998. A high-resolution radiocarbon calibration between 11,700 and 12,400 calendar years BP derived from ²³⁰Th ages of corals from Espiritu Santo Island, Vanuatu. *Radiocarbon* 40(3):1093–1105. doi: [10.1017/S0033822200019147](https://doi.org/10.1017/S0033822200019147).
- Burr GS, Galang C, Taylor FW, Gallup CD, Edwards RL, Cutler KB, Quirk B. 2004. Radiocarbon results from a 13ka BP coral from the Huon Peninsula, Papua New Guinea. *Radiocarbon* 46(3):1211–1224.
- Butzin M, Köhler P, Heaton TJ, Lohmann G. 2020. A short note on marine reservoir age simulations used in IntCal20. *Radiocarbon* 62. This issue. doi: [10.1017/RDC.2020.9](https://doi.org/10.1017/RDC.2020.9).
- Butzin M, Köhler P, Lohmann G. 2017. Marine radiocarbon reservoir age simulations for the past 50,000 years. *Geophysical Research Letters* 44:8473–8480. doi: [10.1002/2017GL074688](https://doi.org/10.1002/2017GL074688).
- Butzin M, Prange M, Lohmann G. 2005. Radiocarbon simulations for the glacial ocean: The effects of wind stress, Southern Ocean sea ice and Heinrich events. *Earth and Planetary Science Letters* 235(1–2):45–61. doi: [10.1016/j.epsl.2005.03.003](https://doi.org/10.1016/j.epsl.2005.03.003).
- Butzin M, Prange M, Lohmann G. 2012. Readjustment of glacial radiocarbon chronologies by self-consistent three-dimensional ocean circulation modeling. *Earth and Planetary Science Letters* 317–318:177–184. doi: [10.1016/j.epsl.2011.11.046](https://doi.org/10.1016/j.epsl.2011.11.046).
- Cheng H, Edwards RL, Southon J, Matsumoto K, Feinberg JM, Sinha A, Zhou W, Li H, Li X, Xu Y. 2018. Atmospheric ¹⁴C/¹²C changes during the last glacial period from Hulu Cave. *Science*, 362(6420):1293–1297. doi: [10.1126/science.aau0747](https://doi.org/10.1126/science.aau0747).
- Craig H. 1957. The natural distribution of radiocarbon and the exchange time of carbon dioxide between the atmosphere and sea. *Tellus* 9:1–17.
- Cutler KB, Gray SC, Burr GS, Edwards RL, Taylor FW, Cabioch G, Beck JW, Cheng H, Moore J. 2004. Radiocarbon calibration to 50 kyr BP with paired ¹⁴C and ²³⁰Th dating of corals from Vanuatu and Papua New Guinea. *Radiocarbon* 46(3):1127–1160. doi: [10.1017/S0033822200033063](https://doi.org/10.1017/S0033822200033063).
- Danilov S, Sidorenko D, Wang Q, Jung T. 2017. The Finite-volume Sea ice–Ocean Model (FESOM2). *Geoscientific Model Development* 10:765–789. doi: [10.5194/gmd-10-765-2017](https://doi.org/10.5194/gmd-10-765-2017).
- Durand N, Deschamps P, Bard E, Hamelin B, Camoin G, Thomas A, Henderson G, Yokoyama Y, Matsuzaki H. 2013. Comparison of ¹⁴C and U-Th ages in corals from IODP #310 cores offshore Tahiti. *Radiocarbon* 55(4):1947–1974. doi: [10.2458/azu_js_rc.v55i2.16134](https://doi.org/10.2458/azu_js_rc.v55i2.16134).
- Edwards RL, Beck JW, Burr GS, Donahue DJ, Chappell JMA, Bloom AL, Druffel ERM, Taylor FW. 1993. A large drop in atmospheric ¹⁴C/¹²C and reduced melting in the Younger Dryas, documented with ²³⁰Th ages of corals. *Science* 260(5110):962–968. doi: [10.1126/science.260.5110.962](https://doi.org/10.1126/science.260.5110.962).
- Fagault Y, Tuna T, Rostek F, Bard E. 2019. Radiocarbon dating small carbonate samples with the gas ion source of AixMICADAS. *Nuclear Instruments and Methods in Physics Research B* 455:276–283. doi: [10.1016/j.nimb.2018.11.018](https://doi.org/10.1016/j.nimb.2018.11.018).

- Fairbanks RG, Mortlock RA, Chiu TC, Cao L, Kaplan A, Guilderson TP, Fairbanks TW, Bloom AL, Grootes PM, Nadeau MJ. 2005. Radiocarbon calibration curve spanning 0 to 50,000 years BP based on paired Th-230/U-234/U-238 and C-14 dates on pristine corals. *Quaternary Science Reviews* 24(16–17):1781–1796. doi: [10.1016/j.quascirev.2005.04.007](https://doi.org/10.1016/j.quascirev.2005.04.007).
- Heaton TJ, Blaauw M, Blackwell PG, Bronk Ramsey C, Reimer PJ, Scott EM. 2020. The IntCal20 approach to radiocarbon calibration curve construction: a new methodology using Bayesian splines and errors-in-variables. *Radiocarbon* 62. This issue. doi: [10.1017/RDC.2020.46](https://doi.org/10.1017/RDC.2020.46).
- Heimann M, Monfray P. 1989. Spatial and temporal variation of the gas exchange coefficient for CO₂: 1. Data analysis and global validation. Report 31, Max-Planck-Inst. for Meteorol., Hamburg, Germany.
- Higdon D, Gattiker J, Williams B, Rightley M. 2008. Computer model calibration using high-dimensional output. *Journal of the American Statistical Association* 103(482):570–583. doi: [10.1198/016214507000000888](https://doi.org/10.1198/016214507000000888).
- Hughen KA, Baillie M, Bard E, Beck J, Bertrand C, Blackwell PG, Buck CE, Burr G, Cutler K, Damon P, Edwards R, Fairbanks R, Friedrich M, Guilderson T, Kromer B, McCormac G, Manning S, Bronk Ramsey C, Reimer PJ, Reimer R, Remmele S, Southon J, Stuiver M, Talamo S, Taylor F, van der Plicht J, Weyhenmeyer C. 2004. Marine04 marine radiocarbon age calibration, 0–26 cal kyr BP. *Radiocarbon* 46(3):1059–1086.
- Hughen KA, Heaton TJ. 2020. Updated Cariaco Basin ¹⁴C calibration dataset from 0–60 cal kyr BP. *Radiocarbon* 62. This issue. doi: [10.1017/RDC.2020.53](https://doi.org/10.1017/RDC.2020.53).
- Keller DP, Lenton A, Scott V, Vaughan NE, Bauer N, Ji D, Jones CD, Kravitz B, Muri H, Zickfeld K. 2018. The Carbon Dioxide Removal Model Intercomparison Project (CDRMIP): rationale and experimental protocol for CMIP6. *Geoscientific Model Development* 11:1133–1160. doi: [10.5194/gmd-11-1133-2018](https://doi.org/10.5194/gmd-11-1133-2018).
- Kennedy MC, O'Hagan A. 2000. Predicting the output from a complex computer code when fast approximations are available. *Biometrika* 87(1):1–13.
- Kennedy MC, O'Hagan A. 2001. Bayesian calibration of computer models. *Journal of the Royal Statistical Society: Series B (Statistical Methodology)* 63:425–464. doi: [10.1111/1467-9868.00294](https://doi.org/10.1111/1467-9868.00294).
- Key RM, Kozyr A, Sabine CL, Lee K, Wanninkhof R, Bullister JL, Feely RA, Millero FJ, Mordy C, Peng T-H. 2004. A global ocean carbon climatology: Results from Global Data Analysis Project (GLODAP). *Global Biogeochemical Cycles* 18: GB4031. doi: [10.1029/2004GB002247](https://doi.org/10.1029/2004GB002247).
- Köhler P. 2016. Using the Suess effect on the stable carbon isotope to distinguish the future from the past in radiocarbon. *Environmental Research Letters* 11:124016. doi: [10.1088/1748-9326/11/12/124016](https://doi.org/10.1088/1748-9326/11/12/124016).
- Köhler P, Fischer H, Munhoven G, Zeebe RE. 2005. Quantitative interpretation of atmospheric carbon records over the last glacial termination. *Global Biogeochemical Cycles* 19: GB4020. doi: [10.1029/2004GB002345](https://doi.org/10.1029/2004GB002345).
- Köhler P, Fischer H, Schmitt J. 2010. Atmospheric δ¹³CO₂ and its relation to pCO₂ and deep ocean δ¹³C during the late Pleistocene. *Paleoceanography* 25:PA1213. doi: [10.1029/2008PA001703](https://doi.org/10.1029/2008PA001703).
- Köhler P, Muscheler R, Fischer H. 2006. A model-based interpretation of low-frequency changes in the carbon cycle during the last 120,000 years and its implications for the reconstruction of atmospheric Δ¹⁴C. *Geochemistry Geophysics Geosystems* 7:Q11N06. doi: [10.1029/2005GC001228](https://doi.org/10.1029/2005GC001228).
- Köhler P, Nehrass-Ahles C, Schmitt J, Stocker TF, Fischer H. 2017. A 156 kyr smoothed history of the atmospheric greenhouse gases CO₂, CH₄, and N₂O and their radiative forcing. *Earth System Science Data* 9:363–387. doi: [10.5194/essd-9-363-2017](https://doi.org/10.5194/essd-9-363-2017).
- Köhler P, Fischer H. 2004. Simulating changes in the terrestrial biosphere during the last glacial/interglacial transition. *Global Planet. Change* 43(1–2):33–55. doi: [10.1016/j.gloplacha.2004.02.005](https://doi.org/10.1016/j.gloplacha.2004.02.005).
- Köhler P, Fischer H. 2006. Simulating low frequency changes in atmospheric CO₂ during the last 740 000 years. *Climate of the Past* 2:57–78. doi: [10.5194/cp-2-57-2006](https://doi.org/10.5194/cp-2-57-2006).
- Köhler P, Knorr G, Bard E. 2014. Permafrost thawing as source of abrupt carbon release at the onset of the Bølling/Allerød. *Nature Communications* 5520. doi: [10.1038/ncomms6520](https://doi.org/10.1038/ncomms6520).
- Krakauer NY, Randerson JT, Primeau FW, Gruber N, Menemenlis D. 2006. Carbon isotope evidence for the latitudinal distribution and wind speed dependence of the air-sea gas transfer velocity. *Tellus B*:58:390–417. doi: [10.1111/j.1600-0889.2006.00223.x](https://doi.org/10.1111/j.1600-0889.2006.00223.x).
- Lasca I, Feinberg JM, Dorale JA, Cheng H, Edwards RL. 2016. Age of the Laschamp excursion determined by U-Th dating of a speleothem geomagnetic record from North America. *Geology* 44(2):139–142. doi: [10.1130/G37490.1](https://doi.org/10.1130/G37490.1).
- Lougheed BC, Metcalfe B, Ninnemann US, Wacker L. 2018. Moving beyond the age-depth model paradigm in deep-sea palaeoclimate archives: dual radiocarbon and stable isotope analysis on single foraminifera. *Climate of the Past* 14:515–526. doi: [10.5194/cp-14-515-2018](https://doi.org/10.5194/cp-14-515-2018).
- Lüthi D, Bereiter B, Stauffer B, Winkler R, Schwander J, Kindler P, Leuenberger M, Kipfstuhl S, Capron E, Landais A, Fischer H, Stocker TF. 2010. CO₂ and O₂/N₂ variations in

- and just below the bubble-clathrate transformation zone of Antarctic ice cores. *Earth and Planetary Science Letters* 297(1–2): 226–233. doi: [10.1016/j.epsl.2010.06.023](https://doi.org/10.1016/j.epsl.2010.06.023).
- MacFarling-Meure C, Etheridge D, Trudinger C, Langenfelds R, van Ommen T, Smith A, Elkins J. 2006. Law Dome CO₂, CH₄ and N₂O ice core records extended to 2000 years BP. *Geophysical Research Letters* 33:L14810. doi: [10.1029/2006GL026152](https://doi.org/10.1029/2006GL026152).
- Marcott SA, Bauska TK, Buizert C, Steig EJ, Rosen JL, Cuffey KM, Fudge TJ, Severinghaus JP, Ahn J, Kalk ML, McConnell JR, Sowers T, Taylor KC, White JW, Brook EJ. 2014. Centennial scale changes in the global carbon cycle during the last deglaciation. *Nature* 514:616–619. doi: [10.1038/nature13799](https://doi.org/10.1038/nature13799).
- McIntyre A, Lamont-Doherty Geological Observatory, Geological Society of America. 1981. Seasonal reconstructions of the Earth's surface at the Last Glacial Maximum. Boulder (CO): Geological Society of America.
- Monnin E, Indermühle A, Dällenbach A, Flückiger J, Stauffer B, Stocker TF, Raynaud D, Barnola J-M. 2001. Atmospheric CO₂ concentrations over the last glacial termination. *Science* 291(5501):112–114. doi: [10.1126/science.291.5501.112](https://doi.org/10.1126/science.291.5501.112).
- Monnin E, Steig EJ, Siegenthaler U, Kawamura K, Schwander J, Stauffer B, Stocker TF, Morse DL, Barnola J-M, Bellier B, Raynaud D, Fischer H. 2004. Evidence for substantial accumulation rate variability in Antarctica during the Holocene, through synchronization of CO₂ in the Taylor Dome, Dome C and DML ice cores. *Earth and Planetary Science Letters* 224(1–2):45–54. doi: [10.1016/j.epsl.2004.05.007](https://doi.org/10.1016/j.epsl.2004.05.007).
- Müller SA, Joos F, Plattner G-K, Edwards NR, Stocker TF. 2008. Modelled natural and excess radiocarbon: Sensitivities to the gas exchange formulation and ocean transport strength. *Global Biogeochemical Cycles* 22:GB3011. doi: [10.1029/2007GB003065](https://doi.org/10.1029/2007GB003065).
- Muschitiello F, D'Andrea WJ, Schmittner A, Heaton TJ, Balascio NL, DeRoberts N, Caffee MW, Woodruff TE, Welten KC, Skinner LC, Simon MH, Dokken TM. 2019. Deep-water circulation changes lead North Atlantic climate during deglaciation. *Nat Commun*, 10:1272. doi: [10.1038/s41467-019-09237-3](https://doi.org/10.1038/s41467-019-09237-3).
- Naegler T, Ciais P, Rodgers K, Levin I. 2006. Excess radiocarbon constraints on air-sea gas exchange and the uptake of CO₂ by the oceans. *Geophysical Research Letters* 33:L11802. doi: [10.1029/2005GL025408](https://doi.org/10.1029/2005GL025408).
- Naegler T. 2009. Reconciliation of excess ¹⁴C-constrained global CO₂ piston velocity estimates. *Tellus B*, 61:372–384. doi: [10.1111/j.1600-0889.2008.00408.x](https://doi.org/10.1111/j.1600-0889.2008.00408.x).
- Oakley J, O'Hagan A. 2004. Probabilistic sensitivity analysis of complex models: a Bayesian approach. *Journal of the Royal Statistical Society Series B* 66:751–769. doi: [10.1111/j.1467-9868.2004.05304.x](https://doi.org/10.1111/j.1467-9868.2004.05304.x).
- Oeschger H, Siegenthaler U, Schotterer U, Gugelmann A. 1975. A box diffusion model to study the carbon dioxide exchange in nature. *Tellus* 27(2):168–92. doi: [10.3402/tellusa.v27i2.9900](https://doi.org/10.3402/tellusa.v27i2.9900).
- Press WH, Teukolsky SA, Vetterling WT, Flannery BP. 1992. Numerical recipes in Fortran, second edition, Cambridge University Press, Cambridge.
- Reimer PJ, Baillie MGL, Bard E, Bayliss A, Beck JW, Blackwell PG, Bronk Ramsey C, Buck CE, Burr GS, Edwards RL, Friedrich M, Grootes PM, Guilderson TP, Hajdas I, Heaton TJ, Hogg AG, Hughen KA, Kaiser KF, Kromer B, McCormac FG, Manning SW, Reimer RW, Richards DA, Southon JR, Talamo S, Turney CSM, van der Plicht J, Weyhenmeyer CE. 2009. IntCal09 and Marine09 radiocarbon age calibration curves, 0–50,000 years cal BP. *Radiocarbon* 51(4):1111–50.
- Reimer PJ, Bard E, Bayliss A, Beck JW, Blackwell PG, Bronk Ramsey C, Buck CE, Cheng H, Edwards RL, Friedrich M, Grootes PM, Guilderson TP, Hafliadason H, Hajdas I, Hatté C, Heaton TJ, Hoffmann DL, Hogg AG, Hughen KA, Kaiser KF, Kromer B, Manning SW, Niu M, Reimer RW, Richards DA, Scott EM, Southon JR, Staff RA, Turney CSM, van der Plicht J. 2013. IntCal13 and Marine13 radiocarbon age calibration curves 0–50,000 years cal BP. *Radiocarbon* 55(4):1869–87.
- Reimer PJ, Austin WEN, Bard E, Bayliss A, Blackwell PG, Bronk Ramsey C, Butzin M, Cheng H, Edwards RL, Friedrich M, Grootes PM, Guilderson TP, Hajdas I, Heaton TJ, Hogg AG, Hughen KA, Kromer B, Manning SW, Muscheler R, Palmer JG, Pearson C, van der Plicht J, Reimer RW, Richards DA, Scott EM, Southon JR, Turney CSM, Wacker L, Adophi F, Büntgen U, Capano M, Fahrni S, Fogtmann-Schulz A, Friedrich R, Kudsk S, Miyake F, Olsen J, Reinig F, Sakamoto M, Sookdeo A, Talamo S. 2020. The IntCal20 Northern Hemisphere radiocarbon calibration curve (0–55 cal kBP). *Radiocarbon* 62. This issue. doi: [10.1017/RDC.2020.41](https://doi.org/10.1017/RDC.2020.41).
- Reimer PJ, Reimer RW. 2001. A marine reservoir correction database and on-line interface. *Radiocarbon* 43(2A):461–463. doi: [10.1017/S0033822200038339](https://doi.org/10.1017/S0033822200038339).
- Revelle R, Suess H. 1957. Carbon dioxide exchange between atmosphere and ocean and the question of an increase of atmospheric CO₂ during the past decades. *Tellus* 9:18–27. doi: [10.1111/j.2153-3490.1957.tb01849.x](https://doi.org/10.1111/j.2153-3490.1957.tb01849.x).
- Rubino M, Etheridge DM, Trudinger CM, Allison CE, Battle MO, Langenfelds RL, Steele LP, Curran M, Bender M, White JWC, Jenk TM, Blunier T, Francey RJ. 2013. A revised 1000-

- year atmospheric $\delta^{13}\text{C-CO}_2$ record from Law Dome and South Pole, Antarctica. *Journal of Geophysical Research: Atmospheres* 118:8482–8499. doi: [10.1002/jgrd.50668](https://doi.org/10.1002/jgrd.50668).
- Sacks J, Welch WJ, Mitchell TJ, Wynn HP. 1989. Design and analysis of computer experiments. *Statistical Science* 4(4):409–423. doi: [10.1214/ss/1177012413](https://doi.org/10.1214/ss/1177012413).
- Sarnthein M, Gersonde R, Niebler S, Pflaumann U, Spielhagen R, Thiede J, Wefer G, Weinelt M. 2003. Overview of Glacial Atlantic Ocean Mapping (GLAMAP 2000). *Paleoceanography* 18:1030. doi: [10.1029/2002PA000769](https://doi.org/10.1029/2002PA000769).
- Sarnthein M, Balmer S, Grootes PM, Mudelsee M. 2015. Planktic and benthic ^{14}C reservoir ages for three ocean basins, calibrated by a suite of ^{14}C plateaus in the glacial-to-deglacial Suigetsu atmospheric ^{14}C record. *Radiocarbon* 57(1):129–151. doi: [10.2458/azu_rc.57.17916](https://doi.org/10.2458/azu_rc.57.17916).
- Skinner LC, Primeau F, Freeman E, de la Fuente M, Goodwin PA, Gottschalk J, Huang E, McCave IN, Noble TL, Scrivner, AE. 2017. Radiocarbon constraints on the glacial ocean circulation and its impact on atmospheric CO_2 . *Nature Communications* 8:16010. doi: [10.1038/ncomms16010](https://doi.org/10.1038/ncomms16010).
- Skinner LC, Muschitiello F, Scrivner, AE. 2019. Marine reservoir age variability over the last deglaciation: implications for marine carbon cycling and prospects for regional radiocarbon calibrations. *Paleoceanography and Paleoclimatology* 34:1807–1815. doi: [10.1029/2019PA003667](https://doi.org/10.1029/2019PA003667).
- Southon J, Noronha AL, Cheng H, Edwards RL, Wang YJ. 2012. A high-resolution record of atmospheric C-14 based on Hulu Cave speleothem H82. *Quaternary Science Reviews* 33:32–41. doi: [10.1016/j.quascirev.2011.11.022](https://doi.org/10.1016/j.quascirev.2011.11.022).
- Stuiver M, Pearson GW, Braziunas TF. 1986. Radiocarbon age calibration of marine samples back to 9000 cal yr BP. *Radiocarbon* 28(2B):980–1021. doi: [10.1017/S0033822200060264](https://doi.org/10.1017/S0033822200060264).
- Sweeney C, Gloor E, Jacobson AR, Key RM, McKinley G, Sarmiento JL, Wanninkhof R. 2007. Constraining global air-sea gas exchange for CO_2 with recent bomb ^{14}C measurements. *Global Biogeochemical Cycles* 21:GB2015. doi: [10.1029/2006GB002784](https://doi.org/10.1029/2006GB002784).
- Talley LD, Pickard GL, Emery WJ, Swift JH. 2011. *Descriptive physical oceanography: An Introduction*. 6th ed. Boston: Elsevier. 560 p.
- Toggweiler JR, Druffel ERM, Key RM, Galbraith ED. 2019. Upwelling in the ocean basins north of the ACC: 1. On the upwelling exposed by the surface distribution of $\Delta^{14}\text{C}$. *Journal of Geophysical Research: Oceans* 124(4):2591–2608. doi: [10.1029/2018JC014794](https://doi.org/10.1029/2018JC014794).
- Waelbroeck C, Lougheed BC, Vazquez Riveiros N, Missiaen L, Pedro J, Dokken T, Hajdas I, Wacker L, Abbott P, Dumoulin JP, Thil F, Eynaud F, Rossignol L, Fersi W, Albuquerque AL, Arz H, Austin W, Came R, Carlson A, Collins J, Dennielou B, Desprat S, Dickson A, Elliot M, Farmer C, Giraudeau J, Gottschalk J, Henderiks J, Hughen K, Jung S, Knutz P, Lebreiro S, Lund D, Lynch-Stieglitz J, Malaizé B, Marchitto T, Martinez Mendez G, Mollenhauer G, Naughton F, Nave S, Nünberg D, Oppo D, Peck V, Penaud A, Portillo RR, Repschläger J, Roberts J, Rühlemann C, Salgueiro E, Sanchez GM, Schönfeld J, Scussolini P, Skonieczny C, Thornalley D, Toucanne S, Van Rooij D, Vidal L, Voelker A, Wary M, Weldeab S, Ziegler M. 2019. Consistently dated Atlantic sediment cores over the last 40 thousand years. *SEANOE*. doi: [10.17882/59554](https://doi.org/10.17882/59554).
- Wanninkhof R. 1992. Relationship between wind speed and gas exchange over the ocean. *Journal of Geophysical Research* 97(C5):7373–7382. doi: [10.1029/92JC00188](https://doi.org/10.1029/92JC00188).
- Wanninkhof R. 2014. Relationship between wind speed and gas exchange over the ocean revisited. *Limnology and Oceanography: Methods* 12(6):351–362. doi: [10.4319/lom.2014.12.351](https://doi.org/10.4319/lom.2014.12.351).