We kindly thank all reviewers for their insightful criticism that helped us to improve this manuscript. Below we reply to the review comments one by one. The review comments are shown in grey, our reply in black. Applied changes to the manuscript are shown in red.

Reviewer #1: Niklas Boers

Summary:

This paper provides a very thorough synchronization of the GICC05 time scale obtained from counting annual layers in ice cores, and (assumed to be) absolute U/Th dates from several (sub-)tropical speleothems via cosmogenic radionuclides with a focus on 14C, for the time period from 10ka to 45ka BP. Based on this synchronization, the timing of the DO events during this interval is compared among ice core and speleothem records, and it is concluded that on average, no systematic lead or lag can be inferred, given the inherent uncertainties.

The paper is written very well, the subject is of great scientific importance, and the employed methodology seems accurate to me. I hence strongly support publication of this study in CP.

Thank you!

However, there are some instances where the presentation is not detailed enough at least for me to be able to precisely follow what is done exactly (see specific comments below). In addition, I have some slight conceptual concerns that I would suggest to be addressed prior to publication. Please note that I'm not a geochemist, so I apologize in advance for potentially trivial or irrelevant questions / concerns below.

Major comments:

1. Necessarily, some of the uncertainties are put in 'by hand', such as treating the MCE of the GICC05 time scale as 1 sigma, but also at several instances of the analysis of the cosmogenic radionuclides. This is not a critique per se, and I agree with the authors that their uncertainty estimates are probably very conservative. However, in the situations at hand, it cannot be quantified _how_ conservative, and this leads to a tricky situation: the more conservative the error estimates are chosen along the way, the harder it is to reject the null hypothesis of synchronous DO events in the different records. The final sigma reported for the average over all DOs and speleothems is 189yr, and a lot can happen with such uncertainties; the statement in the conclusion that on average the DO onsets occurred synchronously is thus misleading, I think. I'd suggest to rather emphasize here that no systematic leads or lags can be inferred given the (partly subjectively introduced) error estimates.

We agree, that 189 years is unfortunately still a substantial uncertainty. Also with respect to the comment by reviewer #2 we have reformulated our manuscript in the respective sections to say: "we reject the hypothesis of leads or lags larger than 189 years at the one sigma level."

In addition, the 189yr are not far from the delay between NGRIP and WAIS inferred to be significant by the WAIS members, would you mind to comment on this?

We don't see a relation between our inferred uncertainty that mainly arises from uncertainty in matching 10Be and 14C records, and the delay of the Southern Ocean response to the bipolar seesaw. Note that our 189 years are an uncertainty (the best guess is 26 years), while fur Buizert et al. (2015) the best guess is 218±46 years. The delay inferred by Buizert et al. is possibly related to the time it takes for eddies to propagate the temperature anomalies related to the bipolar seesaw across the Antarctic circumpolar current (Pedro et al. 2018, QSR). We don't see a reason why this mechanism should be related to our uncertainty estimate.

2. It is stated in the abstract, introduction, and in the discussion that the GICC05 uncertainties are reduced by 50-70%, but I don't understand where these numbers come from, and as far as I can tell, they are not mentioned / explained somewhere else in the manuscript. If you compare the GICC05 MCE to the sigma of the transfer function ensemble, it might be problematic, since the MCE is not really related to a normal distribution, despite the pragmatic approach by Andersen et al.

The 50-70% was indeed derived by comparing our 95.4% probability interval to the MCE (see for example figure 12) adopting the pragmatic approach of Andersen et al. to regard the MCE as a 2sigma uncertainty. But we agree that this quantitative comparison may not be ideal. Hence, we used the more qualitative formulation "strongly" instead.

3. It is not clear to me how exactly the interpolation in Sec. 4.4 is carried out. This is a key part of the study, and I would hence suggest to make this section considerably more detailed. It is written that the AR(1) realizations are used for interpolation, but how? You sample from the PDFs at the tie points, but how do you make sure that

a given AR(1) realization, starting at one tie point, ends up close to the next tie point? Note that I might be completely off track here.

As we write in the manuscript, we use the AR-process only for interpolation. Thus, this is not a random walk that by itself ends up at the tie-point. It is forced to do so. We generate the AR-noise purely based on the MCE (see pp. 19, l. 551-553) and then anchor it at the sampled tie-points, by calculating the difference between the AR-process realization and the PDF samples at the tie-points, and linearly correcting the AR-noise for this offset. As a result, the AR-process realization will be forced to run through the sampled tie-point, but vary freely in between, which gives us our interpolation uncertainty.

Also with respect to the comments below, we see that apparently our description of the way how we infer our interpolation uncertainty is not clear. Hence, we rewrote the entire section (see below). However, as we also write in the paper want to also stress again, that this section is merely an attempt to infer a conservative interpolation uncertainty while still using some constraints GICC05 is giving us.

"To construct a continuous transfer function between GICC05 and the U/Th timescale we apply a Monte Carlo approach. Each iterations consists of i) randomly sampling the PDFs at each tie-point and ii) interpolating in between the tie-points using an AR-process that is constrained by the GICC05 maximum counting error (mce). We use the tie-points shown in figure 7, 9, and 10, i.e., three tie-points between ice cores and tree-rings during the deglaciation, one tie-point between ice cores and the combination of Corals, Speleothems and Lake Suigetsu during the LGM, and one tie-point between ice cores and the Bahamas speleothem around the Laschamp event. For the interpolation, we use the time derivative of the mce (i.e., its growth rate) as an incremental error estimate. During periods when the growth rate is > 0 GICC05 may be stretched (compressed), while a growth rate of 0 does not allow this, independent of what the absolute mce is at that time. By multiplying this growth rate with a random realization of an AR-process, we simulate how much of that uncertainty has been realized in each iteration of the Monte Carlo simulation. Subsequently integrating over the resulting timeseries of simulated miscounts, we obtain again an absolute error estimate, i.e., one possible realization of the mce. In each iteration, this realisation is then anchored at the sampled tie-points (step i) by linearly correcting the offset between the sampled tie-points and the simulated counting error. Hence, this procedure provides us with a correlated interpolation uncertainty over time, taking into account some of the constraints provided by the ice core timescale itself, but giving priority to our inferred tie-points. We note that the treatment of the mce as an AR-process leads to larger interpolation errors compared to assuming a white noise model, which would lead to very small uncertainties that average out over long time (see also discussion in Rasmussen et al., 2006). Furthermore, we treat the mce as ±1σ instead of ±2σ as proposed by Andersen et al. (2006) which additionally increases our interpolation error. We stress that this procedure does not aim to provide a realistic model of the ice-core layer-counting process and its uncertainty which is clearly more complex (see Andersen et al., 2006; Rasmussen et al., 2006), nor should it be interpreted such that the mce was a 1σ uncertainty. However, our approach allows us to infer a conservative estimate of the interpolation uncertainty while at the same time it takes advantage of the fact that GICC05 is a layer counted timescale and hence, cannot be stretched/compressed outside realistic bounds. This procedure was repeated 300,000 times which was found sufficient to obtain a stationary solution, leading to 300,000 possible timescale transfer functions."

Specific comments:

p3, l83: How were the 50-70% uncertainty reduction inferred quantitatively?

As mentioned earlier, this is based on a direct comparison of the MCE and our 95.4% probability estimate. Regarding the earlier comment, we changed it to the less quantitative "strongly".

p3, l91: the “Hence” suggests that the previous sentence implies the _inverse_ relationship, but I don’t think it does, although I don’t question the inverse relationship itself.

Changed to “This causes the production rates of cosmogenic radionuclides to be inversely related…”

p6, l179: what do you mean by “more direct function of the timescale?”

As outlined on p6, L171-178, long term changes of accumulation rates depend on ice flow/thinning models of the ice sheet. On shorter timescales however, this thinning can be assumed to be near constant. In that case, accumulation rate variability depends only on the variability of the determined annual layer thickness, which is a direct product of the timescale that defines the age-depth relationship of the ice core.

We added on p6, L179:

“…more direct function of the timescale that determines the age-depth relationship and, thus, annual layer thickness, and is very precise….”

p6, l190ff: Using both flux-corrected and non-corrected version of the ice core records is fine to infer systematic differences between the records via comparison to the expected error of the mean, but I find it a bit problematic to use such a stack for the synchronization; do you obtain different results when using only flux-corrected or only non-corrected versions of the records?

We agree with this concern and in fact tested this during the analysis. All results shown in the manuscript are robust with respect to whether we chose just single ice cores or versions (flux/flux corrected) of the records. However, we do believe that stacking all ice cores increases the signal to noise ratio and thus, yields the best estimate.

p7 l215: please define “cal”

added: “(calibrated before present, AD 1950)”

p7 l221: Could you motivate the assumption of proportional 10Be and 14C production rate changes here?

We added “(see also following section)” on p7, L.221, where we deal with the question of 14C:10Be production rate ratios in great detail.

Fig.2: the time scales are not equidistant, right? How do you perform the the FFT filtering? Do you interpolate? If yes, using which method, and to which resolution?

Since Figure 2 shows modelled D14C data, the shown records are indeed equidistant, that is annual. The original ice core data are of lower resolution (between a few years to ~150 years). The ice cores were sampled more or less continuously, so that each radionuclide sample integrates over a given depth/age interval. Hence, a 10Be sample is an average of the 10Be concentration (or production) over an interval. Consequently, when producing the ice core stack (section 3.1) we enter each core as a step function. Firstly, this is closest to how the data is sampled, and secondly, this is important because the carbon cycle integrates over production rate changes. Hence, it matters for how long a given production rate is sustained.

However, with respect to the original question, we note, that the sampling resolution of the raw ice core data is sufficiently high (median resolution between 25 years for GRIP 10Be, and 130 years for GISP2 10Be), that calculating a 5000 year high pass filter is not sensitive to the interpolation method.

p10, l315: It may be my fault, but where in the results section are the window length and frequencies given? Can you be more specific?

We believe that we give these details:

P14, L412: “(<1000 years)”

P14, L430-431: “All data are FFT-filtered to isolate D14C variations on timescales <1000 years”

P14, L433-434: “ Each of the lower panels refers to a 2000-yer subsection of the data”

P15, L.444-445: “…we chose to linearly detrend each datset (instead of band-pass filtering)…”

P15, L.446-447: “we have to increase the length of the comparison window to 4,000 years”

P17, L.504-505: “For this tie-point, we merely remove the error-weighted mean between 39-45ka BP from each dataset, since detrending would remove the largest part of the signal”.

However, we now provide the details also more clearly in the method section:

“For the highly resolved tree-ring data we use a 1000 year high-pass FFT filter, while the lower resolved and more unevenly sampled coral/speleothem/macrofossil data is filtered by linear detrending to avoid the interpolation to equidistant resolution required for FFT analysis. Similarly, the high sampling resolution of the tree-ring data allows us to compare the data in 2,000 year windows, while we increase the window length to 4,000 and 5,000 years for the lower resolved data prior to 14ka BP. The exact frequencies and window lengths are also given in the results section.”

p10, l324: I don’t understand this sentence: do you mean that the delay between associated peaks in different sinusoidal signals increases with wavelength? Why?

Correct, that is what we mean. This is a known effect arising from the convolution of the production signal by the carbon cycle, owing to the different reservoirs that have different exchange rates and isotopic signatures. See for example figure 5 in Roth and Joos 2013. As we write 3 lines above (p10, L321-323):

“D14C variations in the atmosphere are dampened and delayed compared to the causal production rate changes. Both factors, attenuation and delay, depend on the frequency of the production rate change (Roth and Joos 2013; Siegenthaler et al., 1980).”

p11, l328: is the box-diffusion model by Siegenthaler et al referred to here?

In section 3.3 (“Carbon Cycle modelling”) we state that we’re using the box-diffusion model by Siegenthaler et al. 1980 (P7, L219-220). However, the statement on P11, L328, is independent of which model is used – there will always be a delay between production rate driven changes of atmospheric and marine D14C due to the carbon cycle.

p13, l284f: what do you mean by “deviations from the transition”? I find it a bit problematic to refer the reader to a paper in preparation here, since it’s not clear given the presentation here, how the change-point detection is carried out. In particular, further below it becomes clear that for each potential change point, PDFs are obtained for its onset, mid point, and end point, but it’s not clear how these PDFs are derived.

By “deviations from the transition” we mean the deviations from the fitted ramp, which we describe as AR(1) noise, compared to the stadial-interstadial transition which would be the “signal” in our case. The PDFs are generated using a MCMC sampler.

We rewrote the method section to hopefully be clearer:

We use a probabilistic model to detect the onset, mid-point, and end of the rapid climate transitions in each individual record. The employed model describes the abrupt changes as a linear transition between two constant states. Any variability due to the long-term fluctuations of the climate records around the transition model is described by an AR(1) process that is estimated in conjunction with the transition model. The model is independently fitted to windows of data on their individual timescales (Table 1 & Fig. 13) around the rapid transitions. Inference was performed using Markov Chain Monte Carl sampling (MCMC) to obtain PDFs of the timing of the onset, the length, and the amplitude of each transition in each record. Using these PDFs we can calculate delays of the onset, mid-point and end of the climate transitions between different records, propagating the respective uncertainties of the parameters. For each record, only events that are well expressed and measured in high resolution have been fitted. The approach and inference procedure are described in more detail in Erhardt et al. (submitted).

Fig.7: -there seems to remain quite a discrepancy between the variability of the bold grey line and the Towai treering data (green) even after synchronization, could you comment on this?

It is true, that the agreement is not perfect even after synchronization. Disagreements can arise from measurement noise, changes in 10Be transport and deposition, or carbon cycle changes. Without dedicated modelling, it is impossible to pinpoint the exact reason for individual discrepancies. These features also exist in the Holocene (see e.g., figure 10 in Adolphi et al. 2016, CP). However, 10Be and 14C agree well before (12-12.7 kaBP) and after (13.2-14.5kaBP) the disagreeing section around 13 kaBP. Given that the ice core timescale has small differential uncertainties, we find it unlikely that the disagreement in between 2 well matching sections can be due to errors in the timescale. Instead we think this highlights our use of relatively long windows to be compared, instead of peak-to-peak matching.

- if I’m not mistaken, none of the time scales of the shown data are equidistant, how to you do the FFT filtering in this case? If you interpolate, how?

As mentioned earlier, the annual resolution of the modelled D14C record arises from the step-function used for the carbon cycle model input. The resolution of the different ice core records is better than 50 years during this interval. The measured 14C data is decadal (tree-rings) to multi-decadal (speleothem). For the high-pass filter, we calculate an error-weighted mean of overlapping 14C data, interpolate annually, and calculate a 1000a low-pass filter. That low-pass filter is interpolated back to the original data resolution, and subtracted. Again, we note that all original data has a sampling resolution substantially higher than the cut-off frequency, so that the filtering is insensitive to the interpolation algorithm.

p15, l445: can you explain what you mean by "to remove offsets"? This also relates to l312 on p10; aren't offsets at longer time scales potentially problematic? I guess _heuristically_ these longer-term offsets are attributed to carbon cycle changes, but a clarifying sentence would be good, I think.

The different 14C records have partly systematic offsets between them, possibly due to reservoir (corals), dead-carbon (speleothems), or other (measurement) effects. Since we are only interested in relative changes, and not absolute values, we can remove those to isolate common co-variability. We added (see figure 1i) to make clear that we are referring to differences between the different 14C records.

Regarding offsets between 10Be and 14C on longer timescales: We think we outline our reasoning clearly on p10, L309-311:

"For our analysis we employ high-frequency changes in D14C since carbon cycle changes have only limited effects on atmospheric D14C on shorter time scales (Adolphi and Muscheler (2016). Similarly, as shown in figure 2, the agreement of the different ice-core records is better on shorter timescales."

In addition we motivate this approach already in the introduction on p4, L137-142:

"It is currently not possible to quantitatively correct either of the radionuclides for these non-production influences since neither past carbon cycle changes nor atmospheric circulation changes are sufficiently well known. However, the potential amplitude of non-production rate changes can be assessed through sensitivity experiments and added as an uncertainty for the production rate signal (Adolphi and Muscheler, 2016; Köhler et al., 2006)."

Furthermore, we discuss on page 6, L 175-181, why we think also the long term trends of the ice core 10Be data have large systematic uncertainties. In this sense, removing the long term trend reduces these uncertainties as well.

Fig.8: You present the result of the synchronization, and show the PDFs for the different windows, but I think one or two extra sentences in Sec 3.4 on how exactly these PDFs are used to shift the record across the windows would be very beneficial.

We do not use the PDFs shown in figure 8 for the final synchronization. We rewrote section 4.4 (see earlier) which now reads more clearly:

"We use the tie-points shown in figure 7, 9, and 10, i.e., three tie-points between ice cores and tree-rings during the deglaciation, one tie-point between ice cores and the combination of Corals, Speleothems and Lake Suigetsu during the LGM, and one tie-point between ice cores and the Bahamas speleothem around the Laschamp event."

Furthermore, we added at the end of section 4.2 where the tie-point is presented L.520:
"We used this tie-point (figure 9) in the final synchronization as it is the best-defined feature in this time interval, and consistent within error with the estimates shown in figure 8."

Fig11: there's no inset and no blue dashed line!

We assume that this comment refers to p18, L537. This should of course read Fig. 10 instead and has been corrected.

Also, shouldn't the four individual speleothem dates correspond to the measured (black) points of NRM/ARM?

No, the U/Th dates have been carried out at different depths than the geomagnetic analyses.

p.19: As noted above, I don't understand how you interpolate between the tie points, the description is too brief in my opinion: by derivative of the MCE, do you mean the increments from one measured point to the next? Why do you multiply the AR(1) with these? Which "cumulative sum back in in time", i.e. from where to where? You say "strong autocorrelation", but what is the value of the parameter alpha?

We hope that we could clarify this in our earlier reply and the rewriting of that section.

p19, l 575: what do mean by "grow/shrink at a rate determined by the mce"? the latter is cumulative and hence always increasing back in time, but your AR(1) based uncertainties decrease when going back in time towards the next time point. I agree that it should decrease this way, but I don't understand the method sufficiently from the given description to understand how, specifically.

We hope we clarified this above. While the mce is typically plotted as a cumulative error back in time, it is really its growth rate that determines the counting error for each time interval.

p22, l639: Here you say that you sample the PDFs for the DO onsets; am I correct in assuming that for each onset, you obtain a PDF of its dates from the change-pointdetection?

Correct. We added "(section 3.5)"

p23, l679: I don't think that this study _shows_ that the counting error can be strongly correlated over extended period; please correct me if I'm wrong!

We do think that it shows exactly that. The results by Adolphi et al. (2016) show that the offset between the tree-ring timescale and GICC05 during the Holocene, requires that nearly every layer, that has been marked "uncertain", has in fact not been a year. Similarly, our results show that to reconcile GICC05 and tree-rings/speleothems between 10-22kaBP require that almost every uncertain year in this period has been a "real" year. Thus, we think that our statement is correct.

But it is true, that we do not derive this explicitly, so we changed it to: "implies".

p25, l727-738: it would be good, I think, to add reference on the relation to the ITCZ position already here.

To provide a theoretical reference to why the ITCZ may migrate in concert with North Hemisphere abrupt events we added a reference to Schneider et al. (2014) in line 731.

p25, l739ff: The fact that the precip increase in El Condor and Cueva del Diamante significantly predates the onset of H2 in Greenland suggests that the southward shift of the ITCZ, proposed to explain the precip increase, was not caused by H2, but rather by long-term solar insolation changes and in particular the NH minimum around this time, right? Also, Fig.16 suggests that the variability in AMOC strength (related to H2) does not substantially affect the position of the ITCZ, but merely the precipitation anomalies north and south of the ITCZ. If this is correct, please revise the paragraph accordingly.

We changed: "…during a weak AMOC state, reduced advection of moisture from the tropical Atlantic leads to lower precipitation north of the ITCZ, while the ITCZ position over South America itself changes very little (Fig. 16)."

And: "It is hence possible, that when northern hemisphere summer insolation reached its lowest values over the past 50 kaBP around H2, the ITCZ migrated to a position south of El Condor and Cueva del Diamante, and during its transition caused the reconstructed precipitation change."

p27, l783: see above regarding the 50-70%

See earlier reply. Changed to "strongly"

p27, l784: note the above comment on the formulation that DO events occur on average synchronously, rather, the null hypothesis of synchronicity cannot be rejected given the uncertainties. Your statement in the abstract is more accurate, I think.

See earlier reply. Changed to "We reject the hypothesis if leads or lags larger than 189 years between Greenland, East Asia, and South America at the one sigma level."

Sorry for the lengthy report, I hope it's helpful!

Best,

Niklas Boers

Thank you for providing this valuable input. We think it improved the manuscript!

Reviewer #2: Jeff Severinghaus

This paper addresses a crucial problem we face in paleoclimatology - namely that many of us are going ahead and using the U-Th-dated speleothems to improve other paleo chronologies, without really having answered the fundamental question of whether the abrupt DO events seen in speleothems are synchronous with those seen in

Greenland ice cores. I am as guilty of this as anyone - in Buizert et al. (2015) Clim. Past, 11, 153–173 we made a physical argument based on known atmospheric and oceanic processes that the Chinese speleothem DO events cannot have lagged Greenland's DO events by more that 50 years. We then proceeded to tie the Greenland and WAIS Divide timescales in a pragmatic fashion to the Chinese speleothems, adopting an uncertainty of 50 years due to the assumption of synchroneity. I do believe that this argument is solid, but it is not enough for the high scientific standards we as a community must ultimately achieve, and the authors of the present paper are attempting to rectify this problem and empirically show that this lag cannot be very large. Therefore this work is essential, timely, and critical to the paleo field, and therefore I think this paper should be published with only very minor revisions.

Thank you!

The ultimate uncertainty that the authors arrive at is large, unfortunately, so it is perhaps best if the language of the conclusions is adjusted to reflect that large uncertainty. Instead of saying that the speleothems and ice cores are synchronous within uncertainty (which is true), it might be more helpful to the reader to write "we can reject the hypothesis of asynchrony larger than 189 yrs" or something equivalent. That way the conclusion shows what has actually been added by the present work.

Changed accordingly.

Minor comments:

The term "synchronicity" is used in psychology (i.e. Carl Jung) and has nothing to do with paleoclimate or chronology. The proper term is "synchroneity". Please change all the uses in the paper accordingly.

Done.

Reviewer#3: Paula Reimer

This manuscript uses cosmogenic isotopes to synchronize the Greenland ice core timescale with the U-Th timescale through a meticulous, multi-step process. The authors minimize the root mean square error in the production rate models from geomagnetic field based reconstructions and the ice cores to resolve the scaling factor for 10Be. They then compare 14C archives from around the Lachamps event with the reconstruction from the scaled ice core stack to select the most suitable ocean ventilation rate for the carbon cycle. The investigation into the effect of delay between ice core reconstructed atmospheric 14C changes and the marine and speleothem archives was insightful. Once the ice cores were synchronized to the U-Th (and dendrochronological) timescale the synchroneity of the proxy response to D-O cycles in a number of speleothem climate records was tested. This represents a very important step in interpretation of palaeoclimate records. The ice core based 14C reconstruction will also provide a guide to improvements for the next IntCal radiocarbon calibration curveupdate.

Thank you.

Specific comments: p. 2, line 52-54 'About one third of the data underlying the current radiocarbon calibration curve, IntCal13 (Reimer et al., 2013), obtain their absolute age from climate wiggle-matching.' The climate wiggle-match records make up about 6% of the total data used in IntCal13 not 1/3 as stated (423 out of 7019 data points; IntCal13 database accessed 9 August 2018 http://intcal.qub.ac.uk/intcal13/)

We are sorry for this imprecision (in multiple ways).

Firstly, we are of course only referring to the glacial part older than 13.9kaBP where IntCal13 only consists of archives other than tree-rings, but which is also the period of the occurrence of DO-events, which is relevant for our discussion. In this section, 1623 14C determinations enter the curve of which 412 are climate wiggle-matched (Cariaco unvarved, Iberian Margin, Pakistan Margin). So that is 25%.

We clarified this in the manuscript:

"The current radiocarbon dating calibration curve (IntCal13, Reimer et al., 2013) is constructed from accurately dated tree-ring chronologies back to 13.9 ka BP (13.9 ka BP, kilo-years Before Present AD 1950). Beyond this time, which encompasses all DO-events, about one fourth of the data underlying IntCal13 obtain their absolute age from climate wiggle-matching."

p. 7, lines 208-210 'The timescale of the Lake Suigetsu record has been inferred from matching its 14C record to the 14C variations in speleothems, additionally constrained by varve counting (Bronk Ramsey et al., 2012).' This statement seems a bit backwards to me since the varve counting provided the initial timescale which was then adjusted by matching the 14C records in speleothems, but if co-author CBR is happy with the way it's written then that is fine.

We changed the statement to:

"The timescale of the Lake Suigetsu record is based on varve counting, corrected for long-term systematic errors by matching its 14C record to the 14C variations in speleothems (Bronk Ramsey et al., 2012)."

p.10, Figure 4. How are the 14C anomalies calculated here? Filtering is mentioned in line 292 but details are not given until section 3.4 and in section 4.3 where the error weighted mean is removed from the data for the Laschamp period. Obviously that was not the case for Figure 4. What do the dashed boxes represent?

We removed the error weighted mean prior to the Laschamp event from all datasets. We added to the caption of figure 4:

"All data are shown as anomalies to their error-weighted mean prior to the Laschamp event. i.e., the $\Delta^{14}C$ increase. The dashed boxes encompass the time periods and $\Delta^{14}C$ uncertainties (error of the error weighted mean) used for the definition of the pre-and post-Laschamp event levels."

Section 3.5: Change-point detection in climate records. This is an abrupt shift from synchronizing 14C records and 10Be in ice core records to comparing to the timing or d18O shifts in climate records. The climate records considered are not even identified here except by a site name in Table 1. Presumably this should be part of Section 5 ?

We agree that this is a relatively abrupt shift. However, we think that this should still be part of the method section 3. We added a short introductory paragraph to the section:

"To test the synchroneity of rapid climate changes, we compare the timing of DO-events seen in Greenland ice cores (Andersen et al., 2004), to a number of well-known U/Th dated speleothems that show DO-type variability from Hulu Cave (Cheng et al., 2016), Sofular Cave (Fleitmann et al., 2009), El Condor, and Cueva del Diamante (both Cheng et al., 2013b)."

Section 5. Figure 13. Why is the NGRIP Ca record used instead of d18O? A word of explanation here would be useful.

We added in section 5:

"We used the NGRIP Ca record (Bigler, 2004), that shows the largest signal to noise ratio across DO-events (compared to e.g., $\delta^{18}O$) making their identification more precise. In addition, the Ca aerosols originate from Asian dust sources (Svensson et al., 2000) and are thus, more directly related to Asian hydroclimate (Schüpbach et al., 2018) making them potentially more comparable to for example the Hulu cave record. Potential phasing differences between different climate proxies in the ice core are small compared to our synchronization uncertainties (Steffensen et al., 2008)."

p.24-25 line 722-723 'Since IntCal13 in principle should be tied to the U/Th-age scale'. This phrase needs some qualification since IntCal13 is tied to dendrochronological time scale for 0 to 14,000 cal BP and while the Hulu cave U-Th agrees well with the tree-ring data it only begins at 10,730 cal BP.

'Since IntCal13 in principle should be tied to the U/Th and dendrochronological age scale…'

True. To be more precise in our formulation we changed the sentence to:

"Since IntCal13 in principle should be tied to the U/Th-age scale for sections older than 13.9 ka BP, this implies either an…"

All figures would benefit from being presented in a larger size.

We hope CP takes care of this request during the layout/typesetting process.

Reviewer #4: Frederic Parrenin

This manuscript discusses the relative timing of DO events observed in Greenland ice cores with those observed in dated speleothems. The methodology is based on the synchronisation via cosmogenic radionuclides. The synchronisation is done during three intervals where variations in production of cosmogenic radionuclides are important: 11-13 ka, 21-23 ka and 41-43 ka (Laschamp event). In-between these three time periods, a kind of interpolation is done and its uncertainty is evaluated thanks to a statistical method which assumes the GICC05 MCE as age interval uncertainty. It is found that DO events are synchronous in ice cores and speleothems within uncertainties (189 yr). Moreover, GICC05 is found to agree with the U-Th chronology of speleothems within its MCE uncertainty, although clearly the MCE is strongly correlated in some intervals (e.g. uncertain layers are always real layers).

This is an interesting manuscript which is very well written. I will focus on the discussion of chronologies since I am not an expert of cosmogenic radionuclides. The only main comment I have is that the title and the formulation of the manuscript are a bit misleading since this manuscript does NOT provide a continuous connection of ice core and speleothems chronologies, but rather a discrete one during only three time periods. The interpolation which is done in-between is just an interpolation and in my opinion should not be treated as a continuous synchronisation.

Thank you for your feedback. We are not sure how to comply with this request though. We want to remind the reviewer that eventually almost any synchronization method is based on more or less discrete tie-points between timescales (volcanoes, rapid CH4 changes, climate-wiggle matching). In between, there is always some sort of interpolation required, which obviously becomes more uncertain as the distance between the tie-points increases. How much more uncertain it becomes depends on whether we have prior information on the stratigraphy of the archives. We exploit this information from the layer counted ice core timescale telling us how this uncertainty is growing width depth/time between horizons.

We believe that we i) never state we would provide a continuous synchronization (but a continuous transfer-function), ii) clearly illustrate in text and figures, that this is only based on a few tie-points, iii) provide conservative interpolation errors by treating the mce as correlated and 1 instead of 2 sigma.

Obviously we hope that more tie-points can be established in the future as new data becomes available. But as we show in figure 12, out transfer function is consistent within error with the few independent tie-points that are available for testing our approach, during a period that is far away from our actual tie-points.

In summary, we hope that our results provide a test-bed for future studies and believe that given the current constraints, we provide the best-guess for the timescale difference between Greenland ice cores and U/Th dated speleothems without applying climate wiggle-matching and the underlying assumptions.

Connecting the Greenland ice-core and U/Th timescales via cosmogenic radionuclides: Testing the synchroneity of Dansgaard-Oeschger events

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Abstract. During the last glacial period Northern Hemisphere climate was characterized by extreme and abrupt climate changes, so-called Dansgaard-Oeschger (DO) events. Most clearly observed as temperature changes in Greenland ice-core records, their climatic imprint was geographically widespread. However, the temporal relation between DO-events in Greenland and other regions is uncertain due to the chronological uncertainties of each archive, limiting our ability to test hypotheses of synchronous change. On the contrary, the assumption of direct synchrony of climate changes forms the basis of many timescales. Here, we use cosmogenic radionuclides (^{10}Be, ^{36}Cl, ^{14}C) to link Greenland ice-core records to U/Th-dated speleothems, quantify offsets between both timescales, and improve their absolute dating back to 45,000 years ago. This approach allows us to test the assumption that DO-events occurred synchronously between Greenland ice-core and tropical speleothem records at unprecedented precision. We find that the onset of DO-events occurs within synchronization uncertainties in all investigated records. Importantly, we demonstrate that there remain local discrepancies in the temporal development of rapid climate change for specific events and speleothems. These may be either related to the location of proxy records relative to the shifting atmospheric fronts or to underestimated U/Th-dating uncertainties. Our study thus highlights the potential for misleading interpretations of the Earth system when applying the common practice of climate wiggle-matching.

1 Introduction

Precise and accurate chronologies are critical for understanding past environmental and climatic changes. Global natural and anthropogenic archives can only be directly compared through the development of robust chronological frameworks, enabling studies of the spatiotemporal dynamics of past change. These findings are crucial for understanding the nature and cause of rapid climate changes in the past, and hence, characterizing the dynamics and feedbacks of past and projected future climate change (Thomas, 2016). However, the applicability, precision, and accuracy of the available dating methods pose strong constraints on our ability to infer leads and lags between climate records, and ultimately, mechanisms of change in the Earth system. Instead, the situation is often reversed: climate changes such as Dansgaard-Oeschger, or DO, events (Dansgaard et al., 1993; Dansgaard et al., 1969) are typically *assumed* to occur synchronously across the Northern Hemisphere in different climate proxies from various regions and then used as chronological tie-points. This so-called "climate wiggle-matching" forms the chronological basis of a large part of paleoclimate records (e.g., Bard et al., 2013; Hughen et al., 2006; Henry et al., 2016; Turney et al., 2015), especially in the marine realm where other dating methods suffer from low precision and poorly constrained biases such as the marine radiocarbon reservoir age (Lougheed et al., 2013). Furthermore, it also plays a central role for one of the most widely used dating methods in paleosciences – the radiocarbon dating method. The current radiocarbon dating calibration curve (IntCal13, Reimer et al., 2013) is constructed from accurately dated tree-ring chronologies back to 13.9 ka BP (13.9 ka BP, kilo-years Before Present AD 1950). Beyond this time, which encompasses all DO-events, about one fourth of the data underlying IntCal13 obtain their absolute age from climate wiggle-matching.

Climate wiggle-matching has the obvious drawback that the leads and lags between different climate records cannot be studied once the records have been forced to align. The approach critically rests on the assumptions, that i) the climate change indeed occurred synchronously everywhere, and that ii) the (sometimes fundamentally different) proxies in question record the changes in a similar way and without intrinsic delays. These assumptions, however, can very rarely be rigorously tested but when they are, ample evidence is revealed that questions their universal validity. Lane et al. (2013) showed that rapid climate change in the North Atlantic region may be time transgressive with regional leads and lags on the order of a century. Nakagawa et al. (2003) argued that the onset of Greenland Interstadial 1e (GI-1e, Rasmussen et al., 2014a) occurred multiple centuries after the associated climate shift in Japan (and subsequent revisions of the underlying timescales (Staff et al., 2013; Bronk Ramsey et al., 2012; Seierstad et al., 2014) did not resolve this conundrum). Buizert et al. (2015) inferred that the Southern Ocean response to DO-events is delayed by about 200 years on average while the atmosphere around Antarctica reacted instantaneously (Markle et al., 2016). Baumgartner et al. (2014) found asynchronicities between ice-core proxies for local Greenland temperature (δ^{15}N) and the tropical/mid-latitude hydrological cycle (CH_4) during some DO-events. They discussed that the climate changes in polar and low-latitude regions may indeed be synchronous, but that atmospheric CH_4 concentrations rise with a delay during some DO-events because of compensating changes in the source strengths of the northern and southern hemisphere wetlands. Alternatively, their findings can be explained via a real delay between Greenland climate change and the activation of CH_4 source areas during certain DO-events. Fleitmann et al. (2009) reported on timing differences of DO-events in Greenland ice cores and speleothems, albeit largely within dating uncertainties. However, they also found significant differences between speleothem records outside their chronological uncertainties. This is complemented by a recent study showing that the duration of a stadial-interstadial transition can differ by up to 300 years between different East Asian speleothems (Li et al., 2017) emphasizing the questions of whether we should expect the onset, mid-point, or end-point of DO-events to occur simultaneously, as this choice will lead to different results when aligning the records.

In this paper, we attempt to provide improved constraints on the paradigm of climate synchroneity. We employ cosmogenic radionuclides as a climate-independent synchronization-tool between the Greenland ice-core timescale (Andersen et al., 2006; Rasmussen et al., 2006; Seierstad et al., 2014; Svensson et al., 2008; Svensson et al., 2006; Vinther et al., 2006) and the U/Th timescale (Broecker, 1963; Edwards et al., 1987; Cheng et al., 2013a) and strongly reduce the absolute dating error of the Greenland ice cores back to 45,000 years BP. This allows us to compare the timing of DO-type variability seen in key paleoclimate records at unprecedented precision: The Greenland ice cores and U/Th-dated (sub-)tropical speleothems.

2 Cosmogenic radionuclides as synchronization tools

Cosmogenic radionuclides (such as ^{14}C, ^{10}Be and ^{36}Cl) are produced in a nuclear cascade that is triggered when galactic cosmic rays (GCR) collide with the Earth's atmosphere's constituents (Lal and Peters, 1967). While the GCR flux outside the heliosphere can be assumed to be constant over the past million years (Vogt et al., 1990), the flux arriving at Earth is modulated by the strength of the helio- and geomagnetic fields (Masarik and Beer, 1999). This causes the production rates of cosmogenic radionuclides to be inversely related to changes in solar activity and/or the strength of the geomagnetic field. This modulation effect leaves a globally synchronous, externally forced signal in cosmogenic radionuclide records around the world. Hence, they can serve as a powerful synchronization tool for climate archives from different regions. The challenge lies in estimating potential non-production-related impacts on radionuclide concentrations in a given archive that may result from geochemical and meteorological processes.

After production, ^{14}C is oxidized to $^{14}CO_2$ and enters the carbon cycle. Changing ^{14}C production rates thus alter the atmospheric $^{14}C/^{12}C$ ratio (expressed as per mille $\Delta^{14}C$, that is, $^{14}C/^{12}C$ corrected for fractionation and decay relative to a standard, denoted Δ in Stuiver & Pollach, 1977). Due to carbon cycle effects, these variations in $\Delta^{14}C$ are dampened and delayed with respect to the causal production rate changes (Siegenthaler et al., 1980; Roth and Joos, 2013). In addition to variable production rates, changes in the exchange rates between the different carbon pools can alter $\Delta^{14}C$. The world's oceans in particular have a significantly lower $\Delta^{14}C$ than the contemporary atmosphere due to their long carbon residence time (Craig, 1957). Thus, variations in the ^{14}C exchange rates between the ocean and the atmosphere will alter atmospheric $\Delta^{14}C$ independent of production rate changes.

^{10}Be attaches to aerosols and is transported from the stratosphere to the troposphere within 1-2 years (Raisbeck et al., 1981) mainly via mid-latitude tropopause breaks (Heikkilä et al., 2011). It has no active geochemical cycle and so its atmospheric concentration is a more direct recorder of production rate changes compared with $\Delta^{14}C$. However, ^{10}Be transport and deposition in the troposphere is guided by local meteorology and thus susceptible to changes thereof (Heikkilä and Smith, 2013; Pedro et al., 2011). This can cause variations in ^{10}Be records that are not related to production rate changes. Furthermore, a so-called "polar bias" (i.e., an overrepresentation of polar as opposed to global production rate changes) has been proposed for ice-core records (Bard et al., 1997). This would lead to subdued geomagnetic and enhanced solar modulation of ice-core radionuclide records due to the geometry of the geomagnetic field. However, there is no consensus in different empirical studies and modelling experiments to whether this effect is present and the results may also vary regionally (Bard et al., 1997; Heikkilä et al., 2009a; Pedro et al., 2012; Adolphi and Muscheler, 2016; Muscheler and Heikkilä, 2011; Field et al., 2006).

The transport and deposition of ^{36}Cl in its aerosol phase is comparable to ^{10}Be. However, in addition to an aerosol phase, ^{36}Cl also has a gaseous phase ($H^{36}Cl$) which is likely dominant in the stratosphere (Zerle et al., 1997). In the troposphere, the partitioning between aerosol and gas phase is not well understood. It may vary in space and time (Lukasczyk, 1994), and can change rapidly depending on pH (Watson et al., 1990). The gaseous $H^{36}Cl$ phase can also be lost from acidic ice in low accumulation sites after deposition which is, however, less relevant for the high accumulation sites studied here (Delmas et al., 2004). In Greenland, similar to ^{10}Be, the dominant deposition process of ^{36}Cl in is wet deposition (Heikkilä et al., 2009b) which is supported by the overall similarity of ^{36}Cl and ^{10}Be variations recorded in ice cores (Wagner et al., 2001b; Muscheler et al., 2005).

As a result, all three radionuclides depend on the same production mechanism which causes their production rates to co-vary globally. This signal can be exploited for global synchronization of paleorecords from natural archives. However, to identify these common changes, their different geochemistry needs to be accounted for. In the case of radiocarbon this is achieved through carbon cycle modelling, to deconvolve the effects of the carbon cycle on the relation between ^{14}C production rates and $\Delta^{14}C$ (Muscheler et al., 2004). For ^{10}Be and ^{36}Cl, fluxes can be calculated from ice accumulation rates. This provides a first-order correction for changing paleoprecipitation rates on the ice sheet and their influence on the radionuclide concentrations. In reality, aerosol transport to the ice sheet is more complex and depends on changes in transport velocity, pathways and scavenging effects en route (Schüpbach et al., 2018), which are, however, difficult to constrain for ^{10}Be due to its stratospheric origin. Instead, comparisons of fluxes and concentrations to other climate proxies can inform about potential climate influences on ^{10}Be and ^{36}Cl transport and deposition (Adolphi and Muscheler, 2016). It is currently not possible to quantitatively correct either of the radionuclides for these non-production influences since neither past carbon cycle changes nor atmospheric circulation changes are sufficiently well known. However, the potential amplitude of non-production rate changes can be assessed through sensitivity experiments and added as an uncertainty for the production rate signal (Adolphi and Muscheler, 2016; Köhler et al., 2006).

The potential of this synchronization tool has been demonstrated multiple times to infer differences between the tree-ring and ice-core timescales (Adolphi and Muscheler, 2016; Muscheler et al., 2014a; Southon, 2002), test the accuracy of the radiocarbon calibration curve (Adolphi et al., 2017; Muscheler et al., 2014b; Muscheler et al., 2008), and synchronize ice cores from both hemispheres (Raisbeck et al., 2017; Raisbeck et al., 2007).

3 Methods & Data

3.1 Ice-Core Data

The ice-core ^{10}Be and ^{36}Cl data used in this study are shown in figure 1. We focus on records that have been robustly linked to the GICC05 timescale (Andersen et al., 2006; Rasmussen et al., 2006; Seierstad et al., 2014; Svensson et al., 2008; Rasmussen et al., 2008). Hence, the majority of the data stems from the deep Greenland ice cores GRIP, GISP2, and NGRIP. In addition, we use Antarctic ^{10}Be fluxes from EDC, EDML and Vostok

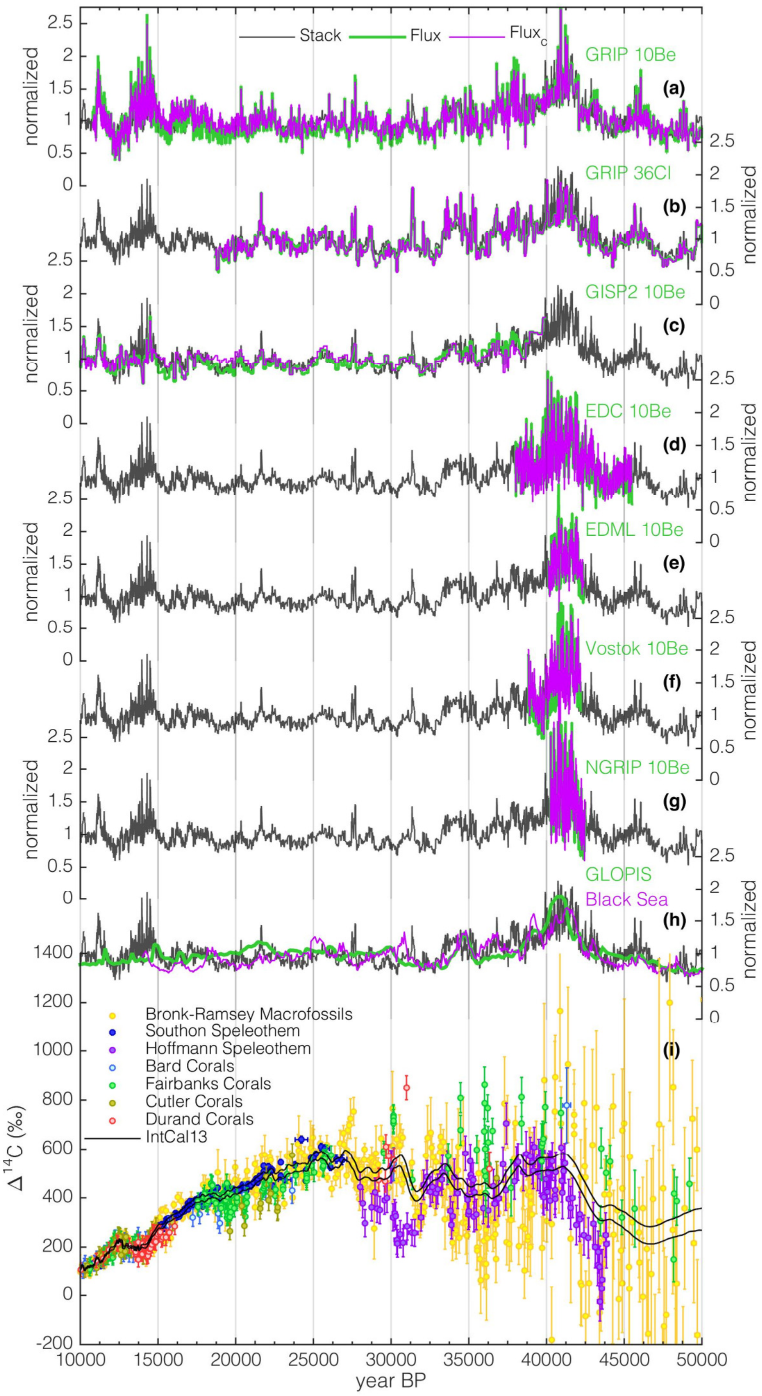

Figure 1: Data used in this study. Panel a-g show individual ice-core records of GRIP ^{10}Be (Baumgartner et al., 1997b; Muscheler et al., 2004; Wagner et al., 2001a; Yiou et al., 1997; Adolphi et al., 2014), GRIP ^{36}Cl (Baumgartner et al., 1998; Baumgartner et al., 1997a; Wagner et al., 2001b; Wagner et al., 2000), GISP2 ^{10}Be (Finkel and Nishiizumi, 1997), and ^{10}Be from EDC, EDML, Vostok, and NGRIP (all Raisbeck et al., 2017). Each record represents deposition fluxes (green) and 'climate corrected' fluxes (purple, see text). In addition, each panel contains the stack of all ice-core records (black, see text). Panel h: ^{10}Be production rates modelled from two geomagnetic field intensity reconstructions: GLOPIS (green, Laj et al., 2004) and based on Black Sea sediments (purple, Nowaczyk et al., 2013) using the production rate model by Herbst et al. (2016). The ice-core radionuclide stack is shown in black. All records in panel a-h are shown on the GICC05 timescale (Seierstad et al., 2014) and normalized to (i.e., divided by) their mean. Panel i: Absolutely dated ^{14}C data from Lake Suigetsu (yellow, Bronk Ramsey et al., 2012), Hulu Cave (blue, Southon et al., 2012), Bahamas speleothems (purple, Hoffmann et al., 2010), and various tropical coral datasets (Bard et al., 1998; Cutler et al., 2004; Durand et al., 2013; Fairbanks et al., 2005, shown in light blue, olive, red, and green, respectively). The black lines encompass the ±1σ uncertainties of IntCal13 (Reimer et al., 2013).

that have been anchored to GICC05 by matching solar variability present in all ^{10}Be records, and volcanic tie-points (Raisbeck et al., 2017).

By calculating fluxes we make a first order correction for the changing snow accumulation rates between stadials and interstadials and their influence on radionuclide concentrations (Wagner et al., 2001b; Johnsen et al., 1995; Rasmussen et al., 2013; Finkel and Nishiizumi, 1997). The accumulation rates for each ice core are based on their annual layer thickness – derived from their individual timescales – corrected for ice thinning. For the Greenland ice cores this thinning function is based on a 1-D ice flow model (Dansgaard and Johnsen, 1969; Johnsen et al., 1995; Johnsen et al., 2001; Seierstad et al., 2014). For the Antarctic ice cores we use the strain rate derived from the Bayesian ice-core dating effort AICC12 (Veres et al., 2013). These strain rates are inherently uncertain and independently derived accumulation rate estimates differ by up to 10-20% in the glacial (Gkinis et al., 2014; Rasmussen et al., 2013; Guillevic et al., 2013). However, these differences are largely systematic and change only on multi-millennial timescales. The shorter term changes in accumulation rates are a more direct function of the timescale that determines the age-depth relationship and, thus, annual layer thickness, and is very precise for increments of the core (Rasmussen et al., 2006). This is important to note, as we mainly exploit production rate changes on centennial to millennial timescales for synchronization.

To test for additional climate influences on ^{10}Be or ^{36}Cl deposition in the ice cores, we followed the approach by Adolphi and Muscheler (2016): For each ice core we calculated multiple linear regression models using $\delta^{18}O$ and snow accumulation rates as predictors for ^{10}Be (^{36}Cl) fluxes and subtracted the obtained model from the ^{10}Be (^{36}Cl) data. We denote the resulting record as the "climate corrected flux" ($Flux_c$). This approach may correct climate effects on ^{10}Be (^{36}Cl) deposition insufficiently, or it may over-correct them, so it cannot be assumed per se that the resulting record is more reliable than the original fluxes. Nevertheless, it provides a first order sensitivity test for the ice-core records with respect to climate-related transport and depositional effects on ^{10}Be (^{36}Cl) fluxes.

To combine all ice-core records, we calculated their mean (denoted as "Stack", Fig. 1) using Monte-Carlo bootstrapping (Efron, 1979). Using 7 ice-core records in two versions (flux and $flux_c$) yields a total number of 14 samples. In each iteration, 14 samples are randomly drawn (with replacement, i.e., each record can be drawn multiple times), perturbed within measurement errors, and stacked. Repeating this procedure 1,000 times we obtain an average relative standard deviation of 8% between the derived stacks, which is comparable to the measurement uncertainty of individual measurements but larger than the expected error of the mean which points to systematic differences between the records. For the period where we have data from both hemispheres this standard deviation is only slightly higher (10%). Even though this is only a relatively short period (see Fig. 1), it contains multiple DO-events which are expressed differently in Northern and Southern Hemisphere climate. Thus, this agreement can serve as indication that climate effects do not dominate the signal.

3.2 Radiocarbon data

For the purpose of this study we have to focus on radiocarbon records that are absolutely dated. Furthermore, the length and sampling resolution of the records need to be sufficient to resolve centennial-to-millennial production rate changes. The records that fulfil these criteria are shown in figure 1 and comprise ^{14}C data from various U/Th dated coral records (Bard et al., 1998; Durand et al., 2013; Cutler et al., 2004; Fairbanks et al., 2005), as well as ^{14}C measured in two speleothems (Southon et al., 2012; Hoffmann et al., 2010). In addition, we use the ^{14}C record from Lake Suigetsu (Bronk Ramsey et al., 2012) since the U/Th dated records do not directly reflect atmospheric ^{14}C but the ocean mixed layer (corals) and, in the case of speleothems, a mixture of atmospheric and soil CO_2, and carbonate bedrock from above the cave. The timescale of the Lake Suigetsu record is based on varve counting, corrected for long-term systematic errors by matching its ^{14}C record to the ^{14}C variations in speleothems (Bronk Ramsey et al., 2012). Hence, it is not truly independently dated. However, similar to ice-core layer counting, this varve count adds constraints especially on centennial timescales, so that $\Delta^{14}C$ variations on these timescales should be relatively unaffected by this tuning to the speleothem ^{14}C data. Thus, even though the timescale may not be independent, this record can still be used to verify the existence of $\Delta^{14}C$ variations in the atmosphere seen in the mixed layer records.

In addition, we use the available tree-ring records back to 14,000 cal BP (calibrated before present, AD 1950) in the revised version by Hogg et al. (2016)(not shown in figure 1 for clarity).

3.3 Carbon cycle modelling

To be able to compare ice-core and radiocarbon records directly we have to account for the effects of the carbon cycle. Following earlier studies (Muscheler et al., 2004; Muscheler et al., 2008), we use a box-diffusion carbon cycle model (Siegenthaler et al., 1980) to model $\Delta^{14}C$ from the ice-core radionuclide records. We assume that ice-core ^{10}Be (^{36}Cl) variations are proportional to ^{14}C production rate changes (see also following section) and model $\Delta^{14}C$ anomalies from each realization of the ice-core stack, as well as the single ice-core records (Fig. 2). It can be seen that the modelled $\Delta^{14}C$ records from the individual ice-core records differ in their long-term trends since the carbon cycle integrates over time so that relatively small but systematic differences in the radionuclide fluxes (possibly arising from uncertainties in the strain rates) have a significant effect on longer time scales. However, all records show the same overall evolution of $\Delta^{14}C$. Furthermore, especially when subtracting the long-term trend and isolating variations on timescales shorter than 5000 years, the agreement is very high (on average within 15‰ at 1σ, Fig. 2b), which is the part of the signal that we will be exploiting in our synchronization effort.

3.3.1 Production rate ratio

Modeling $\Delta^{14}C$ values from ^{10}Be measurements is based on the assumption that ^{10}Be and ^{14}C production rate changes are proportional to each other. However, different production rate models differ in their sensitivity of ^{14}C and ^{10}Be production rate changes to variations in the geomagnetic field (Cauquoin et al., 2014). For a given geomagnetic field change, the production rate model by Masarik and Beer (2009, 1999) yields 30-50% lower ^{10}Be production rate changes than the calculations by Poluianov et al. (2016) and Herbst et al. (2016). For ^{14}C on the other hand, all models yield roughly similar amplitudes. This leads to differences in the $^{14}C/^{10}Be$ production rate ratio for a given change in the geomagnetic field. If Masarik and Beer (1999) are correct, the variations in ice-core ^{10}Be records have to be upscaled by 30-50% to be proportional to ^{14}C production rate changes while no such scaling is necessary when the other production rate models are used. In addition, the amplitudes in ^{14}C and ^{10}Be may differ due to the presence of polar bias (see section 2). If this effect was present, then geomagnetic field changes should cause bigger variations in ^{14}C than ^{10}Be.

Since the presence of a polar bias is debated and the physical reason for the differences between the production rate models is unresolved, we chose an empirical approach to scale the ice-core record appropriately:

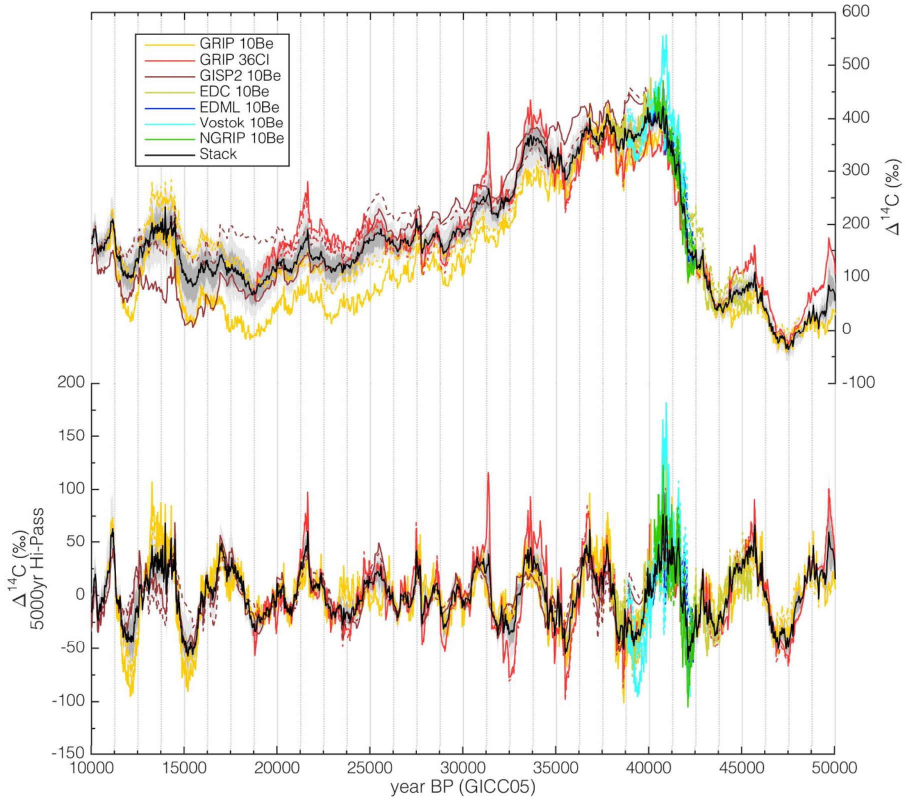

Figure 2: Modelled $\Delta^{14}C$ anomalies from individual ice-core records (see legend, solid lines are based on radionuclide fluxes while dashed lines are inferred from $flux_c$) and the realizations of the ice-core stack (black line shows the mean of all realizations, dark and light grey shading encompass 68.2 and 95.4% probability ranges). The top panel shows the unfiltered model output. The bottom panel displays the records after variations with frequencies <1/5000a^{-1} have been subtracted (FFT-based filter).

We use three geomagnetic field intensity reconstructions around the Laschamp geomagnetic field minimum (Laj et al., 2004; Laj et al., 2000; Nowaczyk et al., 2013) and calculate the resulting ^{10}Be production rate changes using the production rate models by Masarik and Beer (1999) and Herbst et al. (2016) (Fig. 3 a-c). Subsequently, we scale the ice-core ^{10}Be record to minimize the root mean square error (RMSE) between ice-core and geomagnetic field-based records (Fig. 3d). It can be seen that the RMSE reaches a minimum for a ^{10}Be scaling factor of ~1 (for Masarik and Beer, 1999) and ~1.3 (for Herbst et al., 2016). This represents a fortunate coincidence; irrespective of which production rate model is used, the amplitude of the ice-core ^{10}Be variations has to be increased by approximately 30% to match ^{14}C. If the production rate model by Masarik and Beer is used, then the amplitude of the ice-core ^{10}Be record is in agreement with geomagnetic field data, but due to the higher production sensitivity of ^{14}C (see above), ^{10}Be variations have to be increased by ~30%. Similarly, if the production rate model by Herbst et al. is used, then the amplitude of the ice-core ^{10}Be record is 30% smaller than implied by geomagnetic field data (possibly due to a polar bias), while the sensitivity of ^{14}C and ^{10}Be is the same. Again, the net effect is the ^{10}Be variations have to be scaled up by 30% for the comparison to ^{14}C.

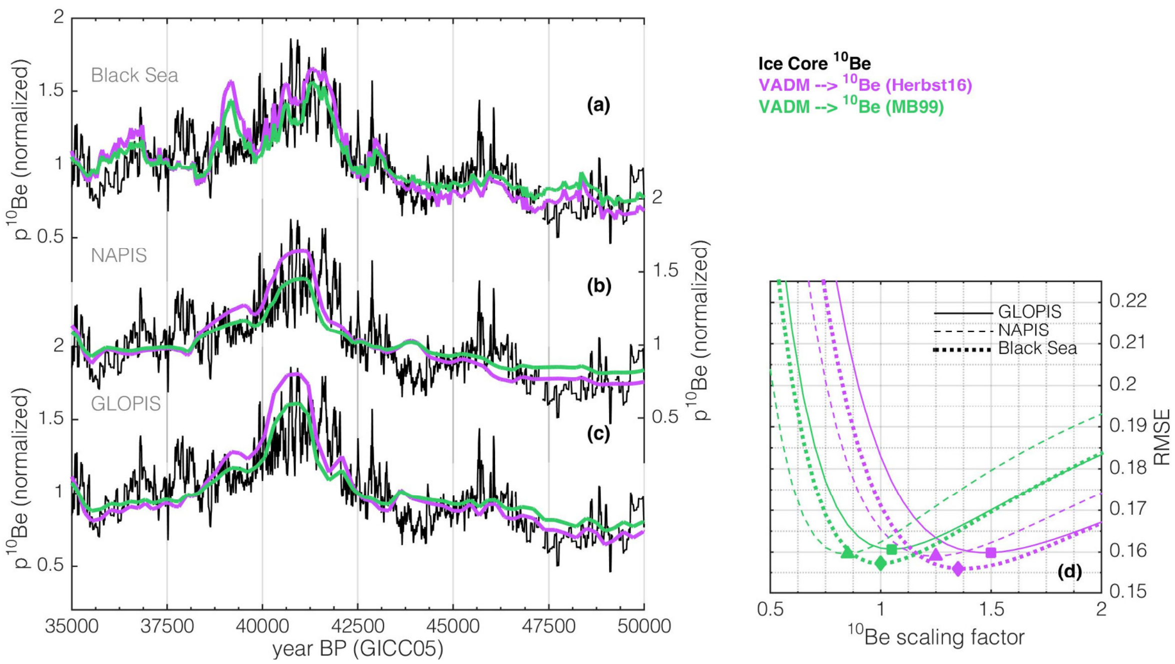

Figure 3: Comparison of ice-core-based and geomagnetic-field-based reconstructions of ^{10}Be production rates. Panel a-c show the ice-core stack (black) in comparison to ^{10}Be production rates based on geomagnetic field reconstructions and 2 different production rate models (Herbst et al. (2016) in pink and Masarik and Beer (1999) in green). Panel a, the Black Sea geomagnetic field record (Nowaczyk et al., 2013), Panel b, the NAPIS geomagnetic field stack (Laj et al., 2000), and Panel c, the GLOPIS geomagnetic field stack (Laj et al., 2004). Panel d shows the RMSE between the ice-core data and the geomagnetic-field-based records when variations in the ice-core record are scaled by different factors (x-axis). The colours correspond to the production rate models. The line styles indicate the geomagnetic field records (see legend) and the symbols denote the RMSE minima.

3.3.2 The state of the carbon cycle

As mentioned in section 2, a quantification of transient carbon cycle changes and their influence on Δ^{14}C is challenged by insufficient knowledge of inventories and processes. The contribution of single processes to Δ^{14}C changes over the last glacial cycle is likely within 30‰ and, due to compensating effects, also their combination is likely not bigger than 40‰ (Köhler et al., 2006). Here we use the Laschamp event to estimate the state of the ocean ventilation around 40 ka BP.

The datasets underlying IntCal13 all show an increase of about 320‰ in Δ^{14}C into the Laschamp event (Fig. 4), albeit at different absolute levels (see Fig. 1). This is ~100‰ more than the compiled IntCal13 curve itself implies. This disagreement can be explained by differences in timing and absolute Δ^{14}C between the different datasets leading to smoothing and dampening of Δ^{14}C variations during the construction of IntCal13. Also, geomagnetic field changes yield a Δ^{14}C change more in line with the individual ^{14}C datasets than with IntCal13, even when assuming a preindustrial carbon cycle.

To estimate the mean state of the carbon cycle during this period, we run our carbon cycle model with different (constant) values of ocean diffusivity. We find that modelled and measured Δ^{14}C around the Laschamp event match best in amplitude when we run the model under conditions where ocean ventilation is reduced to ~75% of its preindustrial value (Fig. 4). This is in broad agreement with previous modelling experiments (Köhler et al., 2006; Roth and Joos, 2013) and proxy data (Henry et al., 2016).

In the following, we will use this estimate for the parameterization of our model. As mentioned above, a transient adjustment of carbon cycle parameters is uncertain and will hence not be attempted. Instead, we ascribe an associated uncertainty to the modelled Δ^{14}C based on the carbon cycle sensitivity experiments by

Köhler et al. (2006). Furthermore, it should be noted, that by only using (filtered) $\Delta^{14}C$ anomalies as synchronization targets, we i) avoid systematic carbon cycle influences on $\Delta^{14}C$ levels, and ii) minimize transient carbon cycle related changes in $\Delta^{14}C$ (Adolphi and Muscheler, 2016).

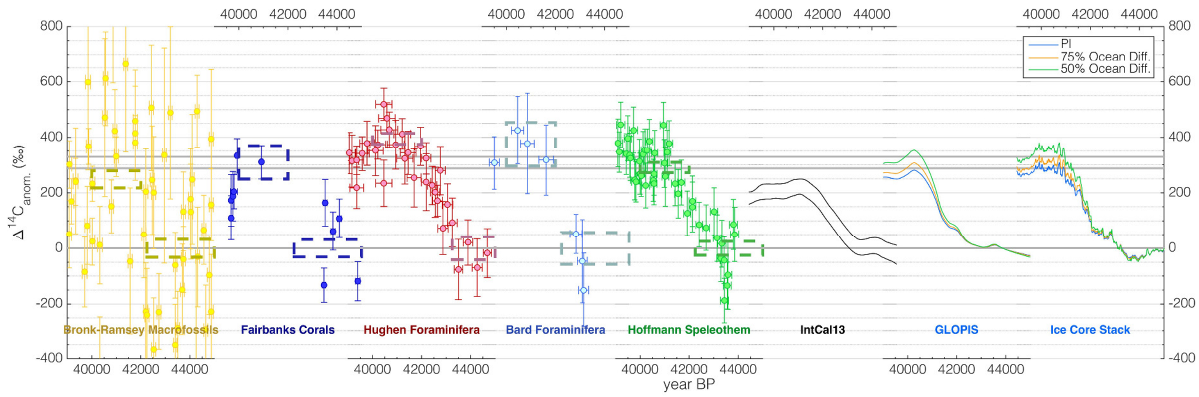

Figure 4: The Laschamp event in measured and modelled $\Delta^{14}C$. The 6 panels to the left show $\Delta^{14}C$ anomalies from macrofossils from Lake Suigetsu (yellow, Bronk Ramsey et al., 2012), tropical corals (blue, Fairbanks et al., 2005), foraminifera from Cariaco Basin sediments (red, Hughen et al., 2006), foraminifera from Iberian Margin sediments (light blue, Bard et al., 2013), Bahamas speleothems (green, Hoffmann et al., 2010), and IntCal13 (black, Reimer et al., 2013). All data are shown as anomalies to their error-weighted mean prior to the Laschamp event. i.e., the $\Delta^{14}C$ increase. The dashed boxes encompass the time periods and $\Delta^{14}C$ uncertainties (error of the error weighted mean) used for the definition of the pre-and post-Laschamp event levels. The two panels on the right show modelled $\Delta^{14}C$ using the GLOPIS (Laj et al., 2004) geomagnetic field record as well as the ice-core stack as production rate inputs. The different coloured lines reflect different carbon cycle scenarios (see legend, PI denotes pre-industrial). The conversion of geomagnetic field intensity to ^{14}C production rate is based on the production rate model by Herbst et al. (2016). Note, that the amplitude of the ^{10}Be variations have been increased by 30% as discussed in section 3.3.1.

3.4 Synchronization – effects of the carbon cycle and the archive

The synchronization method follows Adolphi and Muscheler (2016) and is outlined and tested in detail therein. In brief, sections of modelled (ice-core based) $\Delta^{14}C$ anomalies are compared to the measured $\Delta^{14}C$. For our analysis we employ high-frequency changes in $\Delta^{14}C$ since carbon cycle changes have only limited effects on atmospheric $\Delta^{14}C$ on shorter time scales (Adolphi and Muscheler (2016). Similarly, as shown in figure 2, the agreement of the different ice-core records is better on shorter timescales. In this study, we employ two types of high pass filtering: a FFT-based high-pass filter and simple linear detrending. The choice of filter is based on the data sampling resolution. For the highly resolved tree-ring data we use a 1000 year high-pass FFT filter, while the lower resolved and more unevenly sampled coral/speleothem/macrofossil data is filtered by linear detrending to avoid the interpolation to equidistant resolution required for FFT analysis. Similarly, the high sampling resolution of the tree-ring data allows us to compare the data in 2,000 year windows, while we increase the window length to 4,000 and 5,000 years for the lower resolved data prior to 14ka BP. The exact frequencies and window lengths are also given in the results section. Using the same statistics as for radiocarbon wiggle-match dating (Bronk Ramsey et al., 2001), we then infer a probability density function (PDF) for the timescale difference between the modelled and measured $\Delta^{14}C$ records. For details of the statistics of this methodology we refer the reader to Adolphi and Muscheler (2016). Here we focus instead on additional uncertainties that arise when comparing modelled atmospheric $\Delta^{14}C$ to ^{14}C records from the ocean mixed layer (corals) or speleothems.

$\Delta^{14}C$ variations in the atmosphere are dampened and delayed compared to the causal production rate changes. Both factors, attenuation and delay, depend on the frequency of the production rate change (Roth and Joos, 2013; Siegenthaler et al., 1980). The dampening is largest at high frequencies and decreases with longer periods. On the other hand, the apparent peak-to-peak delay between sinusoidal production rate changes and the resulting $\Delta^{14}C$ change is increasing with increasing wavelengths. Similar effects occur when comparing atmospheric and oceanic $\Delta^{14}C$ changes to each other: the ocean reacts to atmospheric $\Delta^{14}C$ changes with a delayed and dampened response that is wavelength dependent. Hence, we need to take these factors into account when comparing a modelled atmospheric $\Delta^{14}C$ record to mixed layer marine records. However, the frequency dependence of the attenuation and delay makes it difficult to explicitly correct for this since atmospheric $\Delta^{14}C$ changes vary on different time scales simultaneously. Furthermore, the coral records vary in their sampling frequency and often it is not precisely known over how much time an individual ^{14}C sample integrates.

Figure 5 shows a sensitivity test regarding these effects. We modelled $\Delta^{14}C$ from the ice-core stack around the Laschamp event and compared the atmospheric $\Delta^{14}C$ to the mixed layer $\Delta^{14}C$ in the model. To simulate the effect of varying averaging effects of the coral samples, we low-pass filtered the mixed layer signal with increasing cut-off wavelengths. For each filter, we then inferred the apparent delay between the mixed layer (i.e., the "coral") and the atmosphere. In doing so we infer that even though the signal is dominated by a long lasting $\Delta^{14}C$ increase, the inferred delay is small (~30 years) as long as the coral samples do not integrate over long times. Only when assuming that each coral sample averages over more than 1,000 years we infer delays of about 120 years. Nevertheless, this experiment also shows that within reasonable bounds of averaging, the delay of mixed layer to atmospheric signal is limited

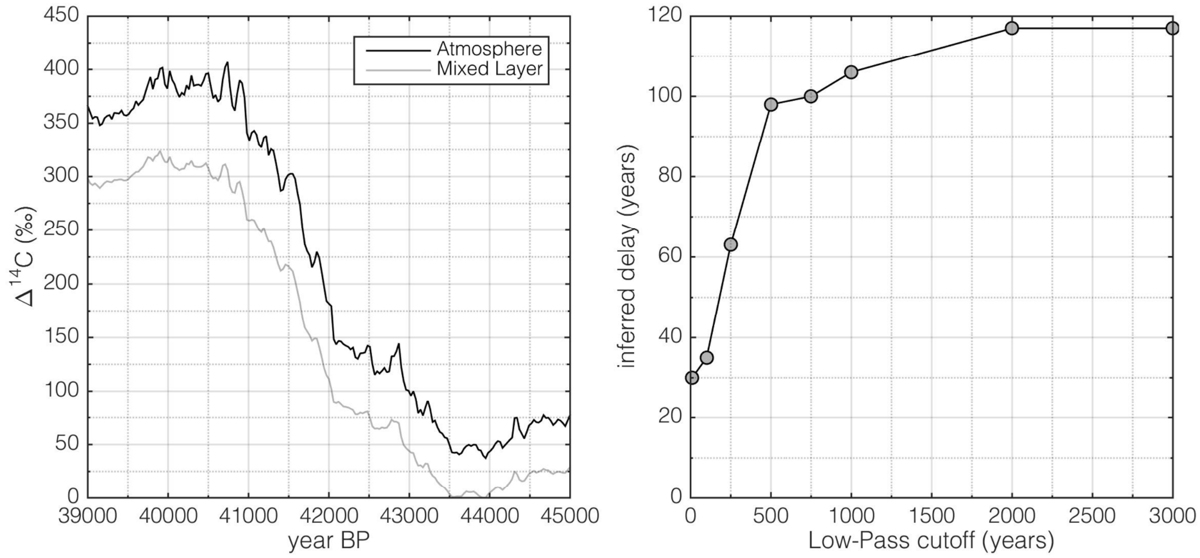

Figure 5: The delay between $\Delta^{14}C$ in the atmosphere and ocean mixed layer. The left panel shows modelled $\Delta^{14}C$ from the ice-core stack around the Laschamp event. The modelled atmospheric $\Delta^{14}C$ is shown in black while ocean mixed layer is shown in grey. They right hand panel shows the inferred delay from our synchronization method when comparing the atmospheric to the mixed layer signal for different low-pass filters of the mixed layer signal (x-axis).

The speleothem $\Delta^{14}C$ reacts differently than the ocean mixed layer. The so-called dead carbon fraction (DCF) of a speleothem consists of two main contributors: i) respired soil organic matter that is older (in ^{14}C years) than the atmospheric ^{14}C signal, and ii) carbonate bedrock that contains no ^{14}C. Applying the model of Genty and

Massault (1999), we model speleothem $\Delta^{14}C$ using different assumptions on the age of the respired soil organic matter and fraction of carbonate bedrock in drip water CO_2. We do this for 2 examples: i) a speleothem with an apparent DCF (i.e., offset from the atmosphere) of 5.8% (resembling the Hulu Cave speleothem record by Southon et al., 2012) and ii) a speleothem with an apparent DCF of 25.7% (resembling the Bahamas speleothem by Hoffmann et al., 2010). By assuming different ages of the soil respired carbon (τ = 10 – 400 years, see Fig. 6), we adjust the fraction of ^{14}C-free CO_2 so that the apparent DCF for each speleothem is matched. The age of the soil respired carbon is defined following Genty and Massault (1999): if, for example, τ = 100 years, then the activity of the soil respired CO_2 is the mean of the atmospheric activity over the past 100 years prior to sampling (also accounting for decay within these 100 years). For simplicity we assume a uniform age distribution for the soil respired carbon. Subsequently, we compare the modelled speleothem $\Delta^{14}C$ to the original atmospheric input using our synchronization method and plot the inferred delay (Fig. 6, right panel). From this experiment it can be seen that the controlling factor on the inferred delay is the age of the soil respired matter that acts as an integrator (low-pass filter) of the atmospheric ^{14}C signal. The fraction of ^{14}C-free carbonate has no influence on the lag between $\Delta^{14}C$ changes in the atmosphere and the speleothem, but only dampens the amplitude of the corresponding change. Realistic ages of soil respired carbon differ from region to region but even though some slow cycling fractions of soil organic matter may be up to several thousand years old (Trumbore, 2000), the major contributors to soil CO_2 are considerably younger and in the order of decades (Genty et al., 2001; Fohlmeister et al., 2011).

From these experiments we conclude that our systematic matching uncertainties to coral and speleothem records are probably below 100 years. We note that this uncertainty is asymmetric since the ocean/speleothem signal cannot lead the atmosphere and so the offset is unidirectional.

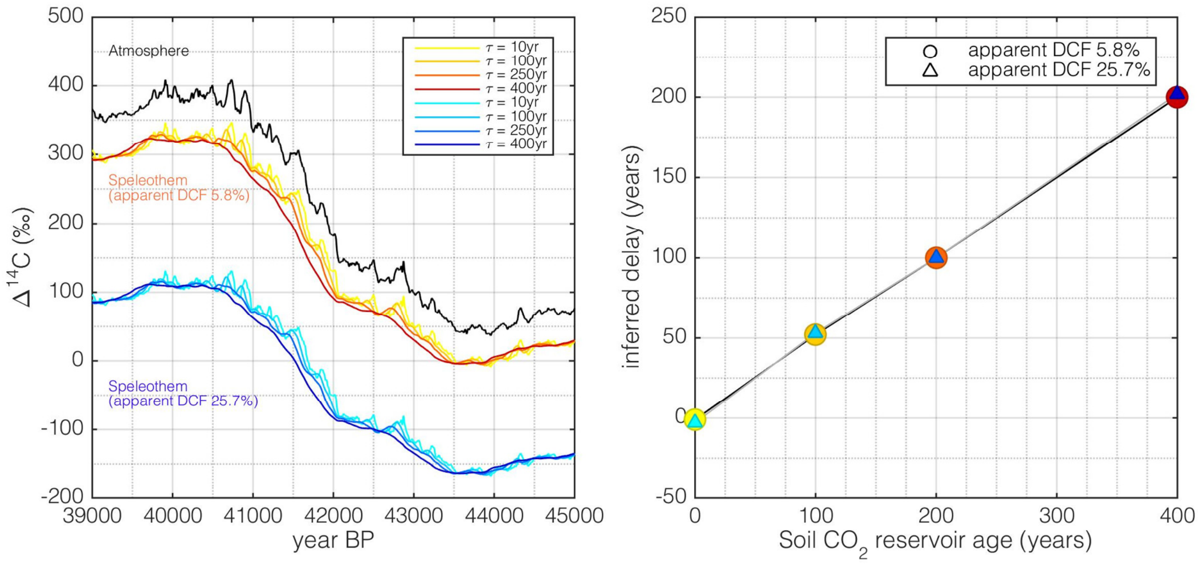

Figure 6: Effect of varying ages of soil respired CO_2 and fractions of CO_2 from ^{14}C-dead carbonate on the $\Delta^{14}C$ in speleothems. The left panel shows atmospheric modelled $\Delta^{14}C$ from the ^{10}Be stack (black) and two modelled speleothem scenarios with a net DCF of 5.8% (warm colours) and 25.7% (cold colours). For each speleothem, a number of different ages for the respired soil organic matter have been assumed (see legend) and the input of ^{14}C-free CO_2 from carbonate has been adjusted to obtain the correct apparent DCF value between 39-40.5 ka BP. The right hand panel shows the inferred delay when we apply our synchronization method to match the atmospheric $\Delta^{14}C$ to the speleothem record.

3.5 Change-point detection in climate records

To test the synchroneity of rapid climate changes, we compare the timing of DO-events seen in Greenland ice cores (Andersen et al., 2004), to a number of well-known U/Th dated speleothems that show DO-type variability from Hulu Cave (Cheng et al., 2016), Sofular Cave (Fleitmann et al., 2009), El Condor, and Cueva del Diamante (both Cheng et al., 2013b).

We use a probabilistic model to detect the onset, mid-point, and end of the rapid climate transitions in each individual record. The employed model describes the abrupt changes as a linear transition between two constant states. Any variability due to the long-term fluctuations of the climate records around the transition model is described by an AR(1) process that is estimated in conjunction with the transition model. The model is independently fitted to windows of data on their individual timescales (Table 1 & Fig. 13) around the rapid transitions. Inference was performed using Markov Chain Monte Carl sampling (MCMC) to obtain PDFs of the timing of the onset, the length, and the amplitude of each transition in each record. Using these PDFs we can calculate delays of the onset, mid-point and end of the climate transitions between different records, propagating the respective uncertainties of the parameters. For each record, only events that are well expressed and measured in high resolution have been fitted. The approach and inference procedure are described in more detail in Erhardt et al. (submitted).

Table 1. Change-point detection window for each record. For each investigated climate event and record, the change-point detection algorithm has been applied between t1 and t2. The windows have been defined visually, ensuring a sufficient amount of data prior to and after the transition. For each record, only events that are well expressed in the climate proxy records at high resolution have been investigated. For the ice-core record t1 and t2 typically encompass 500 years prior to and after the nominal transition ages by Rasmussen et al. (2014a). The exact values have been adjusted to exclude overlap with other transitions where necessary (Erhardt et al. in prep).

			Hulu d18O		Sofular d18O		Sofular d13C		ElCondor d18O		Diamante d18O	
Event	**GICC05 (yr BP)**	**MCE**	**t1**	**t2**	**t1**	**t2**	**t1**	**t2**	**t1**	**t2**	**t1**	**t2**
Holocene	11653	99	12453	10503	12703	10703	12703	11003	12453	11203	13403	11203
GI-1e	14642	186	15442	13942	15442	13942	15442	13942	15442	14192	16392	14192
GS-3 Dust Peak 1	24130	645	25380	24080	-	-	-	-	-	-	25780	24630
GI-3	27730	832	28580	27680	28780	27880	28780	27780	-	-	29030	28080
GI-4	28850	898	30100	28900	30150	29400	30150	29200	29900	29000	30100	29100
GI-5.1	30790	1008	31540	30790	-	-	-	-	31590	30740	32040	30840
GI-5.2	32450	1132	33300	32200	33100	32400	33300	32200	33250	32000	33050	32450
GI-6	33690	1195	34590	33640	34740	33690	34990	33540	34240	33490	-	-
GI-7c	35430	1321	36680	34980	36380	35480	36380	35230	36230	34880	36480	34980
GI-8c	38170	1449	39420	37420	39420	37220	39120	37220	39220	37220	-	-
GI-9	40110	1580	40860	40060	40960	39960	41160	39960	-	-	-	-
GI-10	41410	1633	42110	41060	42460	41590	42460	41460	42210	40960	-	-
GI-11	43290	1736	44240	42940	44840	43540	-	-	44040	42440	-	-

4 Time scale differences between GICC05 and the U/Th timescale

In the following sections we will show the synchronization results for different time windows. We focus our analysis on three distinct windows: 10-14 ka BP, 18-25 ka BP and 39-45 ka BP. The youngest window is defined by the presence of high-resolution tree-ring data for ^{14}C back to 14 ka BP. Going further back in time it becomes increasingly challenging to unequivocally identify common structures in the various $\Delta^{14}C$ records that are suitable for synchronization because the resolution of the individual records decreases back in time while their differences to each other are growing steadily (see Fig. 1i). Hence, we focus on the well-known Laschamp event around 41 ka BP, and the period between 18-25kaBP, i.e., preceding the major carbon cycle changes associated with the deglaciation. We omit the period between 25-39 ka BP. As discussed in Reimer et al. (2013) and seen in figure 1i there is substantial disagreement between the datasets underlying IntCal13 at that time that are impossible to reconcile within their respective age and/or ^{14}C uncertainties. Hence, also any structure in the Δ^{14}C records may be unreliable and thus, lead to erroneous synchronization results.

4.1 10,000 – 14,000 years BP

In the 10-14 ka BP interval, we synchronize the ice-core stack to high-resolution tree-ring and speleothem Δ^{14}C data (Fig. 7). The high sampling resolution of the ^{14}C records allows us to focus on centennial-to-millennial Δ^{14}C changes (<1000 years) where carbon cycle influences on Δ^{14}C can be expected to be small (Adolphi and Muscheler, 2016). In concordance with earlier studies (Muscheler et al., 2014a) we find that GICC05 is ~65 years older than the tree-ring timescale at the onset of the Holocene, but that this offset vanishes over the course of the Younger Dryas interval.

While Muscheler et al. (2014a) argued that this changing offset may be in part due to errors in the timescale of the floating Late Glacial Pines, we can now support this change in the timescale-difference through the U/Th dated speleothems: The synchronization of the ice-core stack to the H82 speleothem from Hulu Cave (Southon et al., 2012) leads to fully consistent results as inferred from the tree-rings. This indicates that the most likely explanation is an ice-core layer counting bias, i.e. that the GICC05 time scale suggests too old ages at the onset of the Holocene, but is accurate within a few decades during GI-1.

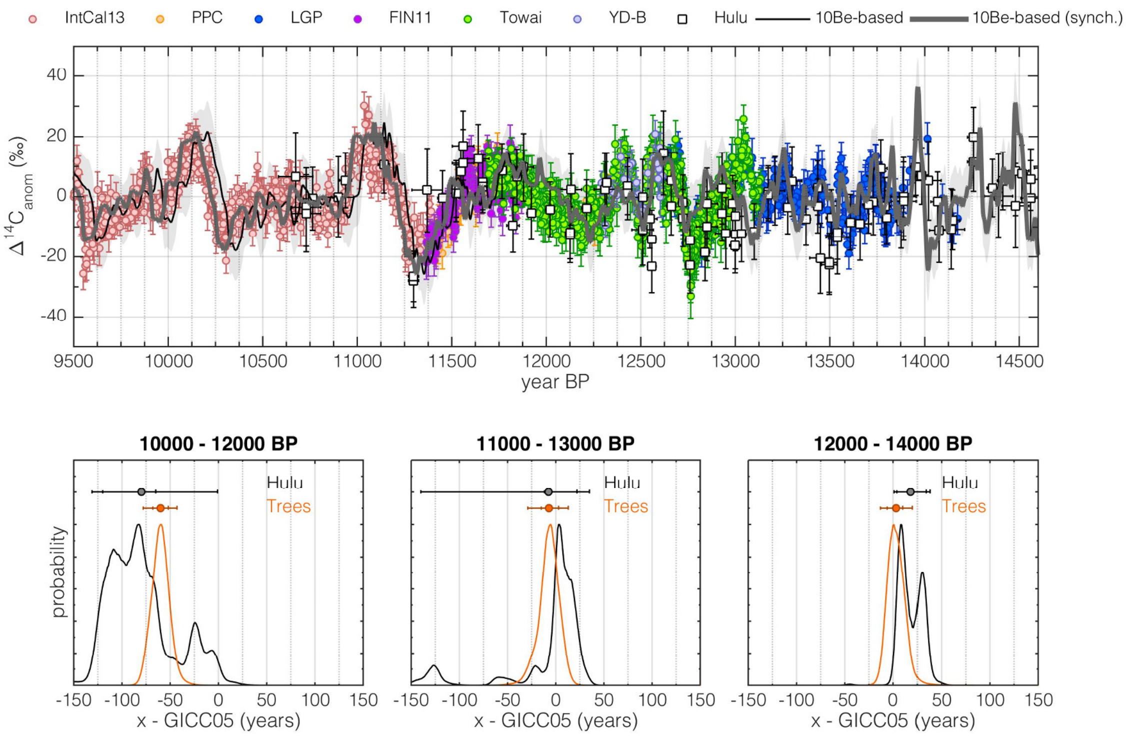

Figure 7: Synchronization of GICC05 to tree-ring and Hulu Cave records during the last deglaciation. Top panel: Ice-core based modelled Δ^{14}C anomalies on the original GICC05 timescale (thin black line, light grey shading encompasses the ±10‰ uncertainty (±1σ) of the modelled Δ^{14}C, based on the carbon-cycle sensitivity experiments by Adolphi & Muscheler (2016)) and synchronized timescale (bold grey line). Tree-ring data underlying IntCal13 are shown in pink. Revised Northern Hemisphere tree-ring data according to Hogg et al. (2016) are shown in orange (Preboreal Pines), dark blue (Late Glacial Pine) and light blue (Younger Dryas-B chronology). New kauri Δ^{14}C data by Hogg et al. (2016) is shown in purple (FIN11) and green (Towai). Hulu Cave H82 Δ^{14}C data are shown as white

squares. All symbols are shown with ±1σ error bars. All data are FFT-filtered to isolate $\Delta^{14}C$ variations on timescales <1000 years. The lower three panels show inferred probability distributions of timescale differences between GICC05 and tree-rings (orange) and Hulu Cave (black). The symbols and error bars denote means, and 68.2% and 95.4% confidence intervals of the inferred timescale difference. Each of the lower panels refers to a 2000-year subsection of the data indicated at the top of each panel.

Interestingly, we do not observe any significant differences between the results stemming from tree-rings and the speleothem records. As shown in section 3.4, we could expect a delay in the speleothem $\Delta^{14}C$ compared to the atmosphere if the respired soil organic carbon contribution to the soil CO_2 was very old. This would result in GICC05 appearing older in comparison to the speleothem than relative to the tree rings. The lack of this delay implies that the majority of the respired soil organic carbon at Hulu Cave must be younger than ~100 years (see Fig. 6). This is supported by the fact that the centennial $\Delta^{14}C$ variations in the tree-ring and speleothem data have the same amplitude (Fig. 7). If old organic carbon significantly contributed to the soil CO_2, we would instead expect to see a stronger smoothing of short-term $\Delta^{14}C$ variations.

4.2 18,000 – 25,000 years BP

Due to the irregular and lower sampling resolution of the ^{14}C records beyond 15,000 cal BP, we chose to linearly detrend each data set (instead of band-pass filtering) to remove offsets between the different ^{14}C datasets (see figure 1i) and highlight common variability. Furthermore, we have to increase the length of the comparison data windows to 4,000 years to ensure sufficient structure in the ^{14}C sequences entering the comparison. Each window is detrended separately in the analysis to isolate short-term $\Delta^{14}C$ variability. We note however, that detrending each ^{14}C dataset over the entire timeframe (18-25 ka BP) instead does not alter the results significantly. Compared to the high-frequency $\Delta^{14}C$ changes studied between 10-14 ka BP, the longer-term variations used for synchronization here may have been increasingly affected by carbon-cycle changes. To account for this, we increase the uncertainty estimate of the modelled $\Delta^{14}C$ changes to ±30‰ (±1σ), which is sufficiently large to account for estimated carbon-cycle-driven $\Delta^{14}C$ changes from modelling experiments during the entire glacial (Köhler et al., 2006). We note that this is a conservative estimate, given that during this period neither modelling (Köhler et al., 2006; Muscheler et al., 2004), nor data (Eggleston et al., 2016) suggest large carbon-cycle changes.

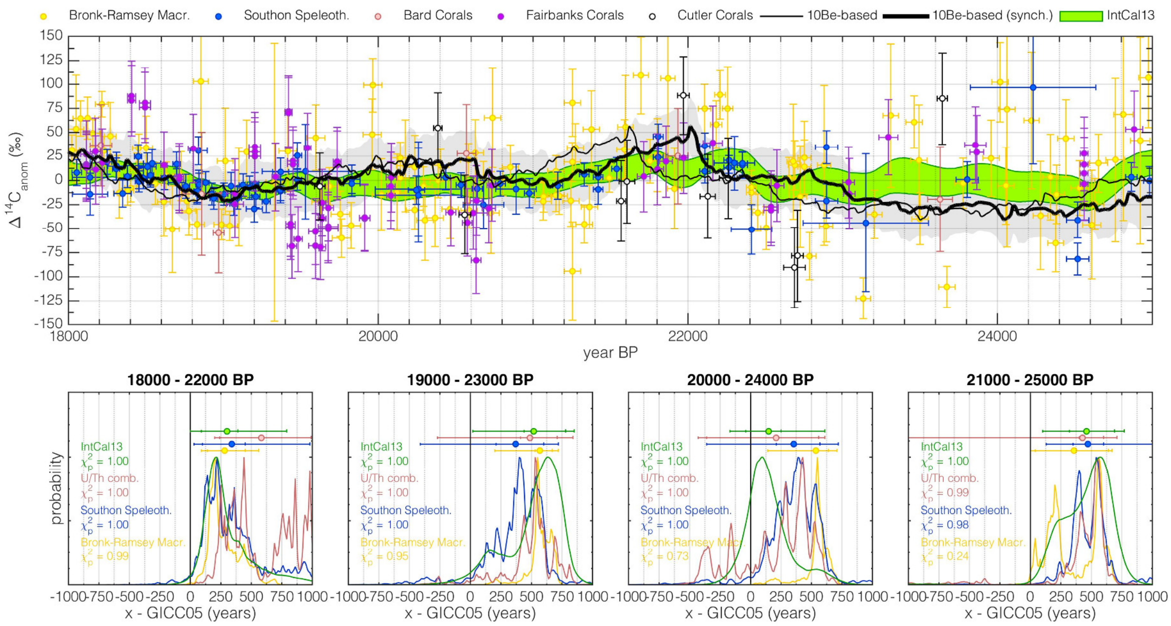

Figure 8: Synchronization results between 18,000 and 25,000 years BP. Top panel: The thin black line shows the modelled $\Delta^{14}C$ curve based on the ice-core stack on its original timescale. The bold black line and grey shading show the synchronized ice-core record including assumed ±1σ uncertainties of ±30‰. The different coloured symbols indicate various ^{14}C datasets underlying IntCal13, which is shown as the green envelope. Lower panels: Each panel shows PDFs of the inferred timescale difference between the ice-core stack and IntCal13 (green), a combination of all U/Th-dated records (speleothems/corals, pink), the H82 speleothem (blue), and Lake Suigetsu (yellow). Symbols of similar colour show the inferred mean and 68.2% and 95.4% confidence intervals. Colour-coded text indicates χ^2 probabilities for the goodness of fit between modelled and measured $\Delta^{14}C$ curves after synchronization. Small (e.g., <0.1) values would indicate significant disagreement. Note that all χ^2 probabilities are relatively high, indicating that our uncertainty estimate for the modelled $\Delta^{14}C$ is very conservative. Each of the lower panels refers to a specific subsection of the data indicated at the top of each panel.

It can be seen in figure 8 that it is challenging to infer robust co-variability in multiple ^{14}C records. However, the millennial evolution of $\Delta^{14}C$ does show common changes in the 18-25 ka BP interval. Synchronizing the ice-core stack to data from i) Hulu Cave H82 speleothem, ii) Lake Suigetsu macrofossils, iii) the IntCal13 stack or iv) a combination of all U/Th dated records (speleothems/corals) leads to consistent results within uncertainties for each choice of time windows: all records imply that GICC05 shows younger ages compared to the ^{14}C records around this time.

The most significant structure that is present in all measured and modelled ^{14}C records during this time is the centennial $\Delta^{14}C$ increase around 22.1kaBP (see Fig. 9). Comparing the ice-core stack to $\Delta^{14}C$ between 21-23kaBP indicates an offset of ~550 years between GICC05 and the U/Th timescale around this time (GICC05 being younger). To account for the potential delay of coral and speleothem $\Delta^{14}C$ compared to the atmosphere, we also modelled the mixed layer $\Delta^{14}C$ signal from the ice-core stack and synchronized this signal to the measured ^{14}C data (Fig. 9). As discussed in section 3.4, we find very little difference in the inferred timing since the $\Delta^{14}C$ variation is relatively rapid (centuries). Comparing the $\Delta^{14}C$ anomalies to geomagnetic field data shows that a small part of the longer-term development of this structure is probably driven by geomagnetic field changes. The amplitude (~50‰) and short duration (centuries) of the $\Delta^{14}C$ increase, however, suggest that this change is mainly driven by a series of strong solar minima, comparable to the Grand Solar Minimum period around the onset of the Younger Dryas (Muscheler et al., 2008). We used this tie-point (figure 9) in the final synchronization as it is the best-defined feature in this time interval, and consistent within error with the estimates shown in figure 8.

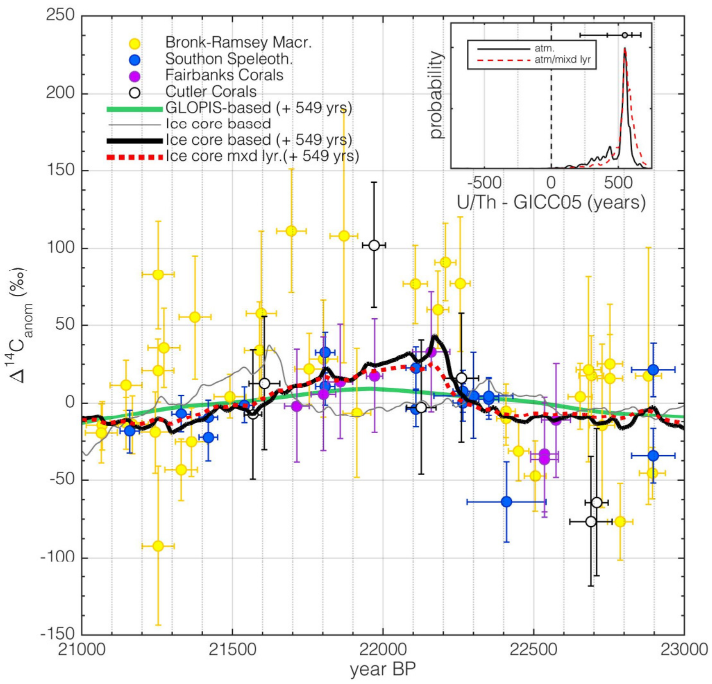

Figure 9: Close-up of measured and modelled $\Delta^{14}C$ anomalies between 21 and 23 ka BP. The thin grey line shows modelled atmospheric $\Delta^{14}C$ from the ice-core stack on the GICC05 time scale. The bold black and dashed red lines show the modelled atmospheric and ocean mixed layer $\Delta^{14}C$ curves after synchronization to the ^{14}C records (yellow: Lake Suigetsu; blue: Hulu Cave; purple and white: corals. The inset panel shows the PDF of the inferred timescale difference between GICC05 and the combination of all ^{14}C records. The black line is based on using only the modelled atmospheric $\Delta^{14}C$. The red dashed line is based on comparing coral and speleothem data to the modelled mixed-layer $\Delta^{14}C$, and Lake Suigetsu data to modelled atmospheric $\Delta^{14}C$. The green line shows modelled $\Delta^{14}C$ based on geomagnetic field changes.

4.3 39,000 – 45,000 years BP

Our oldest tie-point is the previously discussed Laschamp event around 41 ka BP. The only independently and absolutely dated ^{14}C record around this time that has a sufficient sampling resolution is the Bahamas speleothem by Hoffmann et al. (2010). While offset in absolute $\Delta^{14}C$ (see Fig. 1), the U/Th-dated coral data supports the amplitude and timing of the $\Delta^{14}C$ increase seen in the speleothem even though precise synchronization is hampered by the low sampling resolution of the corals. The Lake Suigetsu record is characterized by large uncertainties and scatter around this time. As discussed in section 3.3.2, IntCal13 is smoothed around Laschamp, having a smaller amplitude and a less sharp rise in $\Delta^{14}C$. For this tie-point, we merely remove the error-weighted mean between 39-45 ka BP from each dataset, since detrending would remove the largest part of the signal. Hence, there are large $\Delta^{14}C$ modelling uncertainties associated with unknown carbon-cycle changes, and we assume a Gaussian ±1σ error of 50‰, which we consider conservative since sensitivity experiments imply that the impact of carbon cycle changes on $\Delta^{14}C$ was likely below 40‰ (Köhler et al., 2006).

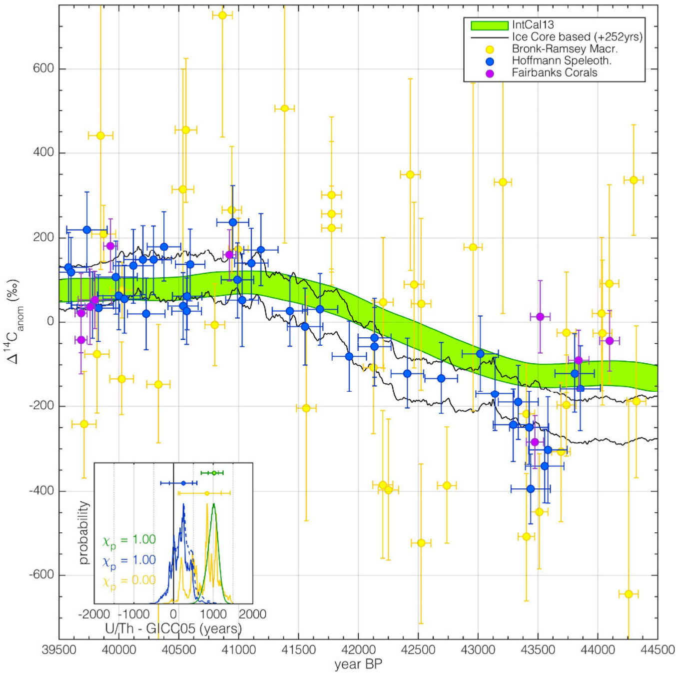

Figure 10. Synchronization of ^{10}Be and ^{14}C around the Laschamp event. The black lines encompass the modelled $\Delta^{14}C$ anomalies (±1σ) from the ice-core data shifted by +252 yrs (68.2% confidence interval = -103 to 477 yrs) according to their best fit to the speleothem ^{14}C data. The green patch shows the ±1σ envelope of IntCal13. The blue and purple symbols show $\Delta^{14}C$ from Bahamas speleothem, and corals, respectively. The yellow symbols show $\Delta^{14}C$ anomalies based on Lake Suigetsu macrofossils. All datasets have been centred to 0‰ by subtracting the error-weighted mean of each dataset. The inset shows the PDF of the inferred age differences between the ice-core data and IntCal13 (green), Lake Suigetsu (yellow) and the Bahamas speleothem (blue). The dashed blue line corresponds to age differences from the modelled mixed layer $\Delta^{14}C$ and the Bahamas speleothem.

Synchronizing the ice-core stack to the speleothem, Lake Suigetsu, and IntCal13 data yields significantly different results. We infer that GICC05 produces ages about 250 years younger than the U/Th dated speleothem data (Fig. 10). The IntCal13 record however, implies a larger difference of ~1,000 years. Using Lake Suigetsu data, on the other hand, leads to multiple probability peaks of which two are in agreement with the speleothem, and one with the IntCal13 record. The large scatter of the Lake Suigetsu data however, leads to poor statistics (low χ^2 probabilities). Furthermore, the Lake Suigetsu timescale is only constrained by varve counting back to 39 ka BP and based on extrapolation for older sections (Bronk Ramsey et al., 2012) and hence, provides less precise constraints on the timing of the $\Delta^{14}C$ increase.

To test which of these links is the most likely we turn to independent radiometric ages of the Laschamp excursion. Pooled Ar-Ar, K-Ar, and U/Th ages on lava flows place the period of (nearly) reversed field direction at 40,700 ± 950 yr BP (Singer et al., 2009), or 41,300 ± 600 yr BP (Laj et al., 2014). In addition, a North American speleothem provides a U/Th-dated transient evolution of the geomagnetic field (Lascu et al., 2016), with the lowest intensities occurring at 41,100 ± 350 yr BP. Comparing the ice-core ^{10}Be stack to these data clearly shows that all of these records rule out the +1,000 year time shift implied by IntCal13, as it would induce a significant disagreement between radiometrically dated magnetic field records and the dating of the ^{10}Be peak in the ice cores (Fig. 11). We hence argue that the 252 yr offset inferred from the comparison to the Bahamas speleothem is the most likely estimate of the timescale difference between GICC05 and the U/Th timescale around this time. Similar as before, assuming that the speleothem represents a mixed-layer signal instead of direct atmospheric $\Delta^{14}C$ does not significantly affect the inferred timescale differences (see Fig. 10 inset, blue dashed line).

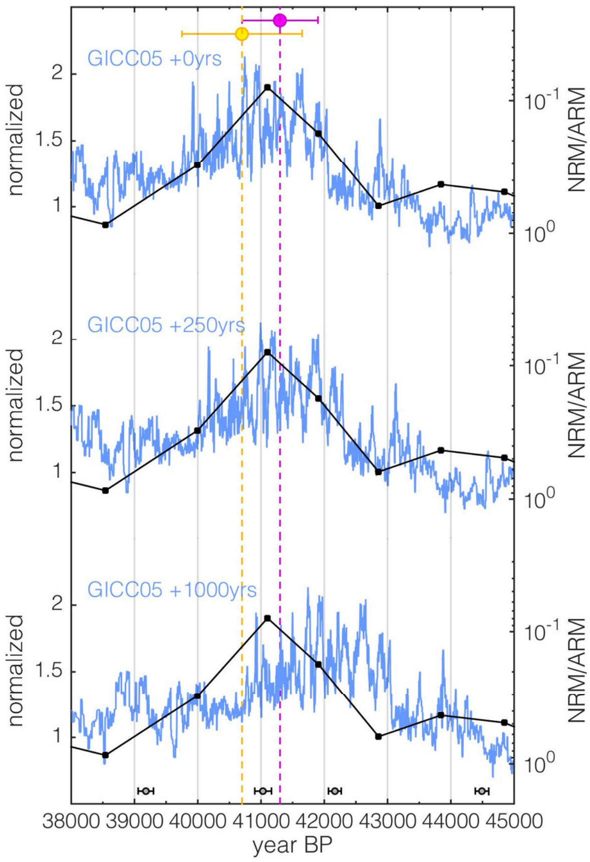

Figure 11: Comparison of the ice-core stack (blue) to Ar-Ar dates of the Laschamp excursion (yellow: Singer et al. 2009, pink: Laj et al. 2014), and relative geomagnetic field intensity (black, NRM/ARM, reversed y-axis) from a U/Th-dated speleothem (Lascu et al., 2016). The individual speleothem U/Th dates are shown on the bottom of the figure with their ±2σ uncertainties. Each panel shows a different shift of GICC05 according to the results from figure 10.

4.4 Transfer Function

To construct a continuous transfer function between GICC05 and the U/Th timescale we apply a Monte Carlo approach. Each iterations consists of i) randomly sampling the PDFs at each tie-point and ii) interpolating in between the tie-points using an AR-process that is constrained by the GICC05 maximum counting error (mce). We use the tie-points shown in figure 7, 9, and 10, i.e., three tie-points between ice cores and tree-rings during the deglaciation, one tie-point between ice cores and the combination of Corals, Speleothems and Lake Suigetsu during the LGM, and one tie-point between ice cores and the Bahamas speleothem around the Laschamp event. For the interpolation, we use the time derivative of the mce (i.e., its growth rate) as an incremental error estimate. During periods when the growth rate is > 0 GICC05 may be stretched (compressed), while a growth rate of 0 does not allow this, independent of what the absolute mce is at that time. By multiplying this growth rate with a random realization of an AR-process, we simulate how much of that uncertainty has been realized in each iteration of the Monte Carlo simulation. Subsequently integrating over the resulting timeseries of simulated miscounts, we obtain again an absolute error estimate, i.e., one possible realization of the mce. In each iteration, this realisation is then anchored at the sampled tie-points (step i) by linearly correcting the offset between the sampled tie-points and the simulated counting error. Hence, this procedure provides us with a correlated interpolation uncertainty over time, taking into account some of the constraints provided by the ice core timescale itself, but giving priority to our inferred tie-points. We note that this treatment of the mce as an AR-process leads to larger interpolation errors compared to assuming a white noise model, which would lead to very small uncertainties that average out over long time (see also discussion in Rasmussen et al., 2006). Furthermore, we treat the mce as ±1σ instead of ±2σ as proposed by Andersen et al. (2006) which additionally increases our interpolation error. We stress that this procedure does not aim to provide a realistic model of the ice-core layer-counting process and its uncertainty which is clearly more complex (see Andersen et al., 2006; Rasmussen et al., 2006), nor should it be interpreted such that the mce was a 1σ uncertainty. However, our approach allows us to infer a conservative estimate of the interpolation uncertainty while at the same time it takes advantage of the fact that GICC05 is a layer counted timescale and hence, cannot be stretched/compressed outside realistic bounds. This procedure was repeated 300,000 times which was found sufficient to obtain a stationary solution, leading to 300,000 possible timescale transfer functions.

Figure 12 shows the resulting mean transfer function along with its confidence intervals. Firstly, it can be seen that all tie-points fall into the uncertainty envelope of GICC05. The implied change in the timescale difference between the youngest two tie-points (i.e., over the course of GS-1), and between 13,000 and 22,000 years BP is slightly larger than allowed by the mce, albeit the latter is consistent within the uncertainties of the tie-point at 22,000 years BP. We can see that the use of the mce to determine the interpolation error leads to small uncertainties wherever the change in the timescale difference is large (e.g. over the 13,000 – 22,000 years BP interval): Stretching GICC05 by as much as the counting error allows, requires that every uncertain layer has in fact been a real annual layer, leaving little room for additional error. Between 22,000 and 42,000 years BP, the interpolation uncertainties are determined by the mce and thus, grow/shrink at a rate determined by the mce.

Our results are in very good agreement with the results by Turney et al. (2016) around Heinrich 3. In this study, a kauri-tree ^{14}C sequence was calibrated onto Lake Suigetsu ^{14}C and also matched on GICC05 via ^{10}Be. The difference of the inferred ages (i.e., kauri on Suigetsu vs. Kauri on GICC05) matches with our proposed transfer function (red star in Fig. 12).

Figure 12 also shows the inferred offset between the ^{40}Ar/^{39}Ar-age of the Campanian Ignimbrite (Giaccio et al., 2017) and a tentatively attributed SO_4-spike in the GISP2 ice core (Fedele et al., 2007). Even though it obviously requires a well-characterized tephra find in the ice cores to ensure that the SO_4-peak is indeed associated with the Campanian Ignimbrite, at least from a chronological point of view, our transfer function does not preclude this link. However, no matching shards were identified in this period (Bourne et al., 2013).

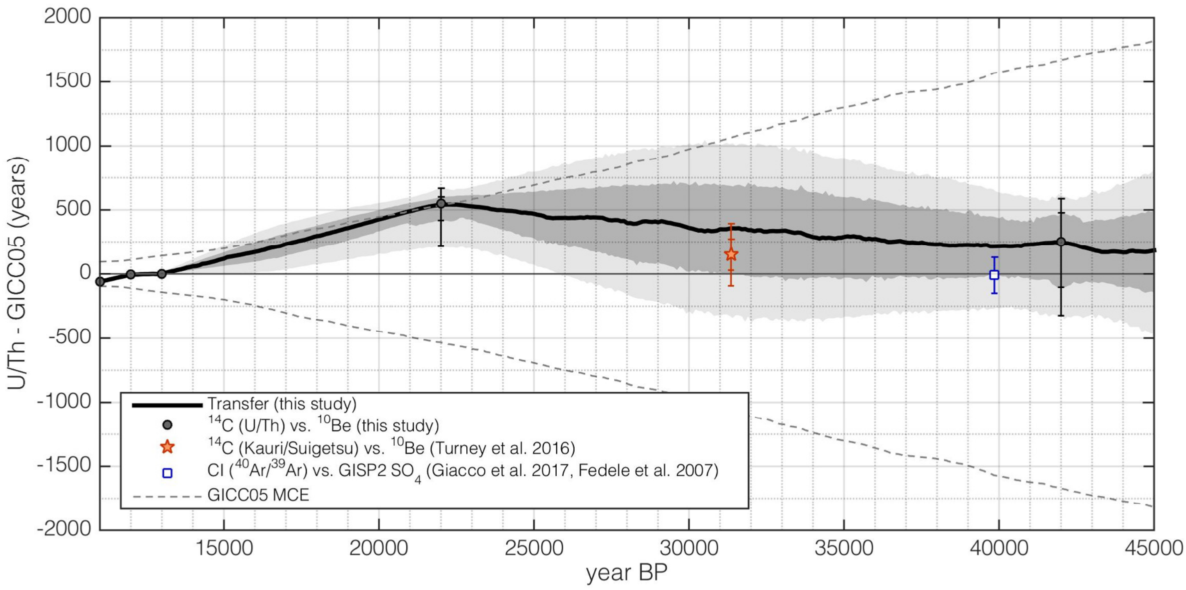

Figure 12: Transfer function between the U/Th timescale and GICC05. The transfer function is shown in black with dark and light grey shading encompassing its 68.2% and 95.4% confidence intervals. The black dots with error bars show the used match points between ^{14}C and ^{10}Be. The red star shows the difference between ages of a glacial kauri tree ^{14}C sequence on Lake Suigetsu ^{14}C and GRIP ^{10}Be (Turney et al., 2016). The blue open square shows the age difference between the $^{40}Ar/^{39}Ar$-age of the Campanian Ignimbrite (Giaccio et al., 2017), and a tentatively associated spike in the GISP2 SO_4 record (Fedele et al., 2007) on the GICC05 timescale (Seierstad et al., 2014).

5 The timing of DO-events

To investigate the synchroneity of climate changes recorded in different parts of the globe, we compare ice-core data to a selection of well-dated speleothem records. The well-known Hulu-Dongge Cave records have become iconic blueprints for intensity changes of the East Asian Summer Monsoon (EASM) anchored on a precise U/Th timescale (Cheng et al., 2016; Dykoski et al., 2005; Wang et al., 2001). The speleothem records from Cueva del Diamante and El Condor reflect changes in precipitation amount over eastern Amazonia associated with the South American Monsoon (Cheng et al., 2013b). The speleothem records from Sofular Cave, Turkey, are not straightforward in their mechanistic interpretation but likely reflect a mix of temperature and seasonality of precipitation ($\delta^{18}O$), and type and density of vegetation, soil microbial activity ($\delta^{13}C$), and hence, effective moisture and temperature (Fleitmann et al., 2009). Hence, while this list of speleothem data can certainly be expanded in future studies, we chose these four speleothem records from 3 different regions that are all well-dated and sensitive to the position of the ITCZ and compare it to the ice-core records. We used the NGRIP Ca record (Bigler, 2004), that shows the largest signal to noise ratio across DO-events (compared to e.g., $\delta^{18}O$) making their identification more precise. In addition, the Ca aerosols originate from Asian dust sources (Svensson et al., 2000) and are thus, more directly related to Asian hydroclimate (Schüpbach et al., 2018) making them potentially more comparable to for example the Hulu cave record. Potential phasing differences between different climate proxies in the ice core are small compared to our synchronization uncertainties (Steffensen et al., 2008).

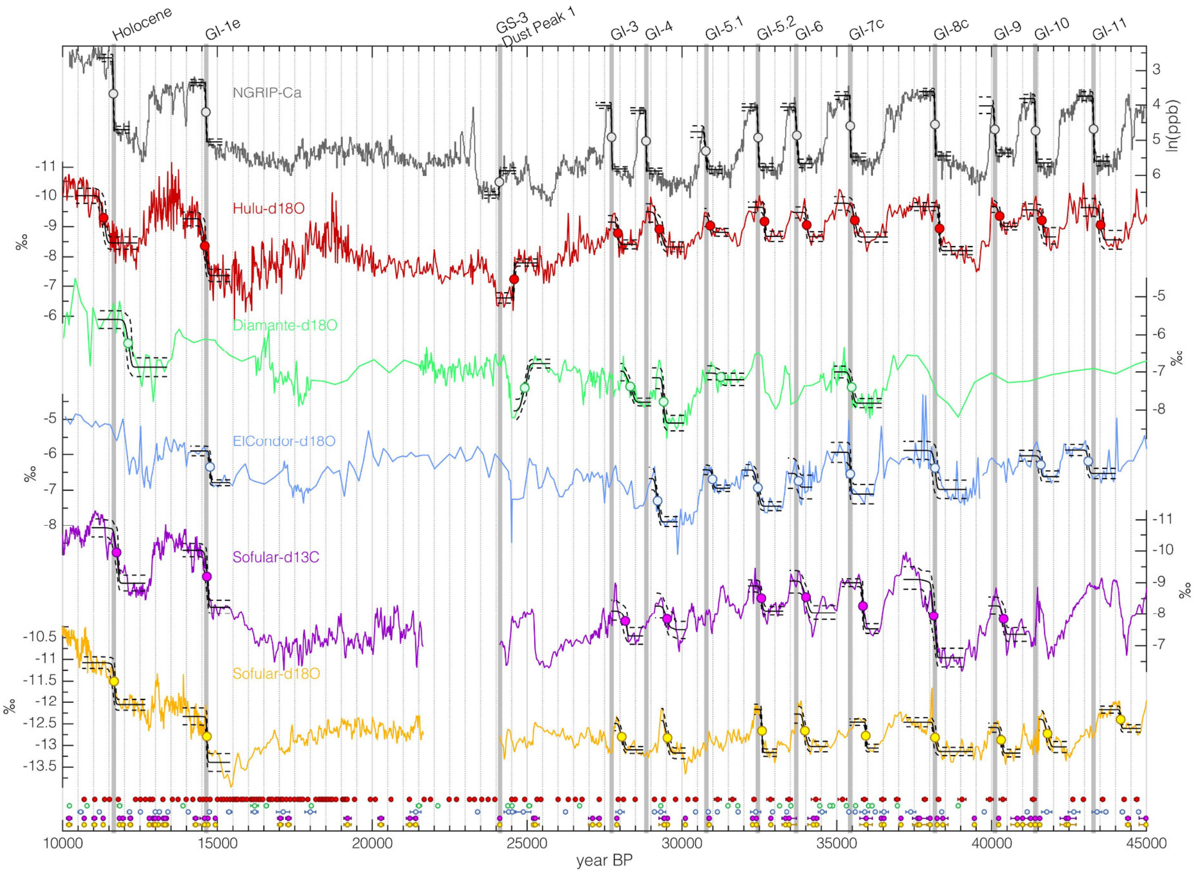

Figure 13: Timing of abrupt climate changes in different climate records. The climate archive and proxy is indicated in each panel. The black lines show the mean of the fitted ramps and their 95% confidence intervals (dashed). The dots mark the midpoint of the mean transition. The U/Th dates and their ±1σ uncertainties of each climate record are shown at the bottom of the figure in colour coding corresponding to the respective climate record. Each time series is shown on its original timescale not applying any synchronization.

Figure 13 shows the ice-core and speleothem climate records on their original individual timescales, along with the fitted ramps to the rapid climate changes. Note that we could not fit each climate event for every record, since the method requires a minimum number of data points defining the levels before and after each transition to produce reliable estimates. Already visually, a lag of climate changes in Greenland compared to the speleothem records can be consistently identified between 20 and 35 ka BP when all records are on their original timescales. Combining the PDFs of the detected change points in Greenland and the speleothems allows us to infer a probability estimate of the timing difference between climate events in Greenland and speleothems. These differences are shown in figure 14 along with our transfer function based on the matching of radionuclides from figure 12. This comparison shows that the differences in the timing of start-, mid- and end-point of DO-events in speleothems and ice cores largely fall within the uncertainties of our radionuclide-based timescale transfer function. Thus, rapid climate changes occur synchronously in Greenland and the (sub-) tropics. Notable exceptions are i) the transition from GS-1 to the Holocene around 11.6 ka BP, ii) Heinrich event 2 at 24 ka BP, and iii) DO-11 around 43 ka BP. However, there is large scatter among the different speleothem-based estimates at these events, indicating that these events are asynchronous in the different speleothems records on their respective timescales. Consequently, some of these records also imply asynchronous climate shifts with Greenland ice cores. This may either be interpreted as an indication of time-transgressive climate changes, or as a bias in individual speleothems – either in how climate is recorded in the speleothem, or their dating (for example through detrital thorium).

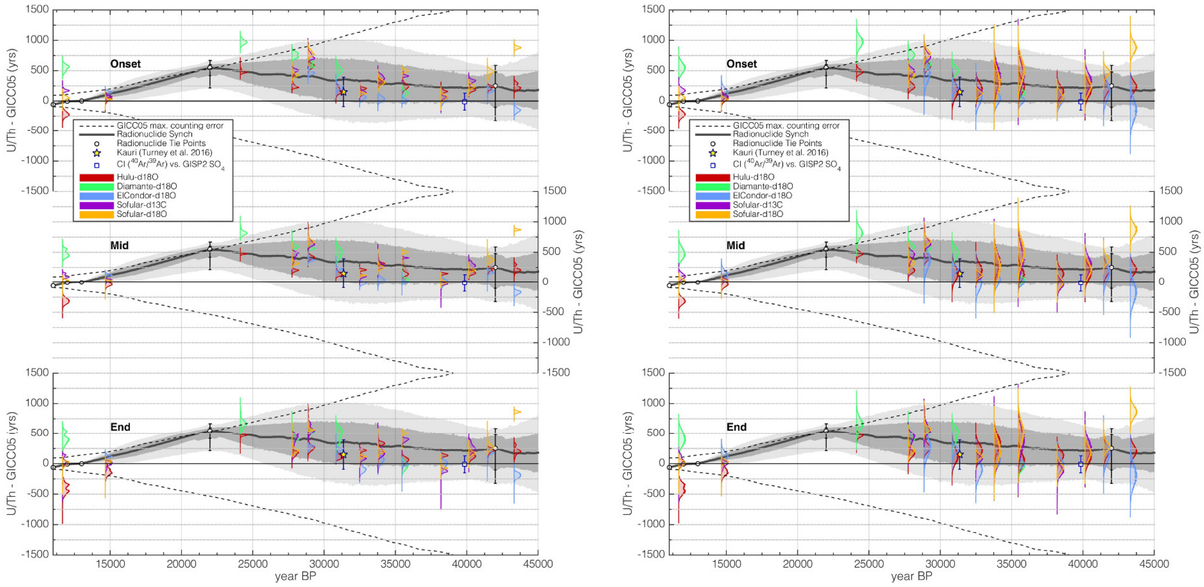

Figure 14: Timing differences of the onset (top), midpoint (middle) and end (bottom) of rapid climate changes in NGRIP and speleothems (coloured PDFs, see legend), and the timescale transfer function inferred from radionuclide matching (black line and grey shadings as in figure 12). The left panels show the PDFs of timing differences including only uncertainties from the determination of the change points in the climate records, while the right hand panels also include the speleothem dating uncertainties.

Averaging over all DO events, we can estimate an overall probability of leads and lags. By using the individual realizations of the radionuclide-based transfer function (see section 4.4) we take into account that the uncertainties of the transfer function are strongly autocorrelated. For each realization, we randomly sample the PDFs for the onset of the DO-events for the ice-core and speleothem records (see section 3.5), perturb the speleothem-based estimates within their U/Th dating errors, determine the lead or lag between the DO-onset in ice-core and speleothem records, and correct it for the expected lag from the realization of our transfer function. By averaging over all DO-events we thus obtain a mean lag for each realization and speleothem. In addition, we combine the different speleothem-based estimates of each realization by averaging over their mean lags to obtain an overall (speleothem & DO-event) mean lag. Converting the obtained lags from each realization into histograms we estimate the PDFs of average lags between ice-core and speleothem records.

Our lag estimates critically depend on our ability to fill the gaps between the widely spaced tie-points and thus, on our assumptions about the ice-core layer counting uncertainty, and how well our AR(1) process model can capture these (section 4). However, we note that by treating the mce as a highly correlated 1σ (instead of 2σ) uncertainty, our error estimate can be regarded as very conservative since it allows for large systematic drifts in each realization of the transfer function that will result in large errors of the mean.

The resulting PDFs of the lag between speleothems and ice cores are shown in figure 15. The uncertainties are mainly determined by our synchronization uncertainty. Thus, the uncertainty is only marginally reduced when averaging over all speleothems (Fig. 15, bottom): Because each realization of the transfer function varies smoothly, the offset between speleothem and ice-core records will be systematic for all speleothems in each realization, and is thus only marginally reduced by averaging.

We find that all speleothem records except Cueva del Diamante (Cheng et al., 2013b) indicate synchroneity with NGRIP within 1σ and that the delay obtained for Cueva del Diamante falls within 2σ. We note that the speleothem data from El Condor (Cheng et al., 2013b) from the same region as Cueva del Diamante does not indicate a significant lag to Greenland. Overall, our analysis cannot reject the null-hypothesis of synchronous DO-events in Greenland ice cores and (sub-) tropical speleothems (lag: $\mu \pm 1\sigma = 29 \pm 189$ years).

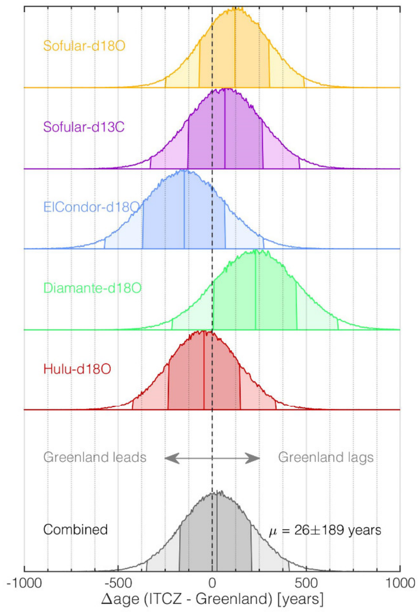

Figure 15: Average lead/lag between the onset of DO-events in the speleothems and NGRIP. Each panel (colour) shows the PDF for the lead/lag of the onset in the speleothem compared to NGRIP, averaged over all investigated DO-events (i.e., excluding the GS-3 Dust Peak/H2). The bottom most panel shows the PDF of the average of all DO-events and speleothems. The dark/light shading of the PDF in each panel indicates 68.2%/95.4% intervals.

6 Discussion

Our proposed transfer function quantifies the long-term differences between the Greenland ice-core and U/Th timescale and allows their synchronization. Even though based on only a few tie-points, this can be used to evaluate the absolute dating accuracy of Greenland ice-core records during the past 45 ka BP, while maintaining the strength of their precise relative dating. In combination with similar work done for the Holocene (Adolphi and Muscheler, 2016; Muscheler et al., 2014a), the picture emerges that the GICC05 counting error may be systematic: when accumulation and data resolution is high (e.g. in parts of the Holocene), too many annual layers have been counted, whereas during periods of low accumulation (e.g. GS-1 and GS-2) there is a tendency to identify too few annual layers. In principle, this is well captured by the GICC05 uncertainty estimate as the derivative of our transfer function is (within error) consistent with the increase of the counting error. However, our results caution against the use of the GICC05 counting error as a 2σ uncertainty as is often done in the literature. Originally, Andersen et al. (2006) pointed out that the MCE is not a true σ uncertainty but proposed that a Gaussian distribution with 2σ = MCE could serve as a pragmatic approximation. In combination with results from the Holocene (Adolphi and Muscheler, 2016) our study implies that the counting error can be strongly correlated over extended periods of time. This is in line with the discussion in Rasmussen et al. (2006) who point out that the main contribution to a potential bias in the layer count is the definition of how an annual layer is manifested in the proxy data. The data resolution as well as the manifestation of annual layers change between different climate states (Rasmussen et al., 2006), likely due to changes in aerosol transport and deposition resulting from variations in the atmospheric circulation and seasonality of precipitation (Merz et al., 2013; Werner et al., 2001). According to our analysis, the largest relative (i.e., year/year) change in the difference between GICC05 and the U/Th and tree-ring timescale occurs over GS-1 (11,653-12,846 years BP) and GS-2 (14,652-23,290 years BP). Both of these periods have likely been characterized by an increased relative contribution of summer precipitation to the annual ice layer (Werner et al., 2000; Denton et al., 2005), and the annual layers in the ice core have been identified in a similar way in both intervals (Rasmussen et al., 2006). In the 11-13 ka BP interval, the offset between GICC05 and the tree-ring timescale changes from -60 (95.4%-range: -77 to -42) years to zero (95.4%-range: -12 to +21) years. During the same interval, the GICC05 maximum counting error grows by 46 years. Albeit consistent within the absolute error margins, this stretch of GICC05 over GS-1 thus slightly exceeds the range allowed by the GICC05 counting error. Muscheler et al. (2014a) discussed that this stretch may be partly explained by errors in the placement of the oldest part of the tree-ring chronologies. However, here, we use a revised late glacial tree-ring dataset in which the different chronologies are connected much more robustly (Hogg et al., 2016). Furthermore, our analysis on the fully independent Hulu Cave ^{14}C data yields similar results (Fig. 7). Hence, our analyses clearly show that the GS-1 interval is about 60 years too short in the GICC05-timescale.

Between 15 and 22 ka BP, our analysis yields a change in the GICC05 offset from +118 (95.4%-range: 2-220) years to +549 (95.4%-range: 207-670) years, while the GICC05 counting error grows by 335 years. Thus, again, our transfer function changes a little faster than the maximum counting error allows during this interval. We note that our ^{14}C-^{10}Be matchpoint around 22,000 years BP has a relatively low signal-to-noise ratio in the ^{14}C data (see Fig. 8-9) and should, thus, be regarded as tentative. However, as shown in figure 8 our results are generally robust against different choices of subsets of the ^{14}C data and time windows. Nevertheless, it can also be seen that the estimates of the most likely age difference (i.e., the peak of the PDFs) differ slightly for different choices of the ^{14}C data. Hulu Cave yields a most likely offset of ~325 years, while Suigetsu implies a bigger age difference of ~550 years that coincides with a secondary probability peak in the Hulu Cave PDF. We note that assuming increased amounts of old soil organic carbon contributing to the speleothem formation would lead to an even stronger difference between these estimates (see section 3.4). Hence, we propose an age difference of +549 (95.4% range: 207-670) years based on the combination of all data (Fig. 9) that is consistent within error with the estimates based on the single datasets shown in figure 8, but stress that this tie-point should be re-evaluated as new suitable ^{14}C data becomes available in the future.

Assuming that the U/Th dates are absolute, our transfer function can be used to account for the bias in the GICC05 timescale and thus facilitate comparisons of ice-core records to other absolutely dated archives. However, we note that our synchronization does not necessarily lead to consistent timescales with radiocarbon-dated records. As discussed in section 3.3.2 (Fig. 4) and section 4.3 (Fig. 10 & 11), discrepancies of the datasets underlying IntCal13 can lead to erroneous structures in the calibration curve. The reduced amplitude of the $\Delta^{14}C$ change around the Laschamp geomagnetic field minimum in IntCal13 compared to its underlying data implies that IntCal $\Delta^{14}C$ must be offset prior to and/or after the Laschamp event. This underlines the challenges in radiocarbon calibration around this time pointed out by Muscheler et al. (2014b). Also more recently, Giaccio et al. (2017) pointed out that paired $^{40}Ar/^{39}Ar$ and ^{14}C-dating of the Campanian Ignimbrite around 40 ka BP yields inconsistent ages when the ^{14}C age is calibrated with IntCal13. Since IntCal13 in principle should be tied to the

U/Th-age scale for sections older than 13.9 ka BP, this implies either an inconsistency between Ar/Ar and U/Th dating or in the reconstructed ^{14}C levels of the calibration curve. The latter would be congruent with the conclusions by Muscheler et al. (2014b). If the problem was indeed the IntCal ^{14}C reconstruction, a synchronization of ice-core ^{10}Be to IntCal ^{14}C would not resolve this bias, since the problem would not be one of chronology, but of ^{14}C measurement and/or archive.
Our analysis provides the first rigorous test of whether DO-events recorded in speleothems and ice cores occur synchronously. We reject the hypothesis of leads and lags larger than 189 years at the one sigma level, consistent with the findings of Baumgartner et al. (2014). Since we compare to speleothem records from different regions, this also highlights that the ITCZ likely migrated synchronously (within uncertainties) over the different ocean basins and continents during the onset of DO-events (Schneider et al., 2014). However, there are also differences between the different speleothem records, which could be due to limitations in their dating or related to how directly individual archives record the rapid climate changes. The most notable examples are the onset of the Holocene and GI-11, which appear to occur asynchronously in the different speleothems (see Fig. 13 & 14). Another example is the younger GS-3 dust peak in the Greenland ice cores that appears to coincide with the East Asian Summer Monsoon decline seen in Hulu Cave, but postdates the precipitation increase seen in El Condor and Diamante. This change in the speleothems is typically attributed to the southward shift of the ITCZ as a response to Heinrich Event 2.

Figure 16 shows the period around H2. Firstly, we note that the younger of the two GS-3 dust peaks in the Greenland ice cores (Rasmussen et al., 2014a) occurs coevally (within chronological uncertainty) with the ITCZ movement recorded by the speleothems. At this time, the East Asian Summer Monsoon is strongly reduced as implied by decreased Hulu Cave $\delta^{18}O$ (Cheng et al., 2016). Coevally, precipitation increases in the South American Summer Monsoon region (Novello et al., 2017; Stríkis et al., 2018). The records thus exhibit more pronounced stadial conditions than normally seen during (non-Heinrich) DO-events. However, taken at face value, the precipitation increase at El Condor and Cueva del Diamante, the two northernmost sites shown in figure 16 (Cheng et al., 2013b), significantly predates the event seen in Greenland and Hulu Cave. It also predates the more southern South American sites Lapa Sem Fim (Stríkis et al., 2018) and Jaragua (Novello et al., 2017) by more than 500 years. This could either point to errors in the dating of the El Condor and Diamante speleothems, or be related to their latitudinal position. A freshwater-only experiment (all other boundary conditions held constant at 19 ka BP levels) with the Community Climate System Model 3 (TraCE-MWF, He, 2011) shows that, during a weak AMOC state, reduced advection of moisture from the tropical Atlantic leads to lower precipitation north of the ITCZ, while the ITCZ position over South America itself changes very little (Fig. 16). El Condor and Cueva del Diamante are both located very close to the LGM position of the ITCZ. It is hence possible, that when northern hemisphere summer insolation reached its lowest values over the past 50 kaBP around H2, the ITCZ migrated to a position south of El Condor and Cueva del Diamante, and during its transition caused the reconstructed precipitation change. As a result, the precipitation response to freshwater forcing would change sign at these cave sites. The sites located slightly further south only show a weak (Pacupahuain) or no (Paixao) response during this period, but are both characterized by increased variability. The two southernmost sites on the other hand (Jaragua and Lapa Sem Fim) remain south of the ITCZ throughout, and hence, show a clear increase in precipitation coeval with the signal in Greenland and Hulu Cave. In this context, the precipitation increase in El Condor and Cueva del Diamante around 25kaBP (i.e., prior to H2) may signify when the ITCZ transitions over the sites. The subsequent reduction in AMOC strength during H2 then leads to a decrease in precipitation in north-west South America, an increase further south, and little change in between. Tentative support for this can be drawn from the response of the El Condor and Cueva del Diamante speleothems to GI-2.2 and GI-2.1 where, albeit weakly, the $\delta^{18}O$ records imply an increase in precipitation during GI-2 which is opposite to their response to DO-events during MIS-3 (Fig. 13, 16). Thus, this analysis indicates, that the seemingly asynchronous response to climate change in different proxy records may indeed only reflect site specific changes in the proxy response. Alternatively, we cannot rule out undetected issues with the U/Th ages of these speleothems. A detailed analysis of this observation feature is beyond the scope of this paper, but in the context of a timescale perspective, which is the focus of this work, it highlights the caveats of climate wiggle-matching between single records, even if the mechanistic link between regional climate changes may be relatively well understood.

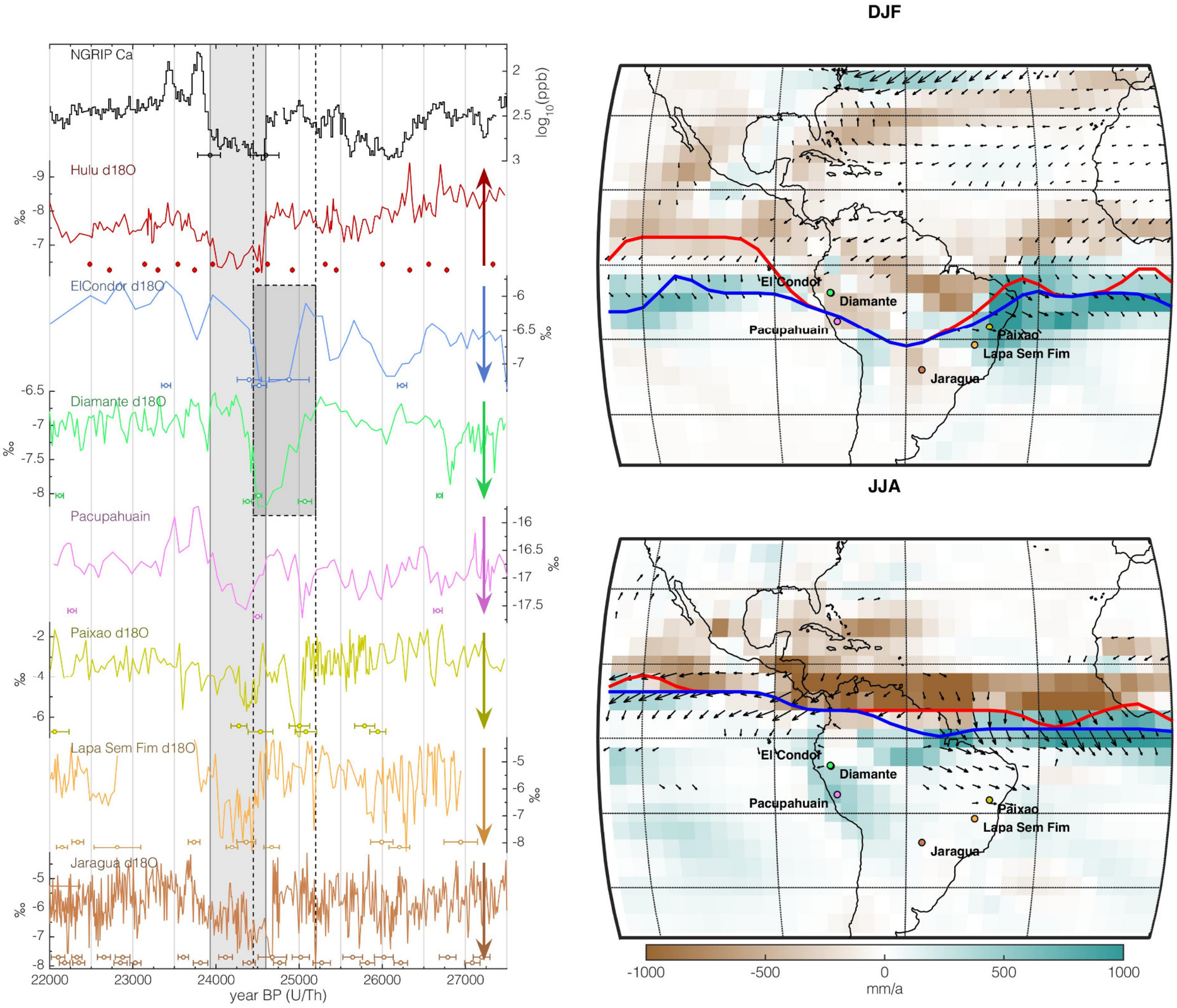

Figure 16: Climate changes around H2. Left (from top to bottom): NGRIP Ca (Bigler, 2004) on the synchronized timescale (Fig. 14), Hulu Cave $\delta^{18}O$ (Cheng et al., 2016), El Condor $\delta^{18}O$ (Cheng et al., 2013b), Cueva del Diamante $\delta^{18}O$ (Cheng et al., 2013b), Pacupahuain $\delta^{18}O$ (Kanner et al., 2012), Paixao $\delta^{18}O$ (Stríkis et al., 2018), Lapa Sem Fim $\delta^{18}O$ (Stríkis et al., 2018), Jaragua Cave $\delta^{18}O$ (Novello et al., 2017). The arrows on the right hand side of each axis point in the direction of the signature of increased precipitation on $\delta^{18}O$ through the amount effect (Dansgaard, 1964). The light grey box marks H2. The dark grey box highlights the preceding $\delta^{18}O$ anomaly in El Condor and Diamante caves. Right: Precipitation (colour) and wind (arrows) response to freshwater forcing in the CCSM3 climate model (freshwater only experiment of TraCE21k, all other forcings are held at 19k conditions, He, 2011). The red (blue) line depicts the latitude of the precipitation maximum during strong (weak) AMOC-states. Only wind anomalies >1m/s are plotted. The cave sites are indicated as dots. The top panel shows the winter (December-

February) response, while the bottom panel shows the summer (June-August) response. Anomalies are plotted as weak-strong AMOC mode.

7 Conclusion

We present the first radionuclide-based comparison between the Greenland Ice Core Chronology 2005 (GICC05) and the U/Th timescale. We find that GICC05 is accurate within its stated absolute uncertainties, but also that the maximum counting error of the GICC05 may be at the limit to capture the total uncertainty accumulated within certain climatic periods. Our analysis indicates that the relationship between GICC05 and the U/Th timescale over the last 45 ka drifts over time and reaches its maximum offset around 22 ka BP. We propose a transfer function that quantifies this drift and facilitates analysis of ice-core and U/Th records, such as speleothems, on a common time scale. Thus, this transfer function allows further integration of key-timescales in paleosciences and contributes to the INTIMATE (INTegration of Ice-core, MArine, and TEerrestrial records) initiative (Bjorck et al., 1996; Rasmussen et al., 2014b; Bronk Ramsey et al., 2014). Provided that U/Th ages are regarded accurate, the transfer function strongly reduces the absolute dating uncertainty of Greenland ice cores back to 45 ka BP. We reject the hypothesis if leads or lags larger than 189 years between Greenland, East Asia, and South America at the one sigma level. We show that the southward ITCZ shift around 24.5 ka BP seen in speleothems, typically associated with H2, coincides with the younger GS-3 dust peak recorded in Greenland ice cores. However, we also highlight inconsistencies between speleothem records around the onset of the Holocene, late GS-3, and GI-11 and thus, caveats to the commonly applied practice of climate wiggle-matching. By comparing various ^{14}C records underlying IntCal13 as well as ice-core ^{10}Be data and geomagnetic field records, we infer that the current radiocarbon calibration curve underestimates the amplitude and rapidity of the $\Delta^{14}C$ change around the Laschamp event 41 ka BP. This adds to previous studies (Giaccio et al., 2017; Muscheler et al., 2014b) highlighting that there are likely systematic errors in IntCal13 that will directly translate into errors of radiocarbon-based chronologies around that time. The combination of several internally inconsistent datasets in IntCal13 can lead to erroneous timing and amplitude of $\Delta^{14}C$ changes. Hence, great care has to be taken when attempting to use sections older than 30 ka BP of IntCal13 directly for studies of ^{14}C production rates and/or carbon cycle changes.

8 Data Availability

The transfer function shown in figure 12 will be made available as a supplementary to this paper and on NOAA.

9 Author contributions

FA designed and carried out analyses, and wrote the manuscript in correspondence with CBR and RM. TE designed and applied break-point detection analysis and wrote the corresponding methods section. FA, RM, CBR, SOR, CT and AC initiated the project. RLE and HC provided speleothem data. AS and SOR provided insights into the ice core chronology. HF and TE gave insights into aerosol transport and deposition. All authors discussed and commented on the manuscript.

10 Competing interests

The authors declare that they have no conflict of interest.

11 Acknowledgements

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