Synchronizing ice-core and U/Th time scales in the Last Glacial Maximum using Hulu Cave ^{14}C and new ^{10}Be measurements from Greenland and Antarctica

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Abstract

Between 15 and 27 ka b2k (thousands of years before 2000 CE) during the last glacial, Greenland experienced a prolonged cold stadial phase, interrupted by two short-lived warm interstadials. Greenland ice-core calcium data show two periods, preceding the interstadials, of anomalously high atmospheric dust loading, the origin of which is not well understood. At approximately the same time as the Greenland dust peaks, the Chinese Hulu Cave speleothems exhibit a climatic signal suggested to be a response to Heinrich Event 2, a period of enhanced ice-rafted debris deposition in the North Atlantic. In the climatic signal of Antarctic ice cores, moreover, a relative warming occurs between 23 and 24.5 ka b2k that is generally interpreted as a counterpart to a cool climate phase in the Northern Hemisphere. Proposed centennial-scale offsets between the polar ice-core time scales and the speleothem time scale hamper the precise reconstruction of the global sequence of these climatic events. Here, we examine two new ^{10}Be datasets from Greenland (NorthGRIP) and Antarctic (WDC) ice cores to test the agreement between different time scales, by taking advantage of the globally synchronous cosmogenic radionuclide production rates.

Evidence of an event similar to the Maunder Solar Minimum is found in the new ^{10}Be datasets, supported by lower-resolution radionuclide data from Greenland and ^{14}C in the Hulu Cave speleothem, representing a good synchronization candidate at around 22 ka b2k. By matching the respective ^{10}Be data, we determine the offset between the Greenland ice-core time scale, GICC05, and the WDC Antarctic time scale, WD2014, to be 125±40 years. Furthermore, via radionuclide wiggle-matching, we determine the offset between the Hulu speleothem and ice core ~~timescales~~time scales to be 375 years for GICC05 (75–625 years at 68% confidence), and 225 years for WD2014 (-25–425 years at 68% confidence). The rather wide uncertainties are intrinsic to the wiggle-matching algorithm and the limitations set by data resolution. The undercounting of annual layers in GICC05 inferred from the offset is hypothesized to have been caused by a combination of underdetected annual layers, especially during periods with low winter precipitation, and misinterpreted unusual patterns in the annual signal, during the extremely cold period often referred to as Heinrich Stadial 1.

1 Introduction

~~Paleoclimatic studies rely on climate records with accurate timescales to allow identification of the driving~~ Well-dated paleoclimatic archives, such as ice cores or speleothems, are essential for reconstructing the mechanisms of past climate change. ~~At present, numerous independent~~Independent chronologies~~, based on vastly different methods, are used to date specific~~ for each climatic ~~archives~~archive are necessary for studying the recorded signals and are determined via, for example, ~~layer~~ counting~~, measuring the~~ annual layers or based on U/Th ~~decay, or radiocarbon dating. The~~ measurements. However, the uncertainties ~~and inaccuracies~~ of each~~ of those~~ time ~~scales are often difficult to assess and are major obstacles to an accurate and complete~~ scale and a lack of matching horizons hamper the global ~~reconstruction of paleoclimate.~~comparison of climatic proxy records.

~~The~~During the Last Glacial Maximum (LGM) ~~was the last period when~~ the ice sheets were at their largest extent ~~before the deglaciation into the Holocene, although its exact stratigraphic definition is being debated (Hughes et al., 2013). For terrestrial records of the Northern Hemisphere,~~(defined as ca. 23.3-27.5 ka by Hughes & Gibbard~~(~~, 2015~~) suggested defining the LGM as the cold period~~ , although other definitions exist) and extreme climatic shifts were recorded in the Greenland ~~Stadial 3 (GS-3: 23.3-27.5 ka b2k following~~ and Antarctic ice cores, as well as in Asian speleothems such as the ones from Hulu Cave, China (Wang et al., 2001). To advance our knowledge of the LGM as recorded in the archives, we need to estimate the leads and lags between these shifts, but a climatic synchronization is not always possible and not free of interpretation caveats. A lack of secure tie points over the LGM by most synchronization studies (Svensson et al., 2020; Corrick et al., 2020; WAIS project members, 2015) makes the reconstruction of a global sequence of events difficult to assess.

In Greenland ice cores, the water stable isotope data (e.g. $\delta^{18}O_{ice}$) records two unusually short and small Greenland interstadials, GI-2.1 and 2.2 (Rasmussen et al., 2014), ~~based on the analysis of NorthGRIP dust and sea-level records. In other ice-extent reconstructions, the LGM was established to have lasted until mid~~briefly interrupting the long cold period formed by the two Greenland stadials GS-2 and GS-~~2, around 20 ka b2k (Clark et al., 2009). In the present study, we refer~~3 (Fig. 1a). Despite other periods of interstadial climate conditions being recorded widely across the Northern hemisphere, there is no counterpart of the brief GI-2.1 and 2.2 in the Hulu $\delta^{18}O_{calcite}$ (Fig. 1b), which hampers the comparison of Greenland ice cores and Asian speleothems.

Around the same time, the $\delta^{18}O_{ice}$ in several Antarctic ice cores reaches a maximum (AIM-2; EPICA Community Members, 2006), interrupting a warming trend (Fig. 1c). Greenland and Antarctic $\delta^{18}O_{ice}$ records are hypothesized to be coupled by the bipolar seesaw mechanism (Stocker & Johnsen, 2003), by which a Greenlandic transition to an interstadial will initiate cooling in Antarctica with a small delay. According to the current chronologies, however, the onset of the Antarctic cooling, i.e. the peak of the AIM-2, leads the onset of the GI-2.2 by about 260 years. The average delay, or signal transmission time, from the North Atlantic to the ~~period 20-25 ka b2k as the LGM for simplicity, since it coincides~~Antarctica is estimated to be 1 to 2 centuries, as shown by ice core measurements (Svensson et al., 2020; WAIS project members, 2015) and in line with ~~the age limits of our new Greenland ^{10}Be dataset, being aware that this is not a formal stratigraphic definition.~~modelling experiments (Pedro et al., 2018), but the lead-lag dynamic appears to be reversed for AIM-2.

During ~~this time, a phase of~~GS-2 and GS-3, at least two massive discharge events of icebergs from the Laurentide ice sheet ~~was~~were inferred from the ice-rafted debris content of North Atlantic marine sediments, defining the occurrence of the Heinrich ~~Event~~Events 1 and 2 (HE-1 and HE-2; Bard et al., 2000; Peck et al., 2006~~). HEs~~), which are each now thought to consist of 2 pulses. Heinrich events occurred during ~~some GSs~~Greenland stadials and added a large amount of freshwater to the surface ocean, likely causing ~~a more~~an extreme shutdown of the Atlantic overturning meridional circulation (AMOC) and, thereby, prolonged climatic conditions of ~~extreme~~severe cold (McManus et al., 2004). The term Heinrich ~~Stadial~~stadial (HS) is often used to indicate the period affected by the ~~HE. The duration of HS-1, for example, is limited to the 14.5-17.5 ka b2k interval within GS-2.1 (Broecker and Barker, 2007), while for HS-2, a correspondence with the late GS-3 is often argued for, based on speleothem water isotope records (e.g. Li et al., 2021).~~Heinrich events.

~~Associated with~~ Signatures of climatic change likely related to HS-1 and HS-2~~, one could expect a cooling signal~~ are recorded in the ~~stable water isotopes of~~ $\delta^{18}O_{calcite}$ of the Hulu speleothems (Fig. 1b; Cheng et al., 2018), but no counterpart is found in the Greenland ~~ice cores (~~$\delta^{18}O_{ice}$~~; Jouzel et al., 1997) but this is not the case.~~. Possible reasons include that the high-latitude areas around the ice sheet were not strongly affected by the cooling or that the isotopes reacted non-linearly (Guillevic et al., 2014). ~~For example, it~~It has also been suggested that, during HS-1, ~~the empirical correlation of temperature and~~ ~~$\delta^{18}O_{ice}$ was disrupted by~~ changes in the precipitation ~~pattern alterations~~ seasonality influenced the annual mean $\delta^{18}O_{ice}$ in a way that obscured the relation of $\delta^{18}O_{ice}$ and annual average temperatures (He et al., 2021~~), which~~). This may also ~~be the case for~~ apply to HS-2.

~~Mineral~~While $\delta^{18}O_{ice}$ does not show a clear imprint of the Heinrich events, Greenland ice-core calcium data show two periods of heightened concentration during GS-3 (Fig. 1a; Rasmussen et al., 2014) that could be related to atmospheric reorganization around HE-2 (Adolphi et al. 2018), although there is debate about this attribution (Hughes & Gibbard, 2015). Since mineral dust aerosols in Greenland ~~ice cores (as reflected by e.g.~~ ~~Ca^{2+} measurements) have their major source on the~~ originate mostly from the Eurasian continent ~~and mainly reflect the dryness of the source region and wind strength, which are highest in periods of extreme cold~~ (Schüpbach et al., ~~2018). In addition to the already high stadial calcium levels, during GS-3 calcium deposition in Greenland shows two periods of further increased concentration (Rasmussen et al., 2014), which could be related to atmospheric reorganization following HE-2, although this attribution is still speculative (Hughes & Gibbard, 2015). Also, methane~~ ~~(CH_4) emissions originating from tropical wetlands and trapped in the ice cores may have strengthened during HSs and could be used as another indicator of the HS-2 climate (Rhodes et al., 2015), since higher methane levels are found in Antarctic ice cores around the same time.~~2018) and the dust record therefore should reflect climate variability in Asia, Dong et al. (2022) synchronized the youngest Greenland calcium peak to the HS-2 signal of an Asian speleothem (Cherrapunji, India). Based on the assumption of synchronous climate signals, they inferred a multi-centennial dating offset between the GICC05 ice-core time scale and the U/Th-dated speleothem record, that is consistent with earlier estimates based on cosmogenic radionuclides (Adolphi et al. 2018).

~~The objective~~In addition to the Antarctic warming indicated by AIM-2, an abrupt rise of ~~this study is~~ methane levels (Fig. 1c) indicates a ~~comparison~~likely response of ~~three~~the tropical methane emission sources to a period of climate change (Rhodes et al., 2015), which is suggested to correspond to HS-2. The phasing of the methane signal and the Heinrich stadial 2 is still unresolved because of possible offsets in the respective chronologies.

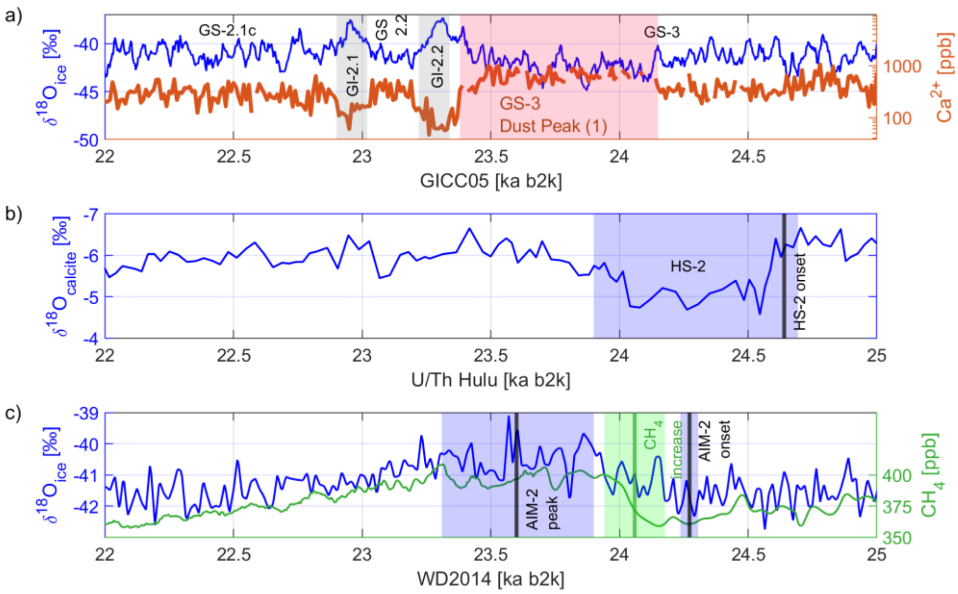


Figure 1 Climate signals on their original time scales~~ in the LGM~~. The time ~~scales we~~direction is from right to left. The references to all datasets are ~~going to examine are the speleothem~~ summarized in Table 1. Notice the relative position of onsets and terminations of the climate shifts. a) Greenland stable oxygen isotopes (GRIP) and calcium (log scale, NorthGRIP), showing the alternating signals of interstadials (GI) and stadials (GS), as defined by Rasmussen et al. (2014), and the dust peak during GS-3. b) Hulu Cave stable oxygen isotopes, on a reversed y-axis, showing the cold phase attributed to HS-2. c) Antarctic (WD ice core) stable oxygen isotopes showing the Antarctic Isotope Maximum 2 (AIM-2; EPICA community members, 2006), and the methane record showing a rapid increase.

The unclear causal relationships between the mentioned signals challenge the interpretation of the climate dynamics of the LGM. The time scale ~~from Hulu Cave, China (Wang et al. 2001; Southon et al., 2012; Cheng et al., 2016), the time scale~~uncertainties of each archive, in the order of centuries for the ice-core chronologies (Svensson et al., 2008; Sigl et al., 2016), mask any absolute inference of the climatic leads and lags. Non-climatic synchronization horizons such as volcanic eruptions or solar events are needed to obtain solid estimates of the relative time scale offsets. For example, Svensson et al. ~~Antarctic WDC ice core (WD2014; Sigl et al., 2016), and the~~ (2020) compared the two layer-counted polar chronologies GICC05 (Greenland ice core chronology ~~(GICC05; Andersen et al., 2006; Svensson et al., 2008).~~; Svensson et al., 2008) and WD2014 (WAIS Divide chronology; Sigl et al., 2016) via bipolar volcanic tie points. They found a negligible offset at 16 ka and only 80 years at 24.6 ka. However, they lack tie points within this entire interval.

~~The Hulu Cave speleothems were dated by U/Th measurements for the period from 15 to 55 ka b2k and have previously been analysed for carbon isotopes (^{14}C; Southon et al., 2012; Cheng et al., 2018) and climatic proxies, such as stable oxygen isotopes ($\delta^{18}O_{calcite}$; Wang et al. 2001; Cheng et al., 2016). The dead carbon fraction (DCF) of ^{14}C in speleothems from this cave, that is, the estimated amount of older carbon that may contaminate the recorded atmospheric ^{14}C signal, is very low and assumed to be constant over time (Cheng et al., 2018) making these speleothems excellent candidates for calibration of the IntCal20 curve (Reimer et al., 2020).~~

~~Throughout the 15 to 42 ka b2k period, the GICC05 was constructed by manually counting annual layers in the electrical conductivity (ECM), continuous flow analysis (CFA) ion records, and visual stratigraphy of the NorthGRIP ice core (Andersen et al., 2006; Svensson et al., 2006). The uncertainty of GICC05 was assessed by identifying so-called 'uncertain annual layers', each contributing to the chronology and to the Maximum Counting Error (MCE) as 0.5±0.5 years (Rasmussen et al., 2006). The2014).~~

~~For Antarctica, the WD2014 time scale was constructed by manual and automatic counting of layers in the WDC ice core until 31.2 ka b2k (Sigl et al., 2016). The data used for counting were the ECM and CFA impurity records; however, for the period 15-27 ka b2k, only ECM data were suitable for layer counting because of insufficient resolution of the CFA record. The uncertainty was assessed by comparison to the previous time scale (WAIS Divide Project Members, 2013), as well as by comparing the manual and automated versions of the layer count (Winstrup et al., 2012).~~

~~Around 23 ka b2k, the $\delta^{18}O_{ice}$ of Greenland indicates two short-lived Greenland interstadials, also known as Dansgaard-Oeschger events, namely GI-2.1 and 2.2 (Rasmussen et al., 2014). Antarctica experienced warming between 24.5 and 24 ka b2k, as shown by the Antarctic Isotope Maximum 2 (AIM-2); subsequently, the water isotope levels remained high until ~23 ka b2k, after which a cooling trend lasted until around 22.2 ka b2k (EPICA Community Members, 2006). Greenlandic and~~

~~Antarctic $\delta^{18}O_{ice}$ records are hypothesized to be coupled by the bipolar seesaw mechanism (Stocker & Johnsen, 2003; Pedro et al., 2018), connecting GI-GS pairs and AIM stages.~~

~~By synchronizing Antarctic CH_4 and Greenlandic $\delta^{18}O_{ice}$, the WAIS project members (2015) stated that, on average, the onset of Antarctic cooling lags the onset of the GI warming by 218±92 years. The authors duly excluded the GI-2–AIM-2 pair from their lead-lag analysis, firstly because the GISP2 CH_4 record did not support synchronicity with the GI-2 temperature increase, and, secondly, because the older HE-4 and HE-5 were similarly associated with higher CH_4 levels. Recently, Svensson et al. (2020) presented a bipolar volcanic match between Greenland and Antarctic ice cores, allowing re-synchronization of the GI-AIM pairs. They find an average delay of 122±24 years but the pair GI-2–AIM-2 is also excluded from their analysis.~~

~~According to the current chronologies, the onset of Antarctic cooling, i.e. the peak of the AIM-2 event, can be visually identified in the $\delta^{18}O_{ice}$ of WDC around 23.6 ka b2k (Jones et al., 2017), leading the corresponding onset of GI-2.2 by about 260 years (Rasmussen et al., 2014; Sigl et al., 2016). Therefore, the pattern of the bipolar seesaw appears different for AIM-2, both because the lead-lag dynamic with Greenland appears to be reversed and because the AIM-2 phase has a disproportionate duration compared to the very short GI-2.1 and GI-2.2 interstadials. Resolving some time-scale issues, which we will delineate shortly, will clarify the distinctive timing factors of the global climate around HE-2, compared to the 'conventional' bipolar seesaw scenario.~~

~~Traces of volcanic eruptions and cosmogenic radionuclides provide synchronization tools that do not rely on the precise identification of climatic match-points and on the assumption of their synchronicity. While aligning GIs across climatic archives provides a broad overview of climate in different regions, the assumed synchronicity of GIs prevents us from assessing the actual leads and lags within the climate system. Moreover, the climatic tie points may be difficult to identify or not available in all time periods. For example, while $\delta^{18}O_{calcite}$ of Asian speleothems show a clear signal at the time of HE-2, consistent with large-scale climate changes (Li et al., 2021), there is no counterpart of the brief GI-2.1 and 2.2, hampering synchronization to Greenlandic $\delta^{18}O_{ice}$.~~

~~Across the LGM, most studies lack bipolar and inter-regional tie points to allow for an accurate reconstruction of the sequence of events.(2020) do not report bipolar volcanic tie points over the entire period 16.5 to 24.7 ka b2k. At 24.7 ka b2k, they evaluate GICC05 to be ~85 years older than WD2014. Another inter-regional synchronization effort by Corrick et al. (2020) offers a climatic synchronization of speleothem and Greenland $\delta^{18}O$, but also lacks tie points over the LGM. Their estimation around GI-3 (27.78 ka b2k) is that GICC05 is 90 years younger than U/Th-dated samples.~~

~~A comparison between WDC CH_4, Hulu $\delta^{18}O$, and Greenlandic $\delta^{18}O$ (Sigl et al., 2016) found that, at the onset of GI-3, WD2014 is 167 years younger than the Hulu time scale and that GICC05 is younger than WD2014 by about 30 years and therefore 197 younger than Hulu. These two latter studies agree that around GI-3, events are younger according to the ice-core time scales than according to the U/Th time scales, albeit disagreeing about how much, while there is indication that the offset between GICC05 and WD2014 changes sign between 27.78 and 24.7 ka b2k.~~

~~Given the scarcity of climatic and volcanic tie points over the LGM, in this work, we focus on new measurements of cosmogenic radionuclides to directly compare polar ice cores and the Hulu time scale. Therefore, thecarbon-cycle models. The model we apply in this study(2014, 2016, 2018) and is based on the box-diffusion model by Siegenthaler (1983). The main assumption that ^{10}Be varies proportionally to the true global production rate of cosmogenic radionuclides may, however, lead to uncertainties. Similarly, ^{14}C may be affected by changes in the carbon cycle, adding additional signals that are not related to production rate changes.~~

~~Currently~~Traces of cosmogenic radionuclides (e.g., ^{14}C and ^{10}Be) allow for global synchronizations because of their common production history and widely recorded signals. So far, the only radionuclide-based tie point in the LGM between ice-core and Hulu Cave records was found by Adolphi et al. (2018). ~~An~~Around 22 ka, they measured an offset of 550 years (95% probability interval: 215-670 years) between GICC05 ~~(younger)~~ and the Hulu time scale ~~(older) was estimated at 22 ka b2k, shortly after~~, the ~~GS-2.1 onset. Here, we test and refine this result by using new higher-resolution ^{10}Be data from Greenland and adding new Antarctic ^{10}Be data to our comparison. We also investigate if and where the ice-core time scales may have accumulated high dating inaccuracies and reconstruct the timing of events across the LGM.~~latter being oldest.

This offset is unexpected considering the good agreement between the two chronologies measured both before and after the LGM; the climate synchronization between speleothems and GICC05 by Corrick et al. (2020) found small offsets of 4 years at GI-1 (15 ka) and of 92 years at GI-3 (27.5 ka). Nonetheless, the climatic synchronization by Dong et al. (2022) suggests that GICC05 is too young compared to the Cherrapunji speleothem U/Th ages and estimates the offset to 320 years (with a 2σ-uncertainty of 90 years) around 24.5 ka. In addition, the WD2014 and Hulu time scales are offset by 167 years around GI-3 (Sigl et al., 2016), suggesting that the Antarctic ice core chronology may also be diverging from U/Th ages.

In this study, to better determine the sequence of events during the LGM, we aim at establishing strong ties between the three chronologies: GICC05, WD2014, both dated by annual layer counting, and Hulu Cave, dated by U/Th measurements. Since the WD2014 chronology in the LGM was never compared against the Hulu and GICC05 chronologies, new Antarctic ^{10}Be measurements are needed to fill the gap. Further, new ^{10}Be measurements from Greenland are needed to strengthen the statistical significance of the 22-ka tie point by Adolphi et al. (2018), which presents a low signal-to-noise ratio. The Hulu cave $^{14}C_{calcite}$ and $\delta^{18}O_{calcite}$ measurements were updated in recent years (Cheng et al., 2018) and directly included in the radiocarbon calibration curve (IntCal20; Reimer et al., 2020), used to date radiocarbon from organic samples worldwide. We therefore aim to evaluate the tie point by Adolphi et al. (2018) using new measurements of ^{10}Be and the recent $^{14}C_{calcite}$. We will use a combination of carbon-cycle modelling and statistical wiggle-matching (Siegenthaler, 1983; Adolphi & Muscheler, 2016) to directly compare proxy records from both polar ice sheets and the Hulu cave.

In this work, we confirm the existence of centennial offsets in the LGM between the three chronologies and we position the mentioned global climatic shifts in relation to each other. The question of why chronological offsets quickly develop remains open, but we suggest that difficulties with annual-layer identification in the very cold parts of the last glacial are the likely source of the observed offsets.

2 Data and Methods

In this study, we aim at comparing the new cosmogenic radionuclide data with other datasets from Greenland, Hulu Cave, and Antarctica. We also ~~analysed~~analyze water stable isotope, methane, and calcium data to assess climatic changes. We summarize the relevant datasets, age resolutions, and citations in ~~table 1~~Table 1. In this work, we refer to years b2k because they are commonly used for the GICC05 time scale; when necessary, we have converted from years BP (before 1950 CE) by adding 50 years.

1 Table 1 Datasets used in this study. *Age resolutions represent the period 20 to 25 ka b2k. For the Hulu Cave, $^{14}C_{calcite}$ data from two speleothems from the cave, labelled H82 and MSD, were published in two separate studies.*

Dataset	Location	Proxy	Avg. Age Res., years	Reference
NorthGRIP	75°10’N 42°32’W	^{10}Be	10	This study
		Ca^{2+}	20	Erhardt et al., 2021
		$\delta^{18}O_{ice}$	10	NorthGRIP members, 2004
GRIP	72°58’N 37°64’W	^{10}Be	27	Yiou et al., 1997; Muscheler et al., 2004
		$\delta^{18}O_{ice}$	10	Johnsen et al., 1997
GISP2	72°36’N 38°30’W	^{10}Be	162	Finkel & Nishiizumi, 1997
		$\delta^{18}O_{ice}$	10	Stuiver & Grootes, 2000
WDC	79°46’S 112°08’W	^{10}Be	67	This study

Dataset	Location	Proxy	Avg. Age Res., years	Reference
		$\delta^{18}O_{ice}$	10	Jones et al., 2017
		Non-sea-salt-Sulfate (nssS)	1	Buizert et al., 2018
Hulu Cave	32°30'N 119°10'E	$^{14}C_{calcite}$	271	H82: Southon et al., 2012
			124	MSD: Cheng et al., 2018
		$\delta^{18}O_{calcite}$	70	H82: Wang et al., 2001; Wu et al. 2009
			31	MSD: Cheng et al., 2016

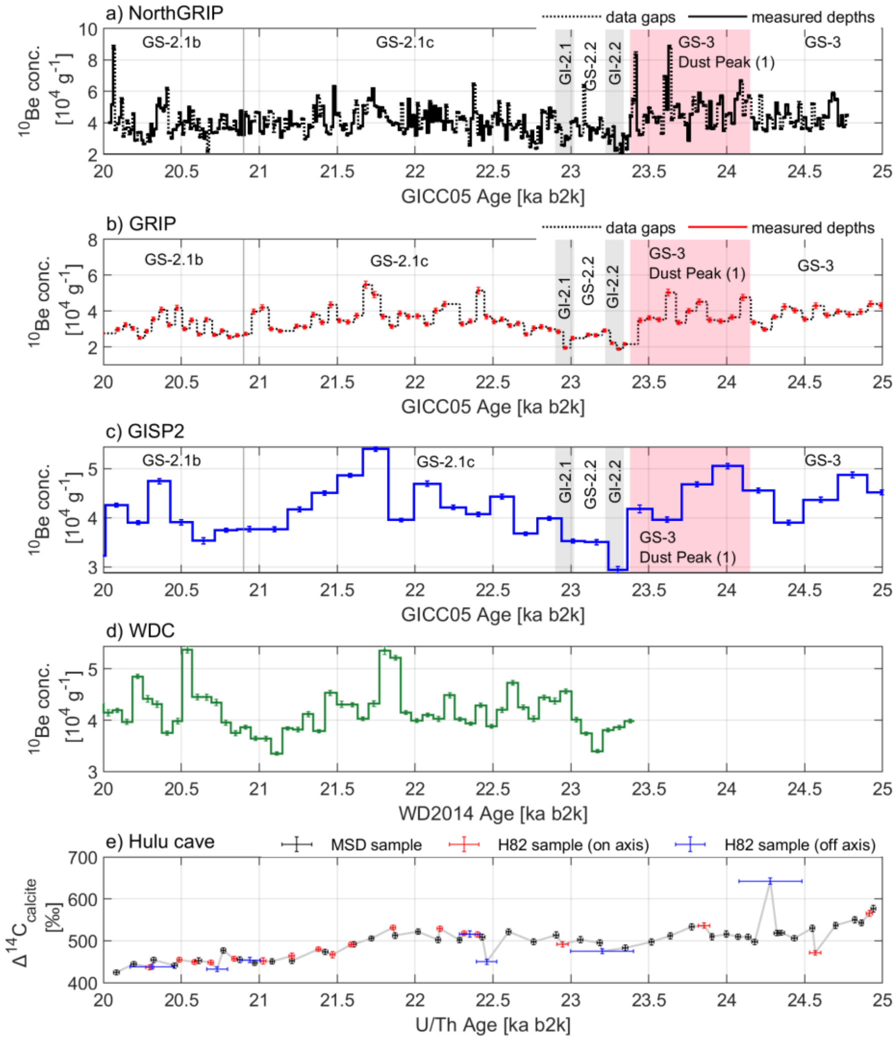


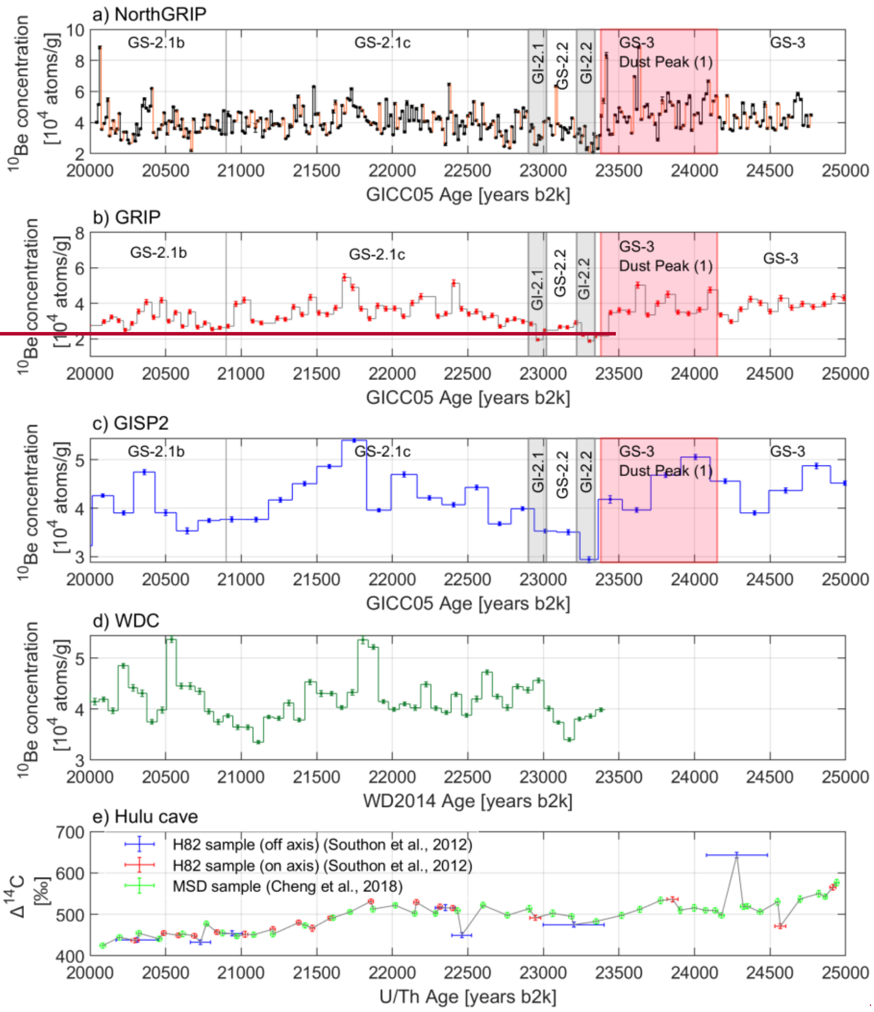
Figure 2 Cosmogenic radionuclide data used in this study (see Table 1 for references). On the GICC05 time scale: ^{10}Be concentrations in (a) NorthGRIP, (b) GRIP, and (c) GISP2. Dotted lines in the GRIP data indicate discontinuities between every 55-cm resolved sample. On the WD2014 time scale: (d) WDC ^{10}Be concentrations. On a U/Th-based time scale: (e) Hulu Cave $\Delta^{14}C_{calcite}$ data as reported in three separate datasets. In the following, we exclude the off-axis H82 measurements (blue), as they show more outliers and wider dating uncertainties.

2.1 Preparation and measurement of the NorthGRIP samples

The 322 new Greenland ^{10}Be measurements were performed at ETH Zurich on samples from the NorthGRIP ice core between 1726.45 m and 1816.51 m depth, which according to GICC05 correspond to ages between ~~20039~~20,039 to ~~24774~~24,774 years b2k (Andersen et al., 2006). The samples have a variable temporal resolution between 7.5 years and 14 years with some smaller gaps (see Methods Appendix, sec. 8.2) and are the so-far best-resolved available radionuclide dataset for the LGM. The extraction of ^{10}Be followed protocols by Nguyen et al. (2021), see Methods Appendix, sec. 8.3. The measured ^{10}Be concentrations are shown in ~~fig. 1a~~Fig. 2a.

~~*Age resolutions are calculated for the period 20 to 25 ka b2k. For the Hulu Cave, ^{14}C data from two speleothems from the cave, labelled H82 and MSD, were published in two separate studies.*~~

~~Dataset~~	~~Location~~	~~Proxy~~	~~Avg. Age Res. [years]~~	~~Reference~~
~~NorthGRIP~~	~~75.10N 42.32W~~	~~^{10}Be~~	~~10~~	~~This study~~
		~~Ca^{2+}~~	~~20~~	~~Erhardt et al., 2021~~
		~~$\delta^{18}O$~~	~~10~~	~~NorthGRIP members, 2004~~
~~GRIP~~	~~72.58N 37.64W~~	~~^{10}Be~~	~~27~~	~~Yiou et al., 1997; Muscheler et al., 2004~~
		~~$\delta^{18}O$~~	~~10~~	~~Johnsen et al., 1997~~
~~GISP2~~	~~72.36N 38.30W~~	~~^{10}Be~~	~~162~~	~~Finkel & Nishiizumi, 1997~~
		~~$\delta^{18}O$~~	~~10~~	~~Stuiver & Grootes, 2000~~
~~WDC~~	~~79.46S 112.085W~~	~~^{10}Be~~	~~67~~	~~This study~~
		~~$\delta^{18}O$~~	~~10~~	~~Jones et al., 2017~~
		~~Non-sea-salt-Sulfate (nssS)~~	~~1~~	~~Buizert et al., 2018~~
~~Hulu Cave~~	~~32.5N 119.16E~~	~~^{14}C~~	~~271~~	~~H82: Southon et al., 2012~~
			~~124~~	~~MSD: Cheng et al., 2018~~
		~~$\delta^{18}O$~~	~~70~~	~~H82: Wang et al., 2001~~
			~~31~~	~~MSD: Cheng et al., 2016~~



~~*Figure 1 Cosmogenic radionuclide data used in this study.*~~

~~*On the GICC05 timescale: (a) NorthGRIP ^{10}Be concentrations (orange lines indicate discontinuities in data collection); (b) GRIP ^{10}Be concentrations (Yiou et al., 1997; Muscheler et al., Data are available from every 2nd ice core portion of 55 cm, leading to discontinuities in the dataset between each 27-year resolved data point (grey lines). (c) GISP2 ^{10}Be concentrations (Finkel & Nishiizumi, 1997). On the WD2014 timescale: (d) WDC ^{10}Be concentrations.*~~

~~*On the U/Th timescale: (d) Hulu Cave Δ^{14}C data as reported in three separate datasets. In the following, we refer to the H82 and MSD samples, but we exclude the off-axis H82 measurements (blue) by Southon et al. (2012), as they show more outliers and wider dating uncertainties.*~~

~~2.1~~2.2 Preparation of the WDC samples and measurement

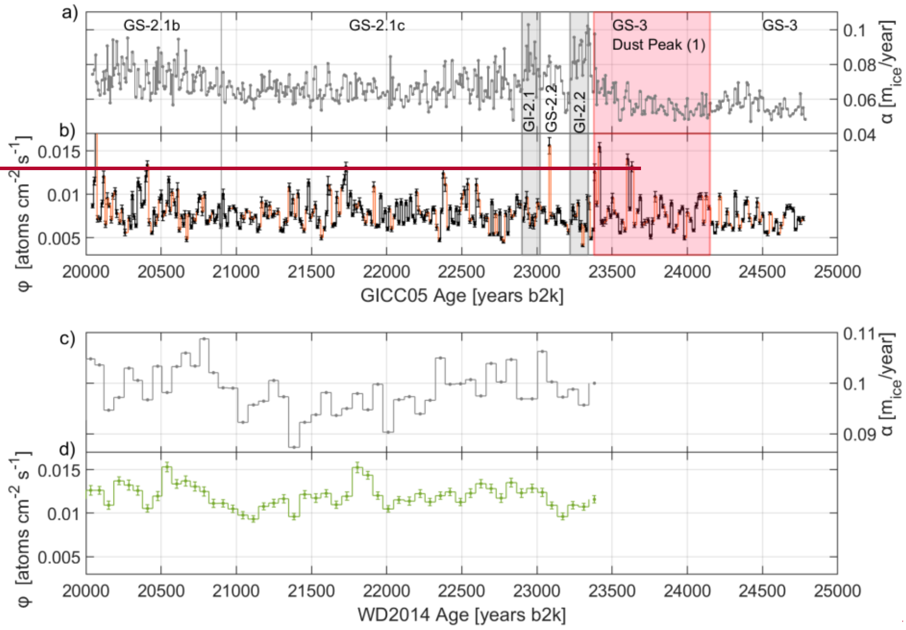
A total of 73 samples in the WAIS Divide 06A ice core (WDC-06A), from 2453 to 2599 m depth, were analysed for ^{10}Be concentrations at Purdue University. These samples represent continuous ice-core sections with a cross-section of ~2 cm^2~~ and a length of ~2 m (~~, varying from 1.89 to 2.12 m~~),~~ ~~corresponding to ~~~ in length (~60-75 years of snow accumulation per sample~~. See~~). The extraction followed established protocols (Woodruff et al. 2013), see Methods Appendix sec. 8.1 for more details. ~~The ~~Fig. 2d shows the measured WDC ^{10}Be concentrations~~ are shown in Fig. 1(d).~~.

~~2.2~~2.3 Conversion of ^{10}Be concentrations to fluxes

To account for the first-order correction of climatic influences on the ^{10}Be signal (Adolphi et al., 2018), we need to convert the ^{10}Be concentrations~~ need to be converted~~ to fluxes, which requires knowledge of accumulation rates (see Methods Appendix sec. 8.3). Accumulation rates for the ice cores are reconstructed using the annual layer thicknesses and an appropriate model for the layer thinning. The thinning in the LGM portion of the ice cores can be approximated by a linear function of depth, although there may be uncertainties that relate to the time scale itself.

For the NorthGRIP ice core, the layer thickness ~~is known by direct layer counting ~~was directly derived from the annual layers of the GICC05 (Andersen et al., 2006), while the thinning was modelled by Johnsen et al. (2001). ~~For GRIP and GISP2, the annual layer thickness is interpolated from the volcanic match to NorthGRIP (Rasmussen et al., 2008; Seierstad et al., 2014). During interstadials with well-resolved volcanic tie points, measurements indicate that the two Summit cores (GRIP and GISP2) have a stadial to interstadial accumulation increase that is 10% higher than at NorthGRIP (Seierstad et. al, 2014). Due to the scarcity of volcanic tie points across the LGM, the accumulation difference for GI-2.1 and 2.2 may be underestimated for GRIP and GISP2. The most recent version of thinning functions for the GRIP and GISP2 was used to calculate fluxes in this study (Lin et al., 2021; Hvidberg et al., 1997). ~~For WDC, the accumulation rate ~~was reported by ~~(Fudge et al~~. (~~., 2016) was inferred from the WD2014 layer thicknesses and modelled thinning (Buizert et al., 2015).

The ~~fluxes of the ~~new ~~datasets~~^{10}Be fluxes are shown in Fig. 2, together with the modelled accumulation rates. The fluxes of NorthGRIP are clearly less affected by changing snow accumulation rates than 10Be concentrations, as seen in the absence of GI/GS-related changes in 10Be, which is like other Greenland ice core 10Be-records (e.g., Muscheler et al. 2004). ~~The fluxes of NorthGRIP were checked for residual correlation with climate proxies (fig. S4), showing~~To show that the flux conversion largely removes the climatic influence on ^{10}Be deposition in Greenland, we checked the residual correlations with other climate proxies in Fig. S3.



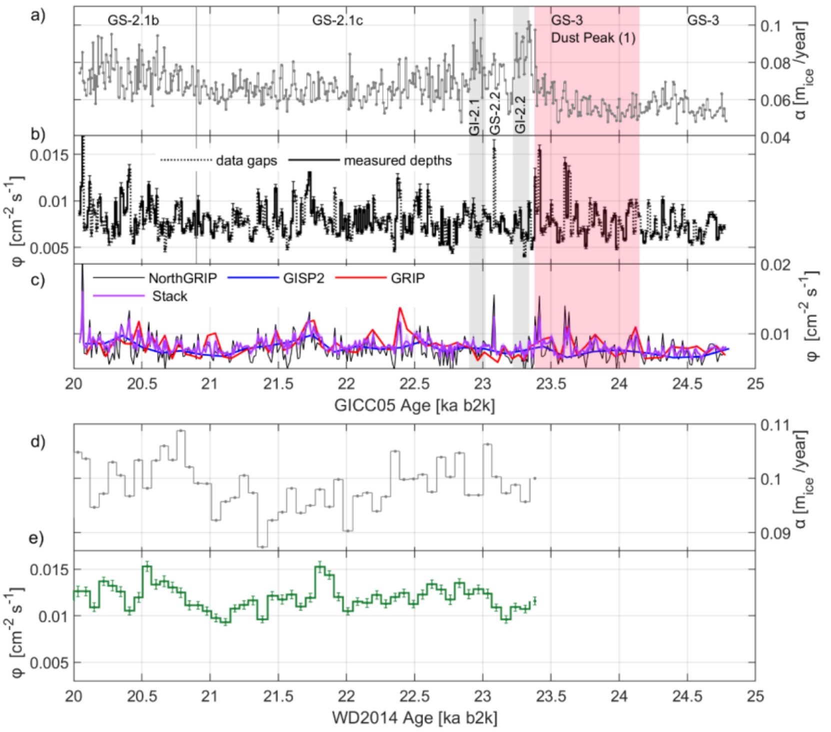


Figure ~~2 Accumulation rates and ^{10}Be fluxes of NorthGRIP (a, b) and WDC (c, d).~~
3 (a) The accumulation rate of NorthGRIP~~ was obtained from the GICC05 layer thicknesses and the strain model (Johnsen et al., 2001). The accumulation rate was converted to the same ~~, down sampled to the same depth resolution of each ^{10}Be data point~~ by averaging the accumulation rate between the top and bottom depths of each ^{10}Be sample and determining the uncertainty from the standard deviation of the annual accumulation within each depth interval. (b) The ^{10}Be fluxes of NorthGRIP show less climate-related fluctuations than the concentrations, as expected, and have ~~. (b) The ^{10}Be fluxes of NorthGRIP have an average of 0.008 ± 0.002 atoms cm^{-2} s^{-1}. ~~(c~~(c) Stack of the fluxes of the three Greenland ice cores, on 10-year resolution. (d) The published accumulation rate of WDC (Fudge et al., 2016~~) was~~), down~~-~~ sampled to the ~~same ~~age resolution of the ~~WDC ~~^{10}Be concentrations. (~~d~~e) The ~~WDC ~~^{10}Be fluxes ~~are shown on the same scaling as the NorthGRIP fluxes for comparison. The ~~of WDC average ~~WDC flux is~~at 0.011 ± 0.001 atoms cm^{-2} s^{-1}, higher than in Greenland, either because of depositional differences between the poles (Heikkilä et al., 2013) or because of accumulation rate inaccuracies.

Over the LGM, the GRIP and GISP2 ice cores were tied to NorthGRIP by identification of volcanic tie points and biomass burning events, and GICC05 was extended to these ice cores by interpolation (Rasmussen et al., 2008; Seierstad et al., 2014). The annual layer thickness is thus less certain for GRIP and GISP2 and, due to a scarcity of volcanic tie points across the LGM, the accumulation difference across the short interstadials GI-2.1 and 2.2 may be at risk of misinterpretation. We used the most recent versions of GRIP and GISP2 thinning functions to calculate fluxes (Lin et al., 2021; Hvidberg et al., 1997), but we actively modified the accumulation history using new ^{10}Be tie points, as delineated in section 3.2.

The ^{10}Be measurements recently performed in the LGM on the NEEM ice core (Zheng et al., 2021) have similar resolution as the GISP2 data set. The NEEM data set does not resolve the ^{10}Be features we are studying well, and we therefore do not consider this dataset further on.

Finally, an average of the ^{10}Be fluxes of the three Greenland ice cores was calculated by stacking the NorthGRIP, GRIP and GISP2 fluxes, using Monte-Carlo bootstrapping (Adolphi et al., 2018). The assumption behind averaging fluxes is that local accumulation effects are mostly removed by the conversion to fluxes and that the climatic effects on ^{10}Be deposition are the same over Greenland. For each iteration, three of the data series are selected with resampling, each dataset is perturbed within its uncertainties, and averaged. The stack is shown in Fig. 3c.

2.4 Carbon cycle modelling and uncertainties

The interaction of galactic cosmic rays (GCRs) with the atmospheric parent atoms (N, O) of ^{10}Be and ^{14}C is modulated by the time-varying helio- and geomagnetic fields. The radionuclides recorded in climatic archives ideally show synchronizable features in their production history (Steinhilber et al., 2012; Adolphi et al., 2018). The ^{14}C atoms enter the carbon cycle, which causes delay and smoothing of the atmospheric ^{14}C concentration relative to the production signal. However, the connection with the rapidly deposited ^{10}Be in ice cores (1-2 years depositional delay; Raisbeck et al., 1981) can be made using ~~A carbon-cycle model (here~~ a carbon-cycle model. Here, the box-diffusion model by Siegenthaler~~,~~ (1983) is ~~necessary~~applied to derive the atmospheric Δ^{14}C signal, i.e. the decay and fractionation-corrected ratio of ^{14}C/^{12}C relative to a standard (Stuiver & Pollach, 1977), from the measured ice-core ^{10}Be. This model was used extensively in works by Muscheler et al. (2000, 2004, 2009, 2014) and Adolphi et al. (2014, 2016, 2018).

We assume here that any variations of ^{10}Be in ice cores can be converted to a global ^{10}Be production rate, proportional to the global ^{14}C production rate. This assumption may, however, lead to uncertainties. For example, changes in the balance between wet and dry deposition or changes in the transport of ^{10}Be to the ice sheet are possible factors that might alter the signal recorded in ice cores from the true production rate.

The ice-core ^{10}Be is normalized relative to its mean, amplified by 20% (as estimated below), and provided as input signal to the model. The model should ~~be~~ run with parameters that best represent the state of the carbon cycle and its changes through time, with the ~~expectation~~assumption that any remaining variability will be related purely to production effects. However, as we do not know these carbon cycle changes well enough, we keep in mind, however, that the measured ^{14}C may have been affected by changes in the carbon cycle that are not considered in the model, adding additional signals that are not related to production rate changes.

~~By varying the model parameters, some residual effect of the parametrization is observable in the amplitude of modelled Δ^{14}C changes, but not in the timing of the changes, which is most important here.~~

In the period 20-25 ka b2k, the Hulu Cave ~~Δ^{14}C~~Δ^{14}C$_{calcite}$ is about 500~~ ‰ (fig. 1e~~‰ (Fig. 2e), which is higher than early Holocene values, which are below 200~~ ~~‰ (Reimer et al., 2020). These higher values may be related to one or more of the following factors: a lower ocean diffusivity during the LGM or any process that similarly reduces the carbon uptake by the ocean (Muscheler et al., 2004), a lower atmospheric ^{12}C inventory resulting in higher ^{14}C/^{12}C ratios (Köhler et al., 2022), or a weaker geomagnetic field. For instance, Muscheler et al. (2004) expect the ^{10}Be production rates in the LGM ~~are expected~~ to have been about 20% higher than today, due to the lower geomagnetic field intensity during the LGM~~ (Muscheler et al., 2004).~~.

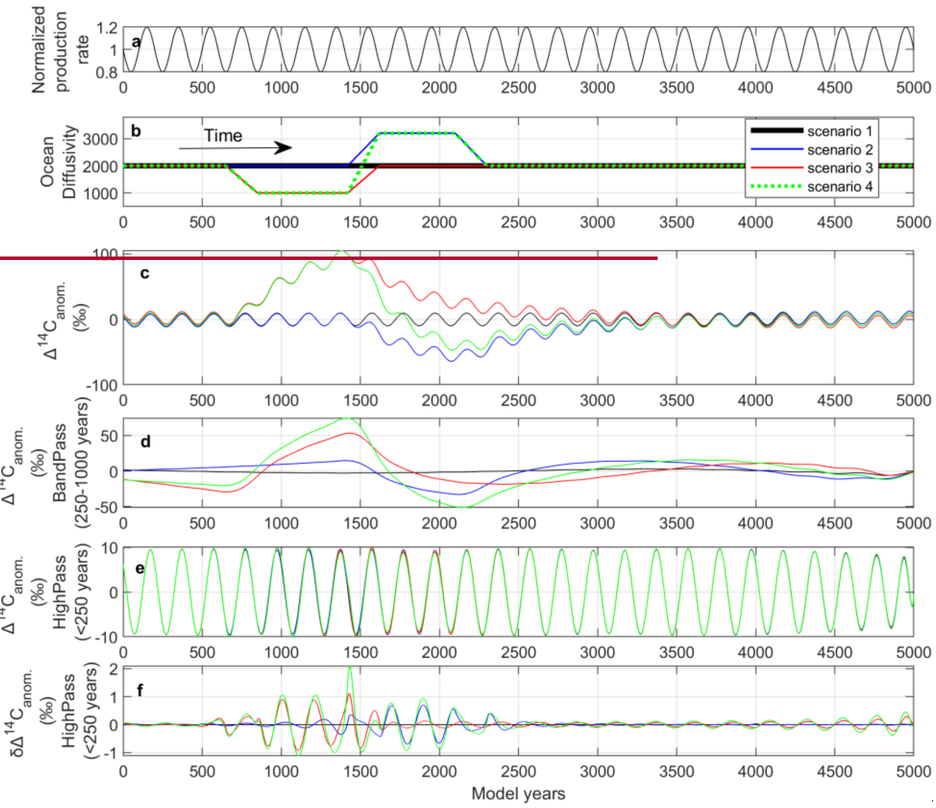
~~The strength of the geomagnetic field directly affects both the ^{10}Be and ^{14}C production rates. Although each radionuclide may be affected differently (Masarik & Beer, 2009), most studies do not find any significant difference in production rates (e.g. Kovaltsov et al., 2012; Herbst et al. 2017). Adolphi et al. (2018) also showed that around a change in the geomagnetic field, ^{10}Be production rates should be amplified by 30% to match the amplitude expected from geomagnetic field reconstructions. This could be explained by incomplete atmospheric mixing of ^{10}Be as the geomagnetic shielding effect of GCR's is largest at the equator.~~

To determine the most appropriate model parameters, we repeat the calibration by Adolphi et al. (2018) around the Laschamps geomagnetic excursion at 41 ka b2k, since the available ^{14}C data has been updated since then (Reimer et al. 2020). We run the model with different ocean ventilation values (~~fig. S2~~Fig. S1), finding that, for values of ocean diffusivity between 25-40% of the pre-industrial Holocene value, the modelled Δ^{14}C match the IntCal20 data best. ~~Moreover, this agrees~~This is in agreement with Muscheler et al. (2004) who performed a time-dependent adjustment of the ocean diffusivity parameter between 10-25 ka ~~to match the model to the measured $\Delta^{14}C$.~~b2k. In their study, the LGM ocean diffusivity was set to ~1000 m^2/yr, about 25% of the pre-industrial Holocene value. ~~We~~In summary, we find that a 20% production rate amplification of the normalized ^{10}Be and an ocean diffusivity of 25% of the pre-industrial Holocene value produce modelled outputs of about 500 ‰, in agreement with the Hulu-~~cave~~Cave measurements, although the model fails to capture the decreasing trend. We note that this is not necessarily a realistic ~~parameter~~parametrization of the state of the carbon cycle, but it allows us to match some of the main features seen in the data. We associate no uncertainty with the model parameters since no setup realistically explains all the ~~$\Delta^{14}C$~~$\Delta^{14}C_{calcite}$ features.

To compare the measured and the modelled $\Delta^{14}C$, in this study we will make use of linear detrending, as this largely removes the systematic offsets associated with the unknown carbon cycle history and inventories. The detrending is performed by, first, selecting data in the 20-25 ka b2k timeframe and, then, using the Matlab function "detrend", which subtracts the best-fit line from the data. Some residual effect of the parametrization is observable in the amplitude of detrended $\Delta^{14}C$, but not in the timing of the changes, which is most important here.

~~2.2.1~~2.4.1 Sensitivity tests: ocean diffusivity changes, accumulation rate uncertainties, measurement uncertainties.

~~We performed three separate sensitivity tests to~~To provide an uncertainty boundary for the detrended modelled $\Delta^{14}C$ curves, we performed three separate sensitivity test. As a first test, we investigated how short-term changes in the ocean diffusivity affect the modelled output, because during the LGM there likely have been changes in the carbon cycle around the GIs and the HS-2 (Bauska et al., 2021).



Normalized production rate
a
b
Time
Ocean Diffusivity
scenario 1
scenario 2
scenario 3
scenario 4
c
$\Delta^{14}C_{anom.}$ (‰)
d
$\Delta^{14}C_{anom.}$ (‰) BandPass (250-1000 years)
e
$\Delta^{14}C_{anom.}$ (‰) HighPass (<250 years)
f
$\delta\Delta^{14}C_{anom.}$ (‰) HighPass (<250 years)
Model years

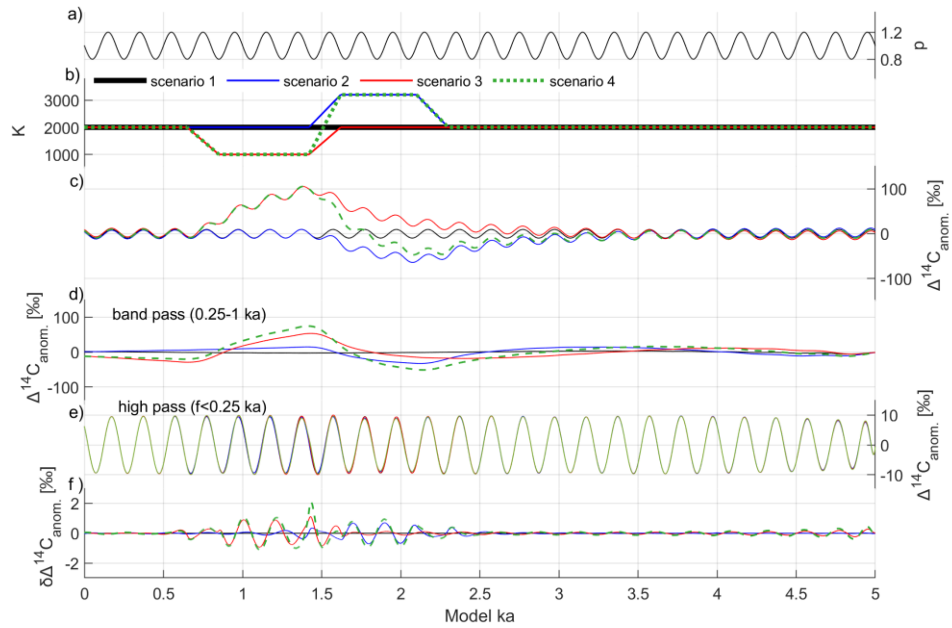
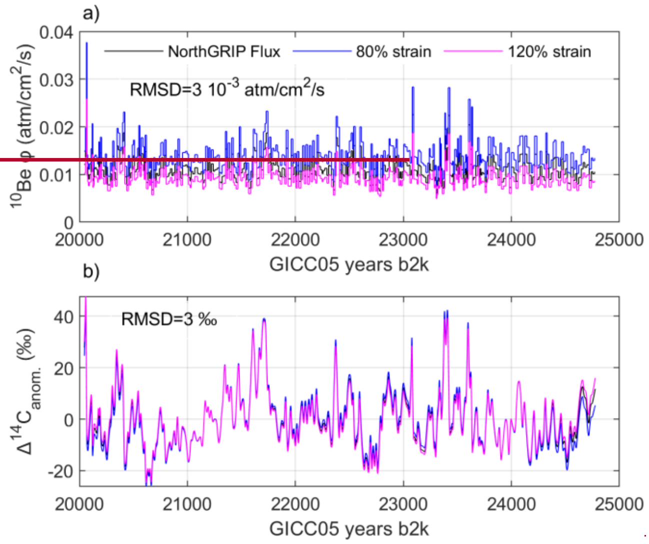


Figure ~~3~~ 4 Sensitivity tests for time-dependent ocean diffusivity changes.
(a) Normalized production rate input ~~showing~~shaped as a ~~cycle~~periodic wave with an amplitude of ±25% and a period length of 200 years, broadly consistent with typical solar de-Vries cycle variability ~~as~~ observed in ^{10}Be ~~data~~ (e.g.~~–~~ Wagner et al., 2001). (b) Scenarios of ocean diffusivity, as described in the text. (c) ~~Modelled~~Model output, after linear detrending. (d) Long-term variations of the output (band-pass filtered ~~output~~ with periodicity 250-1000 years). (e) High-pass filtered output, up to 250-year periodicity, having amplitudes of around 20-‰. (f) Differences of the high-pass filtered curves in (e) between the control scenario 1 and the other three scenarios (similar as in Adolphi & Muscheler, 2016). ~~Differences of around 1-2‰ demonstrate that there is little influence of the ocean diffusivity scenarios on the high-pass filtered outputs of the model.~~

A time-dependent change in ocean diffusivity was induced using 3 scenarios, as shown in ~~fig. 3b: either~~Fig. 4b. The control scenario is set at 50% of the pre-industrial value. We induce an abrupt increase in ocean ventilation (~~higher diffusivity:~~to 80% of the pre-industrial value), an abrupt decrease ~~in ocean ventilation (lower diffusivity:~~(to 25~~% of the pre-industrial value),~~%), or a sequence of both.~~ The control scenario is set at 50% of the pre-industrial value.~~ The duration of the events is chosen to reflect changes in the NorthGRIP calcium record, while the transition time was set to 200 years. ~~The input signal is a 200-year periodic wave with similar values to the NorthGRIP normalized ^{10}Be concentrations.~~

Figure ~~3c~~4c shows that the effects of the perturbations in ocean diffusivity on $\Delta^{14}C$ ~~are quite high and~~ span several tens of ~~‰,~~permille after detrending. Thus, any feature in the ^{14}C records that is in the proximity of an abrupt climate change and has a comparable duration is uncertain and should not be used for matching. On the other hand, ~~fig. 3d~~Fig. 4e shows that short-term variations of the modelled $\Delta^{14}C$ signal are less affected by the diffusivity perturbation. At least in principle, signals exceeding 10-20‰ that are much shorter than the climatic transitions could be used for wiggle-matching, since ~~fig. 3f~~Fig. 4f shows that the different ocean diffusivities scenarios affect the high-passed $\Delta^{14}C$ only within 2‰, hardly above measurement uncertainties.



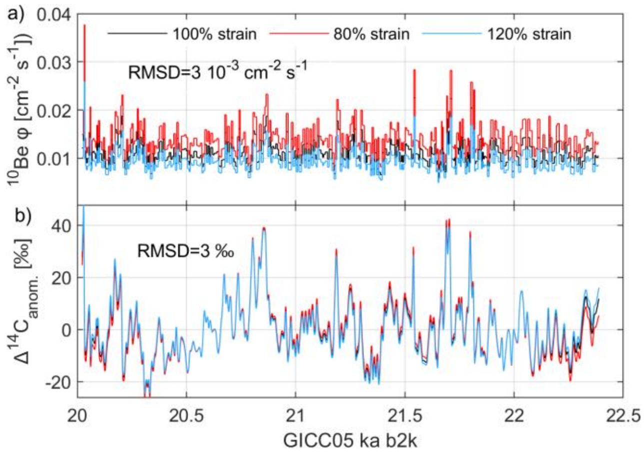


Figure ~~4~~ 5 Sensitivity test of the effects of ~~strain-model related~~ accumulation-~~rate~~related uncertainties on the carbon-cycle modelling.

(a) Two strain-model scenarios produce different flux values. ~~The RMSD quoted in the figure represents the average distance of the curves from each other.~~ (b) The modelled $\Delta^{14}C$ curves are detrended, which largely removes the differences between the scenarios. However, the ~~remaining variability is represented by the~~ RMSD between the curves, 3 ~~‰, which can be taken as the uncertainty derived from the ±20% perturbation of the ice accumulation model.~~‰, represents the remaining variability.

A second sensitivity test was performed to investigate the effect of accumulation~~-rate~~ uncertainties related to the strain model. We performed 2 experimental model runs where we shifted the thinning function (inverse strain) by ±20% of the mean value between 20 and 25 ka b2k, which is a realistic modelling uncertainty between independent studies of the accumulation rate (Rasmussen et al., 2014; Gkinis et al., 2014). Changing the strain rate creates an uncertainty of about 3~~ ~~‰ between the modelled and detrended $\Delta^{14}C$ curves, as shown in ~~fig. 4b~~Fig. 5b. Furthermore, with a similar approach, we quantify the impact of ^{10}Be ~~measurements~~measurement uncertainty on the modelled $\Delta^{14}C$ to be 1~~ ‰ (fig. S3~~‰ (Fig. S2). Adding these independent contributions in quadrature, we set an uncertainty for our modelled $\Delta^{14}C$ of 5~~ ~~‰. This uncertainty serves as an initial parameter for the wiggle-matching algorithm, described in the next section; it furthermore agrees with the uncertainty adopted for the comparison of centennial variations of ^{14}C and ^{10}Be during the stable climate of the Holocene (Adolphi & Muscheler, 2016). However, as we have verified with our test, $\Delta^{14}C$ changes in the vicinity of climate perturbations bear a considerably higher uncertainty.

~~2.3~~2.5 The wiggle-matching algorithm reproduced from Adolphi & Muscheler (2016) and its uncertainty

~~Along with the visual inspection, an important tool for the quantification of offsets between timescales is the~~A wiggle-matching algorithm, adapted by Adolphi & Muscheler (2016) from the original formulation by Bronk Ramsey et al. (2001), ~~and described in detail therein.~~represents an important tool for the quantification of offsets between time scales, along with visual inspection. The first input for the algorithm is the detrended $\Delta^{14}C$ ~~as~~ modelled from the ice-core ^{10}Be ~~concentrations or fluxes.~~. The second input is the detrended ~~$\Delta^{14}C$~~$\Delta^{14}C_{calcite}$ data from the Hulu Cave stalagmite samples. The output of the algorithm is a probability density ~~function~~ of the ~~possible timescale~~time-scale offsets, $\vec{t}_{offset}$.

Following the approach by Adolphi & Muscheler (2016), we do not allow for stretching of the underlying chronologies. We investigated the offset probability within partially overlapping time windows $\vec{W}$, spaced every 50 years, obtaining a two-dimensional probability matrix $\boldsymbol{P}(\vec{W}, \vec{t}_{offset})$. ~~This is intended as a way to study the offset in a time-dependent fashion, since the algorithm does not allow for stretching of the underlying timescales.~~ We summarize the algorithm settings in the Method Appendix.

The ~~position of the time-dependent mode of $\boldsymbol{P}$ is here defined to be the~~ most likely ~~timescale~~ offset, $t^*_{offset}(\overrightarrow{W})$~~). By observing whether this latter quantity~~) is here defined as the maximum of $\boldsymbol{P}$. We identify the boundaries T1 and T2 between which $t^*_{offset}(\overrightarrow{W})$ is stable and ~~weakly time-dependent,~~ we ~~can visually mark some T1 and T2 limits for the purpose of producing~~calculate an average offset~~ within a subset of windows $\overrightarrow{W}$. An estimate of the average offset was thereby produced for the GICC05 and the WD2014 timescales.~~.

After shifting each ice-core dataset by the proposed offset, we computed the χ^2 test between the ~~speleothem and ice-core~~ $\Delta^{14}C$ curves. If the p-value was outside the 0.01-0.99 interval, we repeated the wiggle-matching using the standard deviation of the residuals (RMSD in ‰) as the uncertainty for the modelled $\Delta^{14}C$ curve. ~~We then repeated the wiggle matching again and plotted the new ice-core t^*_{offset} curves.~~ We re-evaluated the boundaries T1 and T2~~,~~ and re-averaged the t^*_{offset} curves, obtaining a ~~second~~final estimation for the offset.

~~For this second offset, we propose~~We aim to establish an uncertainty ~~estimation which~~on the offset that considers the ~~timescale~~ uncertainties of both ice cores and speleothems time scales, as well as resolution limitations.~~ We consider the information gained by running the wiggle matching with multiple ice-core ^{10}Be datasets, including concentrations and fluxes. Moreover, we consider the time-dependent behavior of the uncertainties.~~ We therefore applied a Monte-Carlo protocol to estimate the uncertainty and conducted the following steps:

1) For each modelled $\Delta^{14}C$ dataset and window W^*, we approximate the $\boldsymbol{P}(W^*, \vec{t}_{offset})$ as a Gaussian distribution~~ centered around the mode and~~, with 1σ as the average lower and upper width at half maximum of $\boldsymbol{P}(W^*, \vec{t}_{offset})$. By arbitrarily choosing the mode of **P** as the best offset, we can disregard any other lobe of probability as spurious~~ or unnecessary~~, with ~~some~~high confidence given by our visual inspection of the data.
2) For each dataset and window W^*, we sample randomly from the Gaussian for 10000 times to derive an ensemble of ~~timescale~~time-scale offsets;
3) By iterating steps 1 and 2 across all datasets and all windows within the established time boundaries T1 and T2, we compute the overall histogram of the sampled offsets;
4) We evaluate the 68% confidence interval of this histogram, around the best offset established as the mode.

This procedure is repeated separately for WD2014 and GICC05, and also for the Hulu H82 and MSD datasets. The algorithm is~~ made~~ available in the Supplement.

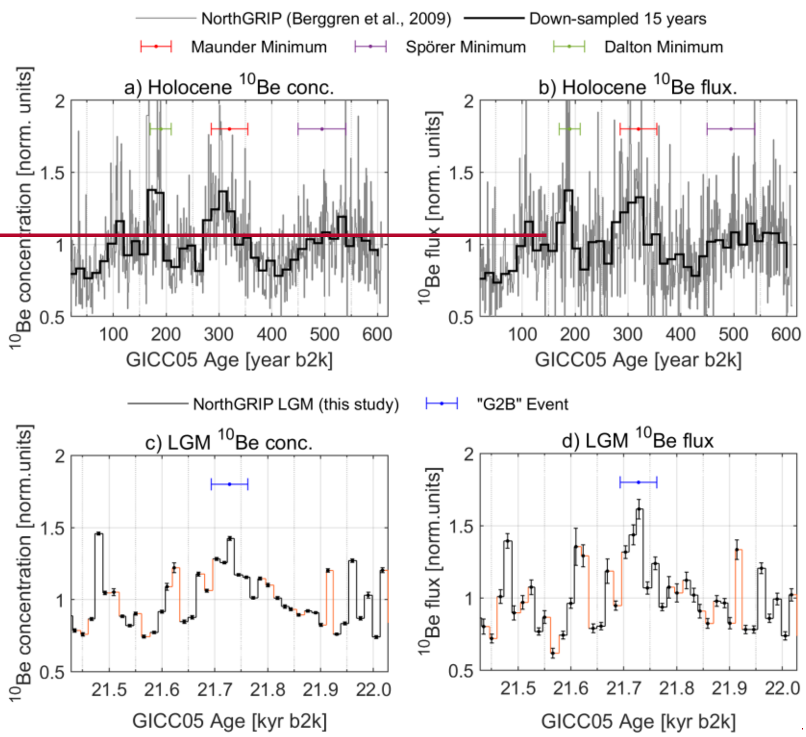
3 Results

3.1 A promising inter-ice-core tie point for ^{10}Be synchronization

Previously used for the matching by Adolphi et al. (2018), a ^{10}Be increase at 21.7 ka b2k (GICC05 age) is visible in the new NorthGRIP data as well as in the WDC dataset (~~fig. 1~~Fig. 2). This radionuclide increase resembles the ^{10}Be signal associated with the Maunder Solar Minimum (1645-1725 CE), which was a period of low solar activity and consequent increase in the global radionuclide production (Berggren et al., 2009; Eddy, 1976). In figure ~~5~~6, we compare the Holocene and LGM counterparts of ^{10}Be at NorthGRIP finding similar shapes and duration~~, which~~. This supports the attribution of the ^{10}Be increase to a solar minimum during GS-2, which could explain the co-registration across ~~^{14}C~~$^{14}C_{calcite}$ and ^{10}Be datasets.

~~In the Holocene, high accumulation rates make the wet deposition of ^{10}Be predominant over dry deposition, so concentrations may be more representative of the true ^{10}Be production rate (Berggren et al., 2009). During the glacial, model runs (e.g. Heikkilä & Smith, 2013) suggest that wet deposition is still predominant, but the abrupt accumulation rate changes observed across GIs require examining the fluxes to separate dilution effects from production effects. We examine both concentrations and fluxes in fig. 5, finding that for both fluxes and concentrations, the LGM and Holocene signals are very similar.~~

Based on the observed similarities in ~~fig. 5~~Fig. 6, we call the ~~22~~21.7 ka b2k increase the "GS-2.1c ^{10}Be Event" (abbreviated *G2B event* in the following~~) without claiming a certain solar origin of the signal.~~). To support the bipolar synchronization, after resampling the NorthGRIP ^{10}Be data on the resolution of WDC, we observe that the two ice cores register similar ^{10}Be amplitudes around the G2B event. The flux increases by 0.003 atoms cm^{-2} s^{-1} at both sites, which represents an increase of 40% and 30%, respectively, from the flux average values at NorthGRIP and WDC. The ^{10}Be concentration, on the other hand, increased only 30% (1.3 x 10^{4} atoms/g) and 20% (0.9 x 10^{4} atoms/g), respectively, from the concentration average values at NorthGRIP and WDC.


NorthGRIP (Berggren et al., 2009)
Down-sampled 15 years
Maunder Minimum
Spörer Minimum
Dalton Minimum
a) Holocene ^{10}Be conc.
^{10}Be concentration [norm. units]
GICC05 Age [year b2k]
b) Holocene ^{10}Be flux.
^{10}Be flux [norm. units]
GICC05 Age [year b2k]
NorthGRIP LGM (this study)
"G2B" Event
c) LGM ^{10}Be conc.
^{10}Be concentration [norm.units]
GICC05 Age [kyr b2k]
d) LGM ^{10}Be flux
^{10}Be flux [norm.units]
GICC05 Age [kyr b2k]

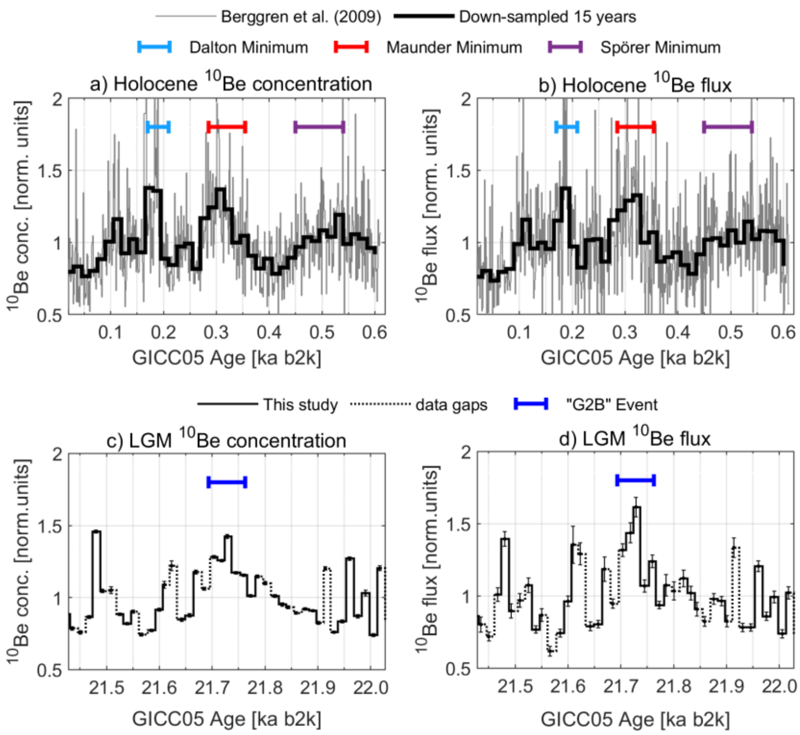


Figure ~~5~~ 6 Comparison of ^{10}Be ice-core data in the Holocene, around the Maunder Minimum, and in the LGM, around the G2B event.

~~By normalizing all data (dividing~~Data was divided by ~~their~~the mean~~)~~ within ±300 years of the central event~~,~~ ~~we ensure a better visual comparison of the data.~~. (a, b) In the Holocene, the NorthGRIP ^{10}Be data by Berggren et al. (2009) ~~are compared~~compare well to the defined durations of the grand solar minima, which were independently recorded in the sunspot observations. ~~The down-sampling to 15 years allows for easier comparison to the LGM dataset below. We can visualize the effect of solar minima as an~~An increase in the ^{10}Be production rate by about 40-50% from the average is produced by the solar minimum. The fluxes show a more abrupt increase in time, while concentrations record a more gradual increase. (c, d) In the LGM, the similarity of shape and duration to the Maunder Minimum supports the identification with a solar minimum.

3.2 Synchronization between ice cores using ^{10}Be

Having established the G2B event as a radionuclide production feature, we can synchronize Greenland ice cores by inserting new ^{10}Be tie points between NorthGRIP~~ and the~~, GRIP, and GISP2

~~ice cores~~. Furthermore, the G2B event can be used to improve the bipolar matching to Antarctica (~~fig. 6~~Fig. 7).

We decided to compare the ^{10}Be data on similar resolutions to facilitate the identification of other common production features. Hence, we down-sampled the high-resolution data of NorthGRIP to the same resolution as GRIP and WDC, ~~as~~since ice-core ^{10}Be measurements are averages over the sampling depth intervals and lose variability with decreasing resolution.

Table 2 ^{10}Be tie-point ages between NorthGRIP, GRIP, and WDC. The internal difference between the Greenland ice cores (δ) reaches 27 years at the G2B event. Likewise, the difference between WDC and NorthGRIP ages (Δ) indicates older ages for WDC at the G2B event and the youngest no. 4 tie point.

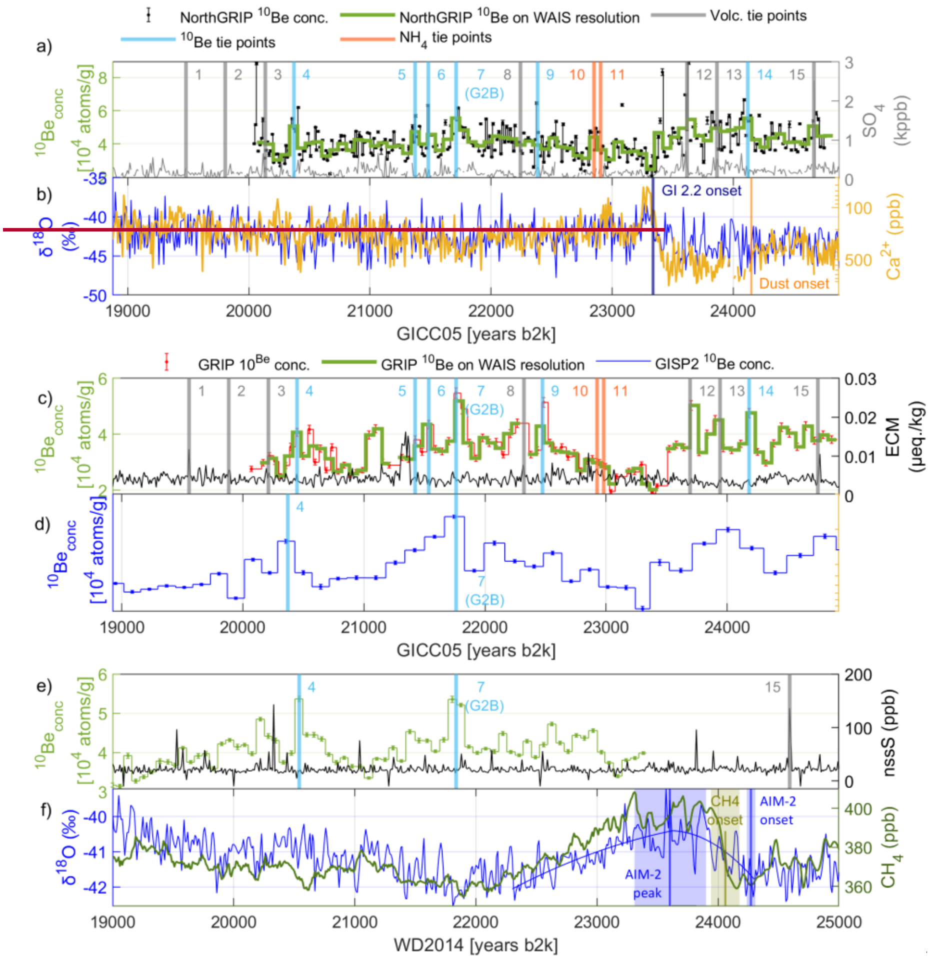
	GICC05 Age (years b2k)			WD2014 Age (years b2k)	
tie point	NorthGRIP	GRIP	δ	WDC	Δ(WDC-NorthGRIP)
4	20373	20370	3	20541	168±40
5	21372	21347	25		
6	21484	21458	26		
7 (G2B event)	21710	21683	27±21	21835	125±40
9	22383	22398	-15		
14	24119	24103	16		

~~In figure 6,~~Figure 7 shows the ^{10}Be concentrations from NorthGRIP, GRIP and WDC ~~are shown~~ on their respective ~~time scales~~chronologies, together with a set of published ~~non-climatic~~ tie points (Seierstad et al., 2014; Svensson et al., 2020). Between NorthGRIP and GRIP, we observe important similarities in the ^{10}Be data, which leads us to suggest 6 new ^{10}Be tie points: a peak at 20.4 ka b2k, a double peak at 21.5 ka b2k, the G2B event at 21.7 ka b2k, a single peak at 22.4 ka b2k, and a triple peak structure between 23.5 and 24.2 ka b2k. These tie points cover the previously tie-point-free section across GS-2.1b/c.

The ~~match~~tie points to WDC, of which ~~the oldest tie point (no. 15) was published by~~ Svensson et al. (2020) published the oldest (no. 15), is extended with the aid of two additional ^{10}Be tie points (no. 4 and no. 7/G2B). The choice of no. 4 as a bipolar tie point is motivated by a similar layer count of ~1300 years from G2B and a similar shape of the signals. The ages of the tie points are summarized in Table 2 and derived from the mid-depth of the highest peak.

~~The ages of the tie points are summarized in table 2 and derived from the mid-depth of the highest peak. The timescales~~The time scales of GRIP and NorthGRIP are slightly misaligned between 21 and 23 ka b2k by up to 27 years at the G2B event, probably ~~due to the fact that~~because GRIP ages were interpolated between widely spaced tie points. The uncertainty of this misalignment can be estimated as half the sum of the resolutions of NorthGRIP and GRIP measurements (± 21 years, 1 σ), as this is likely to limit our matching precision in each direction. Therefore, ^{10}Be measurements cannot be said to resolve matching issues between Greenland ice cores with very high precision, nonetheless we will use the new ^{10}Be tie points in the following to produce an updated accumulation rate for GRIP.

Furthermore, WD2014 and GICC05 are misaligned by 125±40 years at the G2B event, WD2014 being older; the uncertainty on this offset is given as half the sum of the resolutions of WDC and NorthGRIP data.



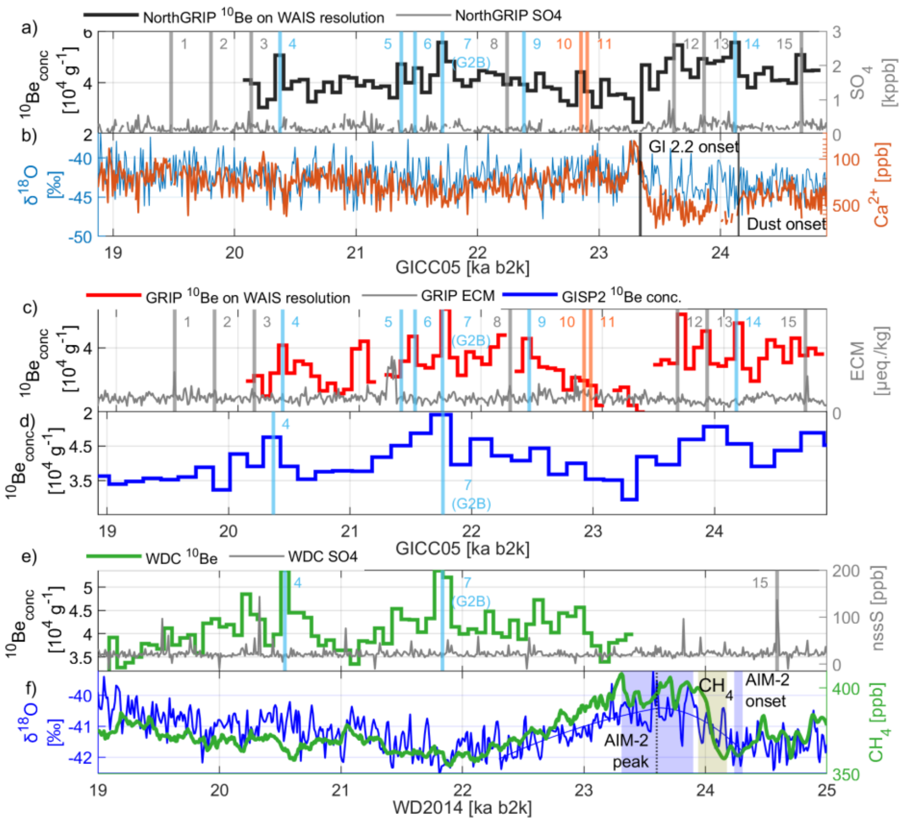


Figure ~~6~~7 Bipolar tie points and climatic proxies. Vertical bars indicate ~~tie-point positioning.~~ the type of tie points (grey: volcanic, blue: ^{10}Be, orange: ammonium). The data were aligned using tie point 7 (G2B event), without stretching the ~~timescales~~chronologies. (a) The NorthGRIP tie points are based on ^{10}Be ~~concentrations~~ (black~~) were down-sampled to the WDC resolution (green). The~~), sulfate data (grey) ~~supports the volcanic match, while~~and ammonium data~~, on which tie points no. 10 and 11 were based, are~~ (not shown ~~for clarity.~~). (b) ~~δ^{18}O and Ca^{2+} at~~ NorthGRIP ~~(calcium is on an~~ $\delta^{18}O_{ice}$ and Ca^{2+} (inverted log-scale) qualitatively represent the climate ~~and determine the timing of the GIs and the GS-3 dust peak.~~

in the LGM. (c) The GRIP tie points are based on ^{10}Be ~~concentrations~~ (red), ~~having a 27-year resolution, were down-sampled to the WDC resolution (green).~~ ECM (~~black) shows volcanic tie points, although some eruptions are better visible in the~~ grey), sulfate ~~signal~~ (not shown~~).~~), ammonium (not shown). (d) The ^{10}Be data of GISP2 (blue) has a low resolution~~, hence the alignment with NorthGRIP cannot be improved~~. Tie points no. 4 and 7 are however sufficiently visible in the data. (e) WDC ^{10}Be data (green) and nssS (~~black~~grey; no. 15 by Svensson et al., 2020). (f) WDC climatic proxies, with CH_4 presented on the gas chronology by Buizert et al. (2015). ~~We observe the occurrence of the AIM-2 warming in Antarctica as an increase of $\delta^{18}O$ (blue). The approximate shape of the AIM-2 was calculated by a second-order polynomial fit. The AIM-2 onset and peak are characterized in the text. The age of the increase of CH_4 by ~50 ppb (orange) was calculated by detecting where the signal increased significantly above the mean, with its gas-age uncertainty (green shading, see text for details).~~The ages of the AIM-2 onset and peak and of the methane increase are described in the text.

~~To improve the GRIP time scale, we calculated a new depth-depth interpolation between the two Greenland ice cores and we obtained a timescale correction, which we apply in the following to the GRIP data. On the resolution of GISP2, ^{10}Be appears to be sufficiently aligned around the G2B event and the peak at 20.4 ka b2k, hence no time scale correction was applied to GISP2.~~

The fact that the G2B event is 27 years older in NorthGRIP than GRIP ~~also~~ means that the GRIP time scale and the accumulation rate ~~needs~~need to be corrected, because ~~of the corresponding changes to the layer thicknesses induced by~~ the new tie points~~. This~~ induced a change in the layer thickness. Hence, we calculated a new depth-depth interpolation between the two Greenland ice cores, and we assigned an updated age scale to the GRIP depths.

The corrected accumulation of GRIP was ~~done~~computed by multiplying ~~the GICC05 timescale of GRIP~~ its layer thickness by a correction function ~~that computes~~of the relative change ~~of the layer thickness~~ between the tie points. For example, since NorthGRIP has 1235 layers between tie-points nr. 4 and 7, while GRIP only has 1210 layers, the correction factor for the accumulation rate between these tie-points is 1.02. Due to these relatively small corrections, we do not find it necessary to apply any smoothing to the correction function. In the following, we denote fluxes of GRIP that were corrected in this way as “corrected fluxes” when we need to distinguish them from the previous version used by Adolphi et al. (2018). Although the effect of the accumulation correction is small, ~~in section 3.3 we will show that this method~~it has ~~a significant~~an impact on the amplitude of the G2B event signal in GRIP, making the NorthGRIP and GRIP modelled Δ^{14}C curves look more alike. At the resolution of the GISP2 data set, ^{10}Be appears to be sufficiently aligned around the G2B event and the peak at 20.4 ka b2k, hence no age correction was applied to GISP2.

In ~~figure 6~~Figure 7, panels b and f, selected climatic proxies are shown to illustrate that a bipolar match across the LGM is already sufficient to change the timing between the two GIs and the AIM-

2, bringing the AIM-2 to be closer to the GI-2.2 onset (a "before version" of the alignment of the climatic proxies ~~can be found~~is shown in ~~fig. 7~~Fig. 1).

In the WDC isotope data, we fit a curve to the ~~$\delta^{18}O$~~$\delta^{18}O_{ice}$ record over the AIM-2 period to determine both the likely age for the onset of the warming slope and the peak – all on the original WD2014 ~~time scale~~chronology. We used the Matlab function 'WDC_breakpoint' provided by WAIS project members (2015), which fits the AIM with a double-polynomial fit to identify the maximum. However, the fit is sensitive to the starting guess for the position of the maximum; by varying the starting guess between 23 and 25 ka b2k, we observed that the fit finds several maxima. This fact is attributed to the shape of the signal being ambiguous and very broad. Moreover, a visual maximum of the AIM-2 is clearly identifiable at 23.6 ka b2k. Taking this as the central estimate, ~~from the distribution of the fitted breakpoints,~~ we ~~obtain~~estimate the location of the AIM-2 peak as 23600 ± 300 years b2k (1 σ). On the other hand, the onset of the AIM-2 is more ~~precisely~~easily defined to be 24272 ± 35 years b2k (1 σ). ~~We also determine the~~

The onset of the CH_4 increase at 24060 ± 118 (1σ) years b2k~~,~~ is defined as the instant at which the signal becomes higher (and remains higher) than 1 standard deviation from the average baseline of the period 24.5-25.5 ka b2k. The uncertainty of the CH_4 onset is quoted from the gas age uncertainty of WD2014 (Fudge et al., 2017), which is mostly useful to compare CH_4 to the ~~$\delta^{18}O$~~$\delta^{18}O_{ice}$ signal of the same ice core, WDC. All values are summarized in ~~table~~Table 3.

By aligning the two ice-core time scales at the G2B event, and without stretching the ~~time scales~~chronologies, we shift the GICC05 data by 125 years compared to WD2014 data. We observe that the AIM-2 peak now happens 125 years before the onset of GI-2.2, instead of 250 years before as was the case on the original time scales. The alignment at the G2B event is not sufficient to alter the order of the GI-2–AIM-2 sequence. What the new alignment shows is that the AIM-2 warming occurs within the dust peak and that the AIM-2 cooling ~~is starting close to~~occurs with the dust peak termination in Greenland, with potential influence of the GIs.

Table 3 Event onsets and terminations discussed in this work, corrected for the time-scale offsets, with 1σ confidence intervals. In the 'Original Calendar age' column, the onset of GIs and dust terminations are based on the original definition by Rasmussen et al. (2014) with reported uncertainties related to the event definition only. The dating uncertainty is dominated by the uncertainty of GICC05, which is about 600 years at this age. The ages of the AIM-2 and CH_4 increase in WDC were assessed as described in the main text. In the column 'Age according to wiggle-matching synchronization', GICC05 has been shifted by 375^{+250}_{-300} years, while the WD2014 ages were shifted by 225^{+200}_{-250} years, both towards older ages. The uncertainty on the new age was propagated from the original uncertainty and the offset uncertainty, using the asymmetric 1σ-boundaries of the offset.

Event	Archive	Age according to wiggle-matching synchronization (y b2k $^{+\sigma^+}_{-\sigma^-}$)	Original Calendar age (y b2k ± σ)
HS-2 onset	Hulu $\delta^{18}O_{calcite}$	24710 ± 40	24710 ± 40
Dust onset	NorthGRIP Ca^{2+}	24525^{+250}_{-300}	24150±10
AIM-2 warming onset	$\delta^{18}O_{ice}$ WDC	24497^{+200}_{-250}	24272±35
Methane onset	WDC CH_4	24285^{+230}_{-270}	24060±118
AIM-2 cooling onset	$\delta^{18}O_{ice}$ WDC	23825^{+360}_{-390}	23600±300
Dust termination	NorthGRIP Ca^{2+}	23755^{+250}_{-300}	23380±20
Start of GI-2.2	$\delta^{18}O_{ice}$ Greenland	23715^{+250}_{-300}	23340±20
Start of GS-2.2	$\delta^{18}O_{ice}$ Greenland	23595^{+250}_{-300}	23220±20
Start of GI-2.1	$\delta^{18}O_{ice}$ Greenland	23395^{+250}_{-300}	23020±20
Start of GS-2.1c	$\delta^{18}O_{ice}$ Greenland	23275^{+250}_{-300}	22900±20

3.3 Carbon cycle modelling and wiggle-matching

~~Finally, an average of the ^{10}Be fluxes of the three Greenland ice cores was calculated by stacking the NorthGRIP, GRIP (corrected flux) and GISP2 fluxes, using Monte-Carlo bootstrapping (Adolphi et al., 2018). fig. S1, with uncertainty bands derived from the standard deviation of the 1000 simulated fluxes.~~

~~Stacking the fluxes combines the information from three ice-core locations and therefore results in a different Δ^{14}C than the interpolation over NorthGRIP data gaps.~~

Inspection of the measured ~~Δ^{14}C (fig. 7a~~$\Delta^{14}C_{calcite}$ (Fig. 8a) and the ^{10}Be-based Δ^{14}C (~~fig. 7~~Fig. 8 b, c, e) confirms the expectation that a shift of the two ice-core ~~timescales~~chronologies towards older ages is required for a better alignment with the Hulu-cave ~~^{14}C~~$^{14}C_{calcite}$ record.

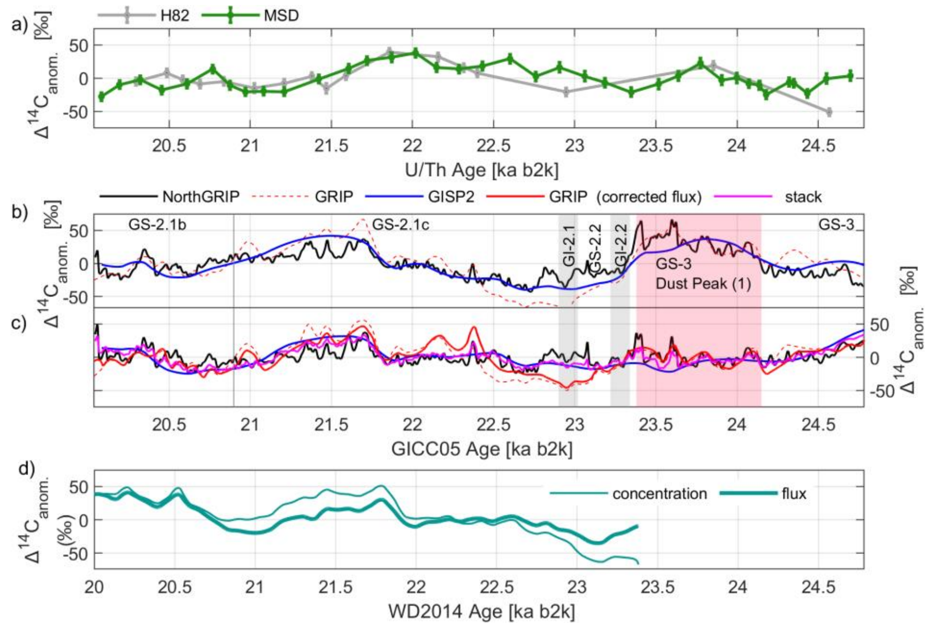


Figure 8 Carbon-cycle modelled $\Delta^{14}C$ compared to measured Hulu Cave data, before synchronization. (a) Hulu Cave $\Delta^{14}C_{calcite}$ (b) Modelled $\Delta^{14}C$ from Greenland ^{10}Be concentrations. (c) Modelled $\Delta^{14}C$ from Greenland ^{10}Be fluxes, including the flux stack (magenta). The effect of the flux correction of GRIP is mostly visible around the G2B event, where the corrected data (solid red line) shows a better agreement with the other cores. (d) $\Delta^{14}C$ modelled from WDC ^{10}Be fluxes and concentrations.

We observe two non-climatic tie-points between all $\Delta^{14}C$ curves:

1) The G2B event: a relatively abrupt increase of 30 ‰ in the modelled $\Delta^{14}C$ ~~from ^{10}Be, reaching its maximum~~ at 21~~,725 years~~.7 ka b2k (GICC05 ages~~), about 100 years after the maximum is reached in ^{10}Be fluxes.~~). The most likely equivalent of this event is observed in the Hulu Cave data at around 22~~,200 years~~ ka b2k (U/Th ages), where an increase of about 25 ‰ happens abruptly between two data points which is followed by a slow decrease. In the H82 data by Southon et al.~~,~~. (2012), the increase is much less abrupt and spans 4 data points.
2) A ~~smaller~~40‰ peak is observed in the ice-core ~~data~~$\Delta^{14}C$ at 20~~,400 years~~.4 ka b2k (GICC05). In Hulu Cave data, one elevated data point at 20~~,800 years~~.8 ka b2k (U/Th) is visible in the MSD dataset, but the H82 record does not show this peak.~~ In the original ^{10}Be fluxes, this event is probably caused by the combination of two peaks, one at about the same age and one~~

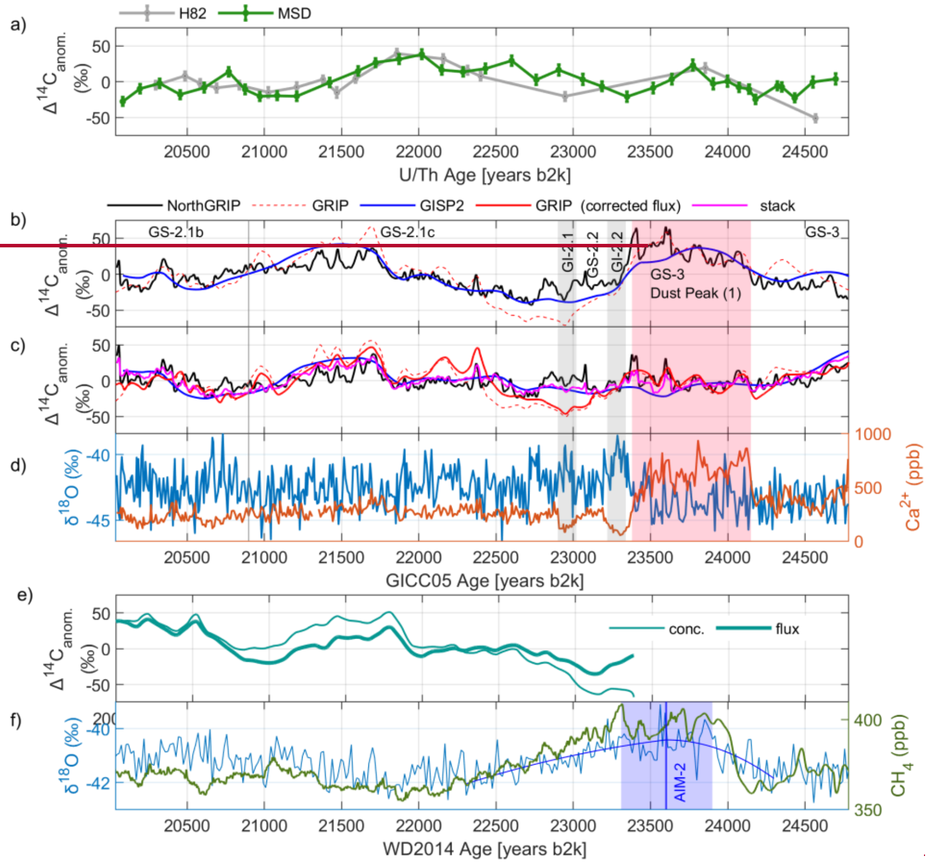
~~at 20,600 years b2k (e.g., see fig. 2). The amplitude of the event is about 40 ‰ in the modelled $\Delta^{14}C$.~~

We assume that the modelled and the measured $\Delta^{14}C$ should, after detrending, be dominated by the same production signal and that the differences we observe in the datasets can be explained by noise, modelling uncertainties, and dynamics affecting the deposition rather than the production of the radionuclides. Therefore, we proceed with the wiggle-matching despite the observed differences.

Based on these considerations, we run the wiggle-matching algorithm exclusively in the 20.5-22.5 ka b2k range (U/Th ages), using data windows of 1300 years with an overlap of 50 years between successive windows. The 1300-year choice is motivated by a compromise between enough signal inclusion in each window and more precision, since preliminary tests showed that this value returned the most stable results, compared to windows of 1000 or 1500 years. For the U/Th time scale, we use both Hulu datasets but we separate our treatment for the H82 and the newer MSD data (Southon et al., 2012; Cheng et al., 2018), because of the different resolution and the slightly different signals recorded. For the WD2014 ~~timescale~~time scale, we use both the WDC ^{10}Be concentrations and fluxes, modelled to $\Delta^{14}C$, since they are similar after carbon modelling. For the GICC05 ~~timescale~~time scale, we use all available $\Delta^{14}C$ modelled from concentrations, fluxes, and the flux stack, since they are similar at least around the production features we are going to match.

By averaging the obtained offset curves across the datasets and the interval (~~fig. S5~~Fig. S4), we estimate the ~~initial time scale~~chronological offset to be ~~370 years~~375years for GICC05 and 225 years for WD2014, using the MSD dataset exclusively. We do not give an uncertainty boundary for these offsets just yet, as we first perform a χ^2 test, following the protocol outlined in the Methods section. We evaluated the χ^2 and the associated p-value between each shifted $\Delta^{14}C$ curve and the Hulu dataset. For the flux-based datasets, the p-values were within the 0.01-0.99 threshold, except for the GRIP 'uncorrected' fluxes. This last exception motivates the exclusion of 'uncorrected' fluxes in the following, as we do not need to keep both GRIP flux-based curves, once we have ascertained that the 'corrected fluxes' satisfy our χ^2 test, possibly because of the improved chronological spacing of the samples.

In the case of all Greenlandic concentration-based curves, because of very low p-values of the χ^2 ($p<<0.01$), we decided to set the modelled-$\Delta^{14}C$ uncertainty to the RMSD of each dataset, which are between 12 and 21‰, less than the amplitude of the matched features. We repeat the wiggle-matching for these cases and base our offset estimation on the full Greenland dataset with the uncertainty method outlined above.



~~*Figure 7 Carbon cycle modelled $\Delta^{14}C$ compared to measured Hulu Cave data and climatic data before synchronization.*~~
~~*(a) Measured Hulu Cave used for synchronization (H82: Southon et al., 2012; MSD: Cheng et al., 2018). (b) Modelled $\Delta^{14}C$ from Greenland ^{10}Be concentrations (NorthGRIP, GRIP and GISP2 datasets). The effect of the flux correction of GRIP is mostly visible around the G2B event, where the corrected data shows a better agreement with the other cores. (d) The climatic data from NorthGRIP ($\delta^{18}O$ and calcium) contain the signature of the dust peak and the GIs. (e) $\Delta^{14}C$ modelled from WDC ^{10}Be fluxes and concentrations. (f) Climatic data from the WDC core ($\delta^{18}O$ and CH_4).*~~

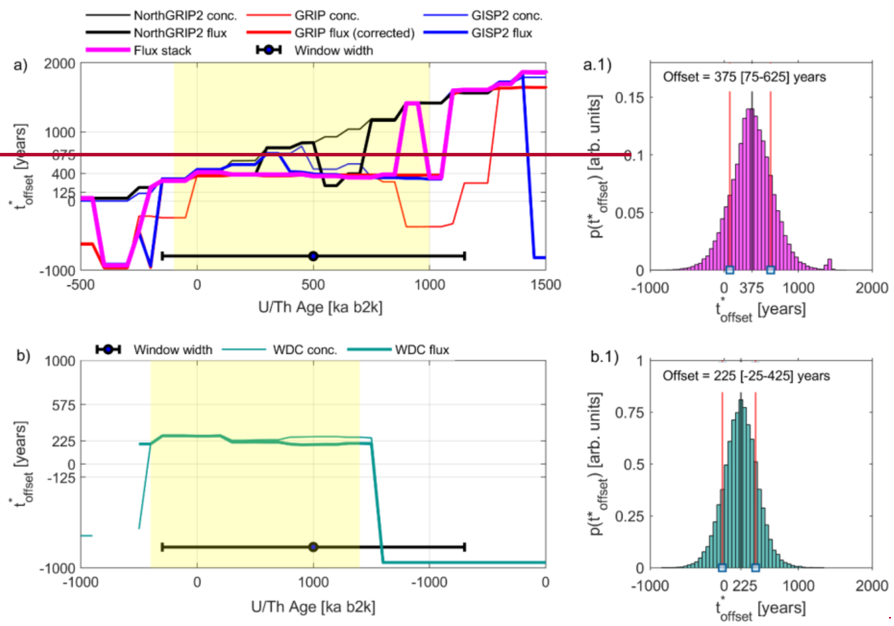
We therefore repeat the averaging of the ~~timescale~~time-scale offset (~~fig. 8~~Fig. 9) between the MSD dataset and the GICC05 ensemble and we apply the Monte-Carlo iteration outlined in the Methods, between 21 and 22.1 ka b2k (the range highlighted in yellow in ~~fig. 8a~~Fig. 9a, where edge effects can be avoided). We obtain the offset estimate of 375 years~~ and~~, a 68% confidence interval from 75 to 625 years, and a 95 % confidence interval from -300 to 1000 years. One assumption behind this approach is that the offset should only change slowly with time, which is supported by the flatness of most curves in ~~fig. 8~~Fig. 9 (with some exception by the concentration-based offsets of NorthGRIP and GISP2).

For the WDC datasets, both flux- and concentration-based curves were shifted by 225 years. Upon performing the χ^2 test against the Hulu Cave data, the p-values were within the 0.01-0.99 tolerance interval, hence the uncertainty of the modelled $\Delta^{14}C$ data did not need to be re-evaluated. We compute the average ~~timescale~~time-scale offset (~~fig. 8~~Fig. 9) between the MSD dataset and the WD2014 ensemble of flux and concentration-based curves and we apply the Monte-Carlo iteration outlined in the Methods, between 20.8 and 21.75 ka b2k (the range highlighted in yellow in ~~fig. 8b~~Fig. 9b). We obtain the offset estimate of 225 years~~ and~~, a 68% confidence interval from -25 to 425 years, and a 95 % confidence interval from -350 to 750 years.

The difference between the ice-core offsets represents another indirect estimate of the offset between the ~~polar timescales~~Greenland and Antarctic chronologies, which in this way is determined to be 150 years, close to the offset of 125±33 years directly obtained by synchronizing the G2B event in the ^{10}Be fluxes.

Our result for GICC05 is smaller than the 550-year offset obtained by Adolphi et al. (2018) but fully consistent within the uncertainties of their estimate (95% probability interval: 215 – 670 years). For the H82 dataset, our analysis leads to offsets of about 500 ± 200 and 220 ± 500 years, for GICC05 and WD2014, respectively, which more closely reproduces the Adolphi et al. (2018) finding.

~~In the following, we will assume the U/Th ages to be correct and shift the ice-core time scales accordingly, although, part of the offset could also be attributed to U/Th being too old because of measurement uncertainties or age modelling issues (Corrick et al., 2020).~~



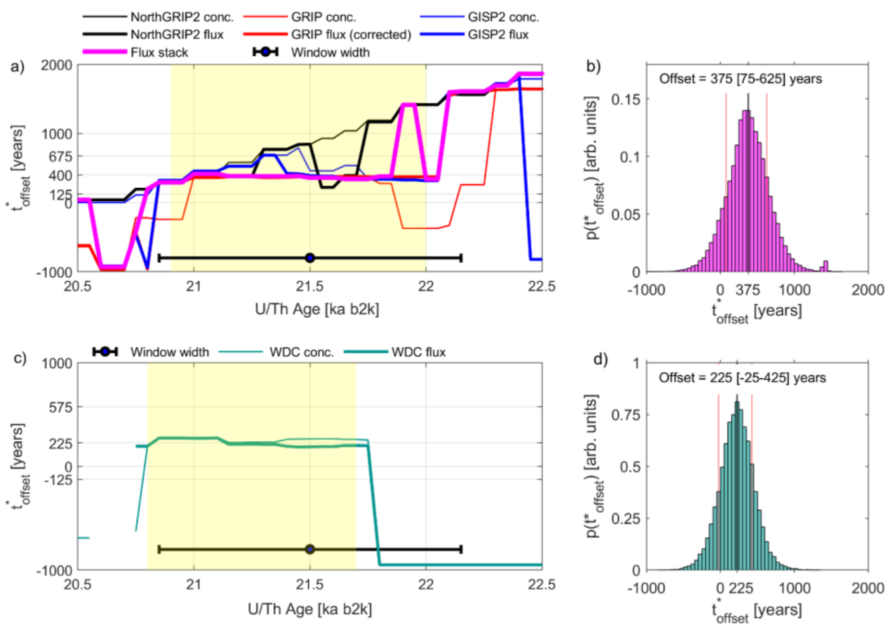


Figure ~~8~~ 9 Wiggle-matching ~~result~~and Monte-Carlo results around the G2B event for Greenland (a, b) and the WDC core (b~~). The~~, c). Matlab code is provided in the Supplement~~ allows to reproduce this figure by running the function 'wiggle_matching_sinnl_et_al_2022'. The time~~ . The curves in the left panels are the time-scale offset ~~function~~ $t^*_{offset}(\vec{W})$ of each ice-core dataset~~ was calculated as the mode of the underlying 2-dim probability density function, estimated by the algorithm (Adolphi et al., 2016). The window width (horizontal black bar) is highlighted to show that each data point represents the data comparison within the windows~~ ~~$\vec{W}$. Across the~~ . The yellow area delimits the time intervals ~~highlighted in yellow, the individual ice-core datasets agree about the offset~~ for ~~each ice-core timescale. Outside these intervals,~~averaging (T1 and T2), outside of which curve instability is caused by the lack of appropriate matching features in the data. ~~(a) The~~ (b) Distribution of the Greenland offset~~ curve, calculated after enlarging the $\Delta^{14}C$ uncertainties of the concentration-based data. (a.1) Monte-Carlo study of the Greenland offset uncertainty. Sampling each offset curve's probability envelope within the yellow interval, as outlined in the Methods, returns a histogram representing the probability of the average offset.~~. From the histogram, we compute the 68% confidence interval of the offset (red lines). ~~(b) For~~ (d) Distribution of the WDC~~, the offset curve of the concentrations and fluxes are the only available datasets for the calculation. The average offset within the interval (yellow) is computed similarly as for GICC05~~ offset, with ~~the Monte-Carlo histogram shown in panel (b.2).~~68% confidence bars.

4 Discussion

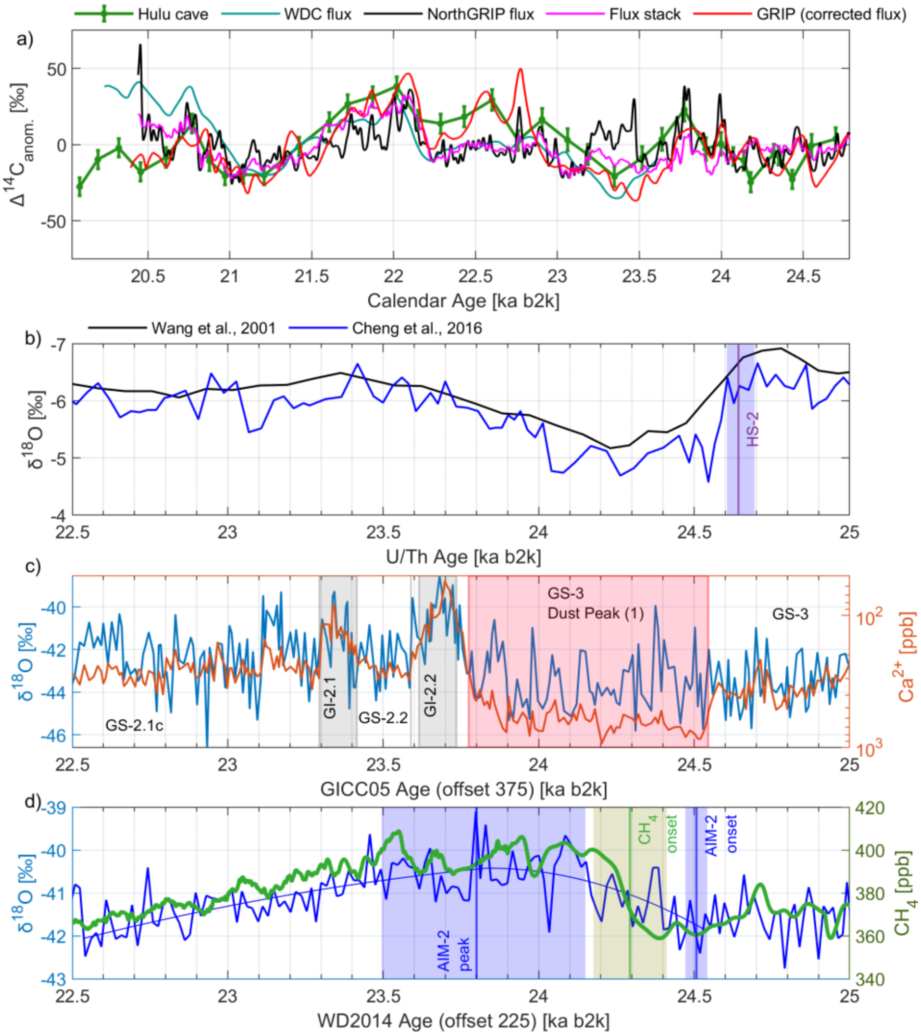
4.1 Climate compared after synchronization

In ~~figure 9~~Figure 10 we show the radionuclide and climatic data after the synchronization proposed in section 3.3.

To reconstruct the sequence of events during the LGM, we need to estimate the onset and termination of HS-2. We suggest the onset of HS-2 to be defined by the $\delta^{18}O_{calcite}$ slope in Asian speleothems (~~fig. 9b)~~ (Fig. 10b; Cheng et al., 2021). Here, we identify the onset of the HS-2 signal at Hulu Cave by the same methods used for the AIM-2, combining the WDC_breakpoint function with a Monte Carlo iteration. By averaging the onset in the two $\delta^{18}O_{calcite}$ onsets at Hulu Cave, one for each dataset, we obtain the HS-2 onset to be 24.71±0.04 ka b2k. This is earlier than, albeit compatible within 2 σ with the definition by Li et al. (2021) of an HS-2 onset at around 24.48±0.08 ka b2k, based on another speleothem (Furong, China). The Hulu onset is also earlier than the Cherrapunji HS-2 onset at 24.47±0.04 ka b2k (Dong et al., 2022).

~~The events are listed from oldest to youngest (although overlap may occur because of the uncertainty bounds). The Greenlandic sequence of GIs and GSs is fixed by the NorthGRIP layer count, therefore the uncertainties of these events are correlated and their order fixed, but they can be shifted as a group. The dating uncertainty is dominated by the MCE, which is about 600 years at this age. The shape of the AIM-2 and CH_4 at~~

~~Event~~	~~Archive~~	~~Age according to wiggle-matching synchronization (y b2k $^{+\sigma}_{-\sigma}$)~~	~~Original Calendar age (y b2k)~~
~~HS-2 onset~~	~~Hulu $\delta^{18}O$~~	~~24710 ± 40~~	~~24710 ± 40~~
~~Dust onset~~	~~NorthGRIP Ca^{2+}~~	~~24525^{+250}_{-300}~~	~~24150±10~~
~~AIM-2 warming onset~~	~~$\delta^{18}O$ WDC~~	~~24497^{+200}_{-250}~~	~~24272 ± 35~~



~~AIM-2 cooling onset~~	~~δ^{18}O WDC~~	~~23825^{+360}_{-390}~~	~~23600±300~~
~~Dust termination~~	~~NorthGRIP Ca~~	~~23755^{+250}_{-300}~~	~~23380±20~~
~~Start of GI-2.2~~	~~δ^{18}O Greenland~~	~~23715^{+250}_{-300}~~	~~23340±20~~
~~Start of GS-2.2~~	~~δ^{18}O Greenland~~	~~23595^{+250}_{-300}~~	~~23220±20~~
~~Start of GI-2.1~~	~~δ^{18}O Greenland~~	~~23395^{+250}_{-300}~~	~~23020±20~~
~~Start of GS-2.1c~~	~~δ^{18}O Greenland~~	~~23275^{+250}_{-300}~~	~~22900±20~~

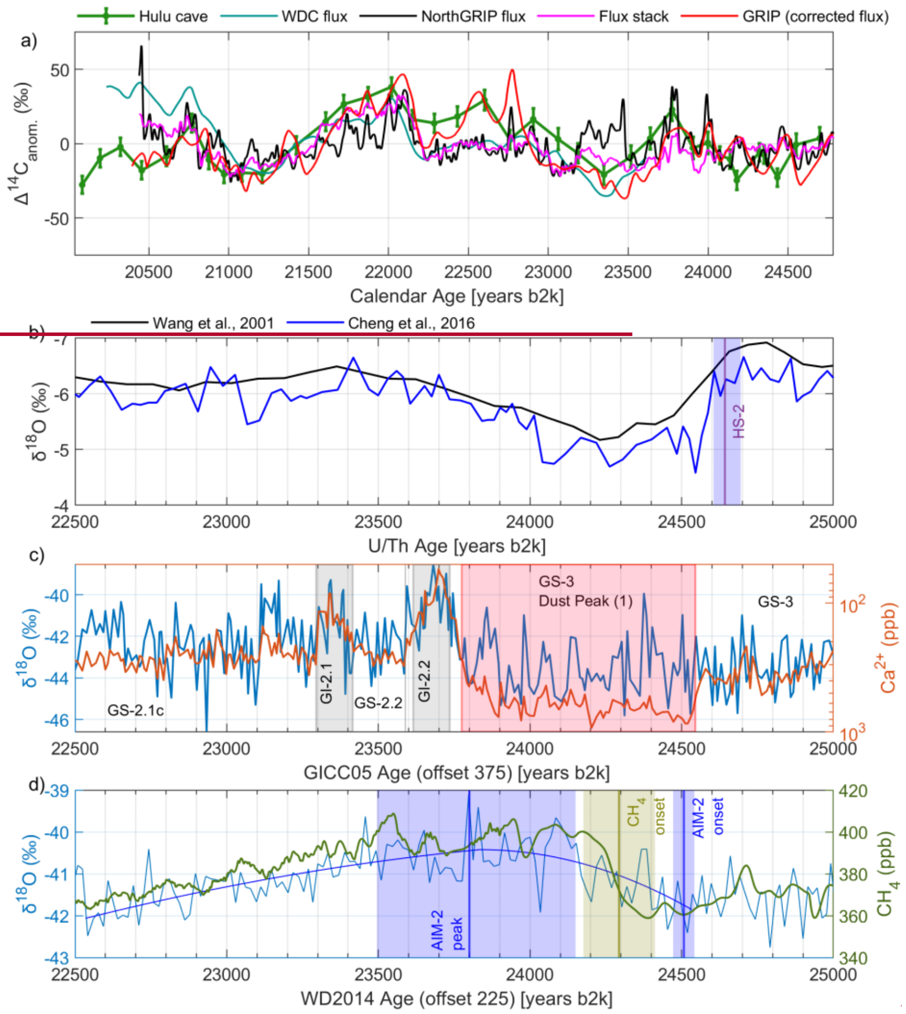


Figure ~~9~~ 10 Radionuclide and climatic data after the synchronization.

The Greenlandic records were shifted by 375 years towards older ages; the WDC records were shifted by 225 years towards older ages. (a) Radionuclide data used for wiggle matching plotted on the synchronized time scales.

(b) Hulu-cave isotopes (reversed y-axis) highlight the suggested onset of HS-2 (Cheng et al., 2016). A lower-resolution dataset is also plotted for comparison (Wang et al., 2001). The termination of HS-2 is not defined, as multiple possibilities arise by comparing the two datasets. (c) Greenlandic proxy data on the shifted GICC05. (d) WDC proxy data on the shifted WD2014.

~~*(c) Greenlandic proxy data on the shifted GICC05 time scale. Calcium is presented on an inverted log-axis to better compare to the Hulu record.*~~

~~*(d) The AIM-2 period aligns both with the dust peak and the HS-2 in Hulu $\delta^{18}O_{calcite}$. The onset of the methane increase occurs about 210 years after the onset of the AIM-2 warming, although we have to consider uncertainties in the gas-age reconstruction. The AIM-2 peak occurs immediately before the GI-2.2, with a wide uncertainty band related to the unclear shape of the AIM-2.*~~

With the time~~-~~-scale correction for WD2014, the onset of the AIM-2 warming occurs synchronously with the HS-2 onset, within the uncertainties determined both by the wiggle matching and by the AIM-2 fit~~, although a~~. A delay ~~would~~could be expected given the centennial-scale response of Antarctic climate to Northern Hemisphere changes (Pedro et al., 2018; Svensson et al., 2020).

The AIM-2 maximum (i.e. the peak of ~~$\delta^{18}O$~~$\delta^{18}O_{ice}$ in Antarctica) was here defined by considering only the data from WDC. By visual inspection, the peak of the AIM-2 signal could either be identified as two marked ~~$\delta^{18}O$~~$\delta^{18}O_{ice}$ peaks at around 23.8 ka b2k (shifted WD2014 age) or as a prolonged plateau between 23.5 and 24.2 ka b2k. We observe that the 23.8 ka b2k ~~$\delta^{18}O$~~ peak in WDC occurs together with or slightly before GI-2.2, which is a similar result one obtains by merely synchronizing the data using the G2B event tie point (~~fig. 6). As the shape of the AIM-2 signal differs across Antarctic ice cores (Veres et al., 2013), a comparative study of other ice cores would improve this result. We cannot provide an Antarctic comparison in this context, as the WD2014 chronology does not currently apply to other ice cores, hence an updated Antarctic synchronization across AIM-2 would be required.~~Fig. 7).

The shape of the AIM-2 signal differs across Antarctic ice cores (EPICA members, 2006), possibly because of registration differences across the continent. A comparative study of all Antarctic deep ice cores would improve our identification of the AIM-2, but the WD2014 time scale was only tied to few ice cores so far. At least, we can compare the recent South Pole ice core (SPICE), chronologically tied to the WD2014 (Winski et al., 2021). The shape of the AIM-2 in SPICE is similar to WDC (Fig. S5). While the onset occurs at the same age, the maximum is reached earlier in SPICE, at 24.1 ka b2k in the shifted WD2014 ages. This indicates that careful consideration of the geographic factors of the Antarctic continent are needed for an ultimate climate reconstruction of the AIM-2. In the meantime, we suggest that caution is exercised when interpreting minor isotopic signals.

The methane increase in WDC is registered ~210 years after the AIM-2 warming started and the HS-2 onset. This supports the theory by Rhodes et al. (2015) of an increased southern biogenic methane production as a delayed response to extreme northern-hemispheric cooling. The high methane levels appear to last between 460 and more than 1000 years, depending on how one defines the end of the methane plateau.

4.2 Discussion on causes of the offset for GICC05

In recent work by He et al. (2021~~)~~), it was suggested that, although not reflected by Greenlandic water isotope records, a shutdown of the AMOC likely occurred during Heinrich Stadial 1, which was stronger than the AMOC shutdown of a 'regular' GS, producing a period of extreme winter cooling. They concluded that, rather than Greenland not experiencing any additional cooling during HS-1 (as proposed by Landais et al., 2018), the imprint of the cooling in the $\delta^{18}O_{ice}$ signal was cancelled by the effect of having less winter snow and by an increase of the $\delta^{18}O_{ice}$ level of summer snow, due to the first warming of the deglaciation. The flatness in the water isotope data is, in their analysis, the result of these counteracting effects of Greenland cooling: the HS-1-specific reduction in winter precipitation and the ^{18}O-enriched summer precipitation.

As ~~much~~far as the GICC05 layers are concerned, He et al. (2021) model a possible scenario over Greenland with a drastic decrease of winter precipitation by about 50% at the onset of HS-1. On the other hand, they predict the summer precipitation to increase steadily as the deglaciation progresses, which is largely unaffected by the AMOC shutdown. This could potentially have produced thin and irregular layers at the onset of HS-1, with the summer part of the layer gradually increasing over time. Acknowledging the 125-~~years~~year offset between GICC05 and WD2014 at 22 ka b2k, and the 375 ~~years~~-year offset between GICC05 and the U/Th ~~timescale~~chronology, we proceed by discussing where the offset could have originated. According to Adolphi et al. (2018), the transfer function from GICC05 to IntCal is near to zero at 13 ka b2k, the closest tie-point younger than the G2B event. Occurring between these two time-horizons, the HS-1 period lasted from about 18 to 14.7 ka b2k, and is most often regarded as ending at the onset of GI-1. The authors of GICC05 retained the prior subdivision of the corresponding stadial, GS-2.1, in 3 sections (termed a, b, and c) of which GS-2.1a starts at 17.48 ka b2k and ends at the onset of GI-1 (Björck et al., 1998; Rasmussen et al., 2014). The onset of GS-2.1a was originally defined by a 2nd-order water isotope dip but occurs synchronously with an increase in the Greenland ice-core calcium profiles at roughly 17.6 – 17.4 ka b2k.

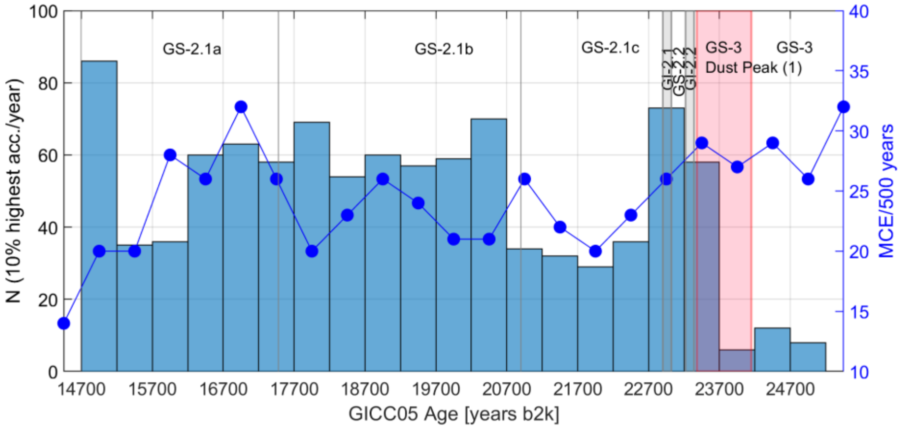


Figure ~~10~~11 Highest accumulation years (histogram) and MCE (dark blue curve) based on the GICC05 ~~timescale~~time scale, computed on the same 500-years intervals.

~~The MCE appears rather constant over GS-2, except for some higher values at ~16.7 ka b2k. On the other hand, the thickest layers (>0.09 m/year; corresponding to the 10% thickest layers in the entire period) are located preferentially across GS-2.1b and 2.1a, with a sharp onset at 20.7 ka b2k. The first bin (14.7-15.2 ka b2k) also contains relatively thick layers, possibly an effect of the onset of GI-1. As expected, the thickest layers are often found in the brief interstadials, while over the dust peak the frequency of high-accumulation years is much lower. The thickest layers across GS-2.1b and 2.1a have as many high-accumulation years as interstadial periods, which is surprising given that this period was very cold and had very low annual precipitation (Kindler et al., 2014).~~

If GICC05 missed a large number of annual layers across HS-1, the age of the onset of GS-2.1a would move towards older ages by a similar amount. We speculate that the onset of GS-2.1a could correspond to the onset of HE-1 (and thus the HS-1 period) and that calcium could be used as a signature of the change of atmospheric circulation influencing dust transport to Greenland during HS-1 as well as HS-2.

Part of the under-counted layers could have contributed to the ~~MCE. The MCE~~ uncertainty of GICC05. This was assessed by identifying so-called ‘uncertain annual layers’, that is ~~derived from the number of~~, features that could be annual layers but are not clearly identified as such~~; it does not properly account for layers not resolved by~~. Each uncertain layer contributed to the chronology and to the ~~data.~~Maximum Counting Error (MCE) as 0.5±0.5 years (Rasmussen et al., 2006). Within GS-2.1a, the MCE grows by ~150 years, meaning that 300 uncertain layers were ~~counted~~identified as 0.5±0.5 years. If we assume that all the uncertain annual layers in GS-2.1a were annual layers, an additional 150 years would appear in the chronology~~, and the MCE would decrease by 150 years.~~ across this section.

As we have seen, according to He et al. (2021), a large number of very thin annual layers is a plausible scenario at least in the first part of HS-1, until the increasing summer precipitation alleviated the problem. As these layers are possibly unresolvable by the data, they would not be counted as uncertain layers, hence the MCE is probably not faithfully representing the full GICC05 uncertainty in sections where layer thicknesses are very small compared to the resolution of the data.

In addition to mis-assigned uncertain layers and layers missed altogether due to marginal data resolution, we here propose another explanation of what could have caused the under-count of GICC05 layers. The layers that are on the higher end of the thickness distribution, even if they were classified as 'certain', might indicate where very low or absent winter precipitation could have made multiple annual layers appear as one.

~~We~~To identify multiple layers, we highlight the ~~10%~~ years with highest accumulation in ~~fig. 10~~Fig. 11, where we also show the MCE, both per 500-year interval. The accumulation threshold is set at 0.09 m/year by integrating the highest 10% of the ~~empirical~~accumulation distribution ~~of the accumulation~~ between 14.7 and 25 ka b2k. ~~In the accumulation~~The histogram ~~of fig. 10, values are consistently high over the entire~~in Fig. 11 thus shows the number of very thick layers in each time-bin.

The thickest layers across GS-2.1b~~, with~~ and 2.1a have as many high-accumulation years ~~occurring as frequent~~ as ~~in the short interstadials.~~ interstadial periods, which is surprising given that this period was very cold and had very low annual precipitation (Kindler et al., 2014).

For the MCE,~~ it appears that~~ observers of GICC05 encountered more issues at the onset of GS-2.1a, whereas the MCE stayed constant over the rest of GS-2.1. Hence, we suggest that a dating bias could have accumulated across GS-2.1a and GS-2.1b due to increased difficulty in identifying annual layers, caused by a weakening of the annual signal due to reduced amounts of winter snow. However, we find no clear evidence that this phenomenon occurs at the onset at HS-1 (17.7 ka b2k).

Recently, Dong et al. (2022) aligned the climate signature of HS-2 in the Cherrapunji speleothem to the calcium peak in Greenland, which was in turn aligned to Antarctica by one volcanic tie point at 24.6 ka b2k (match point no. 15 in Fig. 7; Svensson et al., 2020). Here, the GICC05 is 80 years older than WD2014, i.e., the offset has the opposite sign than at the G2B event. To produce such an inversion of the offset, an about 200 more layers had to be counted in NorthGRIP than in WDC over the interval between G2B and no. 15. However, a uniform stretching of GICC05 with respect to

WD2014 is not advised, as we cannot assume that the offset is evenly distributed. A closer comparison via more bipolar volcanic tie points is needed to assess the time-varying nature of the offset.

4.3 Accuracy of the WD2014 and Hulu time scales during the LGM

For the WD2014 time scale, the offset can be compared to the uncertainties quoted for the ~~time scale~~chronology by Sigl et al. (2016). Between 20 and 25 ka b2k, 1σ uncertainties are between 100 and 125 years, which are smaller than the offset we find. This suggests that the authors of WD2014 may have underestimated their uncertainties, similar to GICC05, and that the layer counting in the glacial was also more challenging than acknowledged at first, but the two estimates are nonetheless within 2σ. However, the gradual warming ~~observed~~recorded in Antarctica between 19 and 15 ka b2k (Pedro et al., 2011) likely did not lead to fast changes in the annual layers' expression. The WD2014 issues may thus be mostly related to the low resolution of the data used for counting between 15 and 26 ka b2k (Sigl et al., 2016), which may have limited the accuracy of the ~~time scale~~chronology.
As a final remark, we have assumed here that the U/Th ages are correct and we have shifted the ice-core chronologies accordingly, on the basis of the challenging layer identification in the ice cores and of the small absolute uncertainties on the U/Th ages. However, the Hulu time scale could in principle also carry some error. For example, Corrick et al. (2020) speculate on possible issues with the Hulu chronology caused by interpolation and growth-rate modelling related to differences in the sample positioning for U/Th or $\delta^{18}O_{calcite}$ measurements. As we have seen above, there are at least two instances of Asian speleothems with a younger onset of HS-2 (Li et al. 2021; Dong et al. 2022), which agree with each other within their combined uncertainties. Apart from possible climatic delays across the speleothems, this could indicate issues in the absolute Hulu chronology, which would imply a smaller offset from the polar chronologies. Since, to our knowledge, there is no available radionuclide data for the Cherrapunji or Furong speleothem, our methodology relies on the Hulu $^{14}C_{calcite}$ record. It is crucial to resolve these differences in the speleothem chronologies as Hulu also forms the backbone of the ^{14}C-dating IntCal20 curve.

5 Conclusion

In this study, we have presented two new ^{10}Be datasets that allow both a bipolar synchronization and a comparison of polar climate to the Hulu speleothem archive during the 20-25 ka b2k period. The new NorthGRIP ^{10}Be dataset suggests a dating offset between GICC05 and U/Th time scales of

375^{+250}_{-300} years, which is less than the 550-year estimate by Adolphi et al. (2018), albeit consistent within uncertainties. Likewise, an offset was found between U/Th Hulu Cave dates and the WD2014 time scale of 225^{+200}_{-250} years, which is supported by the visual alignment of the datasets.

The rather large uncertainties on the time-scale offsets could be reduced in future studies. In terms of a bipolar match, a tight volcanic synchronization across the LGM, with the aid of our ^{10}Be match, would certainly link the two ice sheets with the most precision.

The issue of large uncertainties in the wiggle matching lies in resolution problems of the $^{14}C_{calcite}$, in the low signal-to-noise ratio of the G2B event, and in the carbon-cycle model likely not capturing every aspect of the LGM climate. The combination of these factors produces multiple possible alignments of our datasets. Therefore, higher-resolution ^{14}C data from absolutely-dated tree rings or from another speleothem, with a lower dead carbon fraction than Hulu, would improve our chances to find reliable signal structures to match. From the ice-core side, high-resolution ^{10}Be from other ice cores would further limit the archive-specific and climate noise contributions.

In terms of the sequence of events between 20 and 25 ka b2k, ~~our time-scale offset does not change their order. However, the onset of~~ the Greenlandic dust peak ~~moved~~shifted to be roughly synchronous with the signal in the Hulu speleothem ~~that has been~~ linked to the HS-2 onset. ~~Observing the~~ The onset of AIM-2 ~~occurring~~occurs relatively near to the dust and Hulu signals, ~~we support~~supporting the hypothesis that the AIM-2 warming is related to an AMOC shutdown during HS-2. The CH_4 increase onset in Antarctica also moved closer to the HS-2 onset, supporting the theory by Rhodes et al. (2015) of a long overlap of the methane plateau with the HS periods.

The termination of the Greenlandic dust peak and the almost synchronous onset of GI-2.2 were brought closer to the AIM-2 peak. We observe that the AIM-2 maximum, at least within WDC ~~$\delta^{18}O$~~$\delta^{18}O_{ice}$ data, occurs together or before the GI-2.2 onset. This may indicate an exception to the average delay of 122±24 years found within other GI-AIM pairs, where the GI generally occurs before the AIM breakpoint (Svensson et al., 2020). Another piece of evidence suggests a deviation from the normal bipolar seesaw mechanism: It seems likely that the very brief GIs are not the primary cause of the sustained Antarctic gradual cooling taking place during the first millennia of the GS-2.1, but could be an indication that the AMOC resumed or gained more strength after GI-2.1 and into GS-2.1.

We investigated several possible reasons why the annual-layer counters of GICC05 could have missed annual layers across GS-2.1. We find that the MCE of GICC05 alone cannot account for all missing layers in the GS-2.1, but that either very thin annual layers or missing winter precipitation could have made it difficult, if not impossible, to identify the thinnest layers with the available data, and thus made the MCE estimation less robust.

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~~6~~7 Data and Supplement

^{10}Be data of NorthGRIP and WDC are made available in the excel documents provided with the Supplement of this paper. Moreover, Matlab codes for reproducing ~~fig. 8~~Fig. 9 and ~~fig. 9a~~Fig. 10a are provided in the Supplement, as well as the supplementary figures.

~~7~~8 Methods Appendix

~~7.1~~8.1 WDC measurements

Ice samples from the WAIS Divide ice core were stored and processed at the NSF Ice Core Facility in Lakewood, Colorado. Of each 1-m section from 2453-2599 m depth, a thin slice (DD-16) with a cross-section of ~2 cm^2 from the outside of the core was cut for ^{10}Be analysis and shipped in frozen form to Purdue University. For each ^{10}Be sample, two consecutive core sections were combined, yielding a total mass of 350-400 g per sample. Ice samples were weighed, melted, and acidified with a solution containing ~0.18 mg of Be carrier. The samples were passed through a 30 µm Millipore filter and loaded onto a 3 ml cation exchange column (Dowex 50WX8) from which the Be fraction was eluted and purified following established procedures (Finkel and Nishiizumi 1997; Woodruff et al. 2013). The Be fraction was precipitated as $Be(OH)_2$, transferred to a small quartz vial and heated in a tube furnace at 850 ºC. The BeO was mixed with Nb powder (Alfa Aesar, -325 mesh, Puratronic, 99.99%) and pressed into a stainless-steel cathode. The ^{10}Be/^{9}Be ratios of samples and blanks were measured by accelerator mass spectrometry at Purdue's PRIME laboratory (Sharma et al. 2000), relative to well-documented ^{10}Be/Be AMS standards (Nishiizumi et al. 2007). The measured ^{10}Be/Be ratios of the samples were corrected for an average blank ^{10}Be/^{9}Be ratio of $(10 \pm 3) \times 10^{-15}$ which corresponds to typical blank corrections of 1-2% of the measured values. The blank-corrected ^{10}Be/^{9}Be ratios, combined with the sample mass and amount of Be carrier added yield ^{10}Be concentrations ranging from 3.1 to 5.4×10^4 atoms/g (~~fig. 1d~~Fig. 2d). Typical uncertainties (1σ) in the measured ^{10}Be concentrations are 1.5-3.5%. The measured ^{10}Be concentrations were not corrected for radioactive decay, which would increase all values by 1.0-1.2%.

~~7.2~~8.2 NorthGRIP measurements

~~7.2.1~~8.2.1 Ice cutting

At present, the required minimum weight for each ^{10}Be measurement is around 120 g. About half of the samples, in alternate order, had previously been cut in the shape sticks for gas measurements (section area of 3.5x3.5 cm^2), with a weight of around 600 g per bag (55 cm). The "gas sticks" were thus cut into 4 parts, resulting in pieces of 13.75 cm, corresponding to a resolution of about 7.5 years. The other half of the samples were cut from the archive piece to have a section area of ~2x3 cm^2.

Each bag of these was then cut into two parts, hence each of such pieces weighs around 180 g and corresponds to about 14 years resolution. The campaign was performed with the necessity of keeping the total number of samples at around 320 for cost reasons. Since the total number of cut ice samples was 470, the pieces were selected to alternate between adjacent samples at 7.5 years resolution, to minimize the age gaps; the resulting ^{10}Be data consequently shows some data gaps.

~~7.2.2~~8.2.2 Sample preparation and AMS Measurement (ETHZ)

Ice samples of 150-180 g were weighed and a solution containing 0.15 mg Be carrier (^{9}Be) was added, as well as 1 mg of Cl carrier. The samples were melted, not filtered, and run through cation exchange columns, from which the Be fraction was extracted. The melt water was further run through anion exchange columns to retain the chlorine content and stored for future ^{36}Cl measurement. The resulting $Be(OH)_2$ was heated in steps to 850° to obtain BeO and mixed with Niobium powder. Five blank samples of Milli-Q water were also measured for background assessment. More details about the most recent preparation protocol which is used at the Lund Laboratory can be found in Nguyen et al. (2021).

The $^{10}Be/^{9}Be$ ratios of the samples were measured in July 2020 at the AMS facilities at ETH in Zurich (Christl et al., 2013), where a total of 322 measurements were performed. Measured $^{10}Be/^{9}Be$ ratios were normalized to the ETH Zurich in house standards S2007N and S2010N which in turn have been calibrated relative to primary standards provided by K. Nishiizumi (2007). An average blank correction of $^{10}Be/^{9}Be = 1.3 \pm 0.3\ 10^{-14}$ was applied to correct for ^{10}Be introduced with the Be-carrier. The blank correction corresponds to 2-3% of the measured ^{10}Be ratios. Calculated ^{10}Be concentrations range between 1.8 and 8.7 10^4 atoms/g and were decay corrected using a half-life of $1.387 \pm 0.016\ 10^6$ years, which, at these ages, produces a signal increase of only about 1%. The final uncertainties of the ^{10}Be concentrations were between 1 and 10 %.

~~7.3~~8.3 From concentrations to fluxes

The concentration to flux conversion, for ice cores, is obtained by calculating $\varphi = \rho_{ice}\gamma\alpha$, where φ is the flux, γ is the ^{10}Be concentration, α is the accumulation rate, and ρ_{ice} is the ice density (0.917 g/cm^3).

~~7.4~~8.4 Wiggle-matching algorithm settings

Thanks to the higher resolution of ice-core data with respect to Hulu data and thanks to the carbon-cycle model producing a continuous output, we approximate the ice-core modelled $\Delta^{14}C$ input as a continuous function of time, i.e. we resample it annually by linear interpolation. The second input, the Hulu-cave data, is a collection of discrete data points. The $\Delta^{14}C$ uncertainties are about 6‰ for the Hulu Cave data and 5‰ for the ice-core data. The datasets are detrended within each observation window before computing the probabilities. In particular, the ice-core $\Delta^{14}C$ is first shifted according to the scanned time offset, then detrended with respect to the observation window, and resampled to the Hulu-cave sampling points.

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