

We thank the two reviewers for their time and valuable comments that were taken into account as detailed in the following.

Comments from Christo Buizert.

- Could you speculate on the physical mechanisms for the decoupling? In other words, why are H-events not recorded in Greenland temperature records? My sense has always been that Greenland temperatures are “saturated”. Once the winter sea ice edge is far enough south (say 45N), driving it further south by additional AMOC weakening will not cool Greenland any further. Other explanations are also possible of course. Since the decoupling is the main topic of the paper, it would be appropriate to have some discussion on what could cause such a decoupling.

>> We propose to add the following text: “Uniformly cold conditions are generally observed during Heinrich Stadials of the last glacial period with temperature and $\delta^{18}\text{O}$ levels that are not significantly lower than temperature levels observed during Greenland Stadials (Kindler et al., 2014; Guillevic et al., 2014). Because Greenland is surrounded by large sea ice extent during Greenland Stadials and Heinrich Stadials (Hoff et al., 2016), an explanation may be that central Greenland temperatures are saturated during cold periods so that AMOC modifications occurring south of the sea ice edge are not influencing significantly Greenland temperatures.”

- Line 63-64: The 16.2 ka event is also seen very beautifully in Cariaco basin reflectance (Deplazes et al., 2013). Since Cariaco is such an iconic record, it would be worth mentioning this (or even showing it in one of the figures).

>> We propose to add this reference in the following sentence: “At low latitudes, an ITCZ shift at 16.2 ka is clearly expressed through a weak monsoon interval in East Asian speleothem records and through change in hydrology in the low-latitude Pacific region, Cariaco Basin and Brazil (Partin et al., 2007; Deplazes et al., 2013; Russell et al., 2014; Strikis et al., 2015).”

- Another paper that should be referenced here is (Zhang et al., 2016). They argue for stronger links to Antarctica than to Greenland, as also proposed here. The decoupling between Greenland and mid-latitude hydrology had been noted explicitly by others also, which could be acknowledged more clearly (Rhodes et al., 2015; Zhang et al., 2016; Zhang et al., 2014).

>> The reference to the work of Zhang et al. (2016) has been added (see below) and we propose to modify the present conclusion by adding: “These new measurements hence confirm the previous studies of Zhang et al. (2014, 2016) and Rhodes et al. (2015).”

- The motivation on line 70-76 for using ice cores is not very strong, in my view. Since there is no H-event signal in Greenland temperature records anyway, there is no benefit in having a hydrological record on the same time scale! All the other records (CH₄ and CO₂ from Antarctica, sediments and speleothems) are on independent chronologies. Why not just state that ice core $\delta^{18}\text{O}$ and $\delta^{17}\text{O}$ can provide additional evidence to supplement what we know already from sediments and speleothems? That is a strong enough motivation in my view.

>> What we want to compare is the Greenland temperature record (high latitude) with lower latitude hydrological changes on a common timescale. In this sense, measuring $\delta^{18}\text{O}$ (+ $\delta^{15}\text{N}$ of air), $\delta^{17}\text{O}$ -excess and $\delta^{18}\text{O}$ -excess on the same ice core has really a sense and it is much better than comparing speleothem or sediment records with a Greenland ice core record. Such common measurements become increasingly important when going back in

time as larger uncertainties become associated to the record chronologies. We would thus like to keep this motivation for our work.

- Line 81: Please just call it Maximum Counting Error, and leave out the 1 sigma. I know that it is often advertised in terms of standard deviations, but I think it's incorrect. A 200 yr MCE means that GICC05 encountered 400 uncertain layers (each counted as half a year, representing a 50-50 chance the layer is real). So the 200 yr error is an extreme case where 400 coin flips all landed face-up. Not exactly a 1 sigma event.

>> Indeed, this was a mistake to call it a 1 sigma uncertainty. It will be corrected.

Line 86: "using a PICARRO laser cavity: : ." (change word order)

>> done

Line 101: INSTAAR (typo)

>> done

Line 156: "reflect" instead of "reflects"

>> done

Line 194: "the same timescale". Same timescale as what? As the Greenland records? Did you plot the Rhodes et al. record on its original timescale, or a different one? The caption to Fig. 2 suggests records are on the GICC05/AICC2012 time scale. How did you convert the WAIS Divide records to GICC05?

>> Actually, we kept the original timescales for the two cores but did a translation since the WAIS timescale (WD2014) is referred to year 1950 and GICC05 to year 2000. So, the conversion is only to refer both timescales to year 2000. We propose to remove "presented on the same timescale" to prevent any confusion and say that the WAIS results are presented on the WP2014 timescale referred to year 2000.

Line 195: "hypothesized to reflect" instead of "understood to reflect".

>> done

Line 199: ": : carbon fluxes and/or enhanced air-sea: : ." Both could be true.

>> Yes, that was the aim when writing "and/or"

Line 247: A southward shift of source regions is also what I would expect. This could explain the apparent SST increase of the source. However, increasing both RH and SST is hard to do through meridional shifts in atmospheric circulation. SST decreases with latitude, but RH increases. So at lower latitude, RH should be lower, actually. Any thoughts on what circulation change could cause both signatures?

>> A possibility here would be that we have mixed contributions of continental and marine sources for precipitation in Greenland at that period in agreement with the modeling study of Werner et al. (2001). Continental sources are from North America and we have a transition from a big dry to a big wet in that region at 16.2 ka. A contribution from the big wet North America from 16.2 ka would thus explain an increase of relative humidity from that period.

This idea was probably not expressed clearly enough in the present manuscript and we now propose the following text: "The Greenland signal of source humidity increase may at least partly explained by wetter conditions in the continental North America evaporative source regions, which are known to partly affect Greenland moisture today in addition to the main source in Northern Atlantic (Werner et al., 2001; Langen and Vinther, 2009)."

Line 254: how does the "big wet" transition fit in dynamically? Presumably the storm tracks and polar jet stream over N-America shift southward (Asmerom et al., 2010)?

>> This is indeed a possibility, this was added. Thanks.

"This transition to a big wet period can be explained by a southward shift of the storm tracks and polar jet stream over North America shift during this period (Asmerom et al., 2010)."

Line 260: The Pa/Th discussion is hard to follow without seeing the data. Please remind the reader that more positive values mean weakened circulation. I am no expert on this proxy, but my understanding is that Pa/Th integrates over the water column via particle scavenging. So I am not sure one can interpret the depth of the site as the depth to which the AMOC was affected. Of course AMOC changes at the surface and at depth are linked. The Pa/Th discussion should be clarified or left out.

>> The simplest is probably indeed to remove this discussion which is not needed for our conclusion

Line 283: Ice shelf destabilization by subsurface warming was suggested independently by (Marcott et al., 2011); please cite both.

>> This was added.

Line 293: The apparent stability from 20-14.7ka is somewhat misleading, because we know Greenland must have warmed in response to CO₂ and insolation. I think this is due to a masking effect; (summer) warming due to insolation and CO₂ rise is masked in Greenland temperature records by winter cooling during HS1 driven by AMOC weakening (Buizert et al., 2018). That explains why Greenland and the Laurentide retreat significantly prior to 14.7ka, while it appears there is no warming in Greenland records.

>> A reference was added to this work:

"During Heinrich 1 occurring during the last deglaciation, the story may be more complicated because of the CO₂ concentration and insolation increases. In this case, the occurrence of Heinrich Stadial 1 may counteract the increase in Greenland temperature records induced by CO₂ and insolation forcing through winter cooling driven by AMOC weakening as suggested by Buizert et al. (2018)."

Line 296: the link to EDML had also been suggested by Zhang et al. (2016), and possibly others?

>> The reference to Zhang et al. (2016) has been added:

"a link between EDML $\delta^{18}\text{O}$ record and low latitude signal over Heinrich Stadial 1 has already been suggested by Zhang et al. (2016)."

Line 299: consider removing "their coherent chronology". I don't think this adds to much new insight, personally.

>> Done

Figures: Please add panel labels 1a, 1b, 1c etc, which will make it easier to look up in

the caption, and refer to specific records in the text.

>> This will be done in the revised version.

We thank the two reviewers for their time and valuable comments that were taken into account as detailed in the following.

Response to Referee 2:

I think this a great study which is appropriate for the journal. My background is in triple-oxygen from a perspective outside of the hydrology community so I will stick to this. Although the change in ^{17}O -excess is not clearly located at 16.2 ka (as it is really indicated by δD), the interpretation that the trend in ^{17}O -excess over this time period is due to an increase in the relative humidity of the source region through HS1 seems valid. I have no issues with the interpretation and most everything seems to be consistent with previous work. I do have some question on the ^{17}O -excess data itself as mentioned in my most major comment below.

Line 78: The definitions for $\delta^{18}\text{O}$, ^{17}O -excess and D-excess should be given at first occurrence.

>> Done

My biggest issue with the study is the analytical methods from line 102 to 108 for the CoF3 based water fluorination and ^{17}O calibration. The methods for this are insufficient. The methods as is simply state that the reaction is done and gives a citation for Barkan and Luz 2005. High precision ^{17}O -excess measurements require extreme care in gas handling, the right mass-spec setup and consistent methodology. Some errors in gas handling can generate incorrect values at the 100 ppm level, let alone at the 5 ppm level reported as precision. Impure gas in particular can even yield a false measured relationship between $\delta^{18}\text{O}$ and ^{17}O -excess through scale distortion.

>> The method for water ^{17}O -excess measurements through fluorination was already described in several papers (all details are given in Barkan and Luz, 2005) but we agree with the reviewer that it is important to detail again here the methodology, especially since we improved precision compared to previous studies at LSCE using a new mass spectrometer (MAT 253).

The minimum things that would need to be known to trust the data in no particular order and not intended to be exhaustive are:

1: What mass-spec is being used?

>> MAT 253

2: How is the sample being introduced to the mass spec.

>> To reach the 5 ppm precision, we use the mass spectrometer in dual inlet mode and hence introduce the sample in gas phase (pure oxygen) through the classical sample bellow of the mass spectrometer. The sample is run (measured) against a standard (pure commercial oxygen).

3: What is the composition of the in-house reference relative to the samples (raw data would do) ?

>> The in-house references are several water standards calibrated every 3 years with respect to SMOW and SLAP provided by IAEA. For this study, we used in-house standards with the following isotopic composition: NEEM ($\delta^{18}\text{O} = -33.56 \text{ ‰}$; ^{17}O -excess = 32 ppm); OC ($\delta^{18}\text{O} = -54.05 \text{ ‰}$; ^{17}O -excess = 12 ppm); ROSS ($\delta^{18}\text{O} = -18.64 \text{ ‰}$; ^{17}O -excess = 37 ppm)

4: Some basics about the CoF₃ technique. (He carrier gas?, source of the CoF₃ for purity questions, reaction temperature etc.)

>> We used the published procedure: He carrier gas is purified through a trap of liquid nitrogen (-196°C); CoF₃ is bought by Sigma-Aldrich following numerous tests by several producers (Sigma-Aldrich CoF₃ gives the best reproducibility and precision according to the species measured by the mass spectrometer); the temperature reaction is 370°C.

5: How is the resulting O₂ gas purified?

>> The purification is done through a molecular sieve trap immersed in liquid nitrogen. Tests of GC purification were also performed during the development of the line at LSCE but did not improve the precision. Indeed, a systematic correction to a V-SMOW – SLAP scale is performed.

Every two-three weeks, three in-house standards bracketing the d18O and d17O values of the samples are run on the fluorination line. These standards are then used to put the d18O and d17O values on the V-SMOW – SLAP scale following procedure described in Schoenemann et al. (2013) and Landais et al. (2013). Then, everyday, only one in-house standard is run to check the day to day stability of the whole system (line + mass spectrometer) but this house standard is not used alone for shifting the d17O and d18O data, a full scale compression on the V-SMOW – SLAP scale is performed.

This will be explained better in the new manuscript and we propose the following paragraphs:

“In order to perform ^{17}O -excess measurements on water samples at LSCE, we follow the method described in details in (Barkan and Luz, 2005). In short, for each sample, 2 mL of water are injected in a helium flow purified by passing through a trap immersed in liquid nitrogen. Water vapor then reacts with CoF₃ (producer Sigma-Aldrich) in a nickel tube heated at 370°C to produce oxygen and fluorhydric acid which is trapped in liquid nitrogen at the outlet of the nickel tube. Oxygen is first trapped in a molecular sieve tube immersed in liquid nitrogen and then separated from helium and purified through 2 cycles of warming (+30°C) and cooling (-196°C) of the tube with molecular sieves. The oxygen is finally trapped in a manifold immersed in liquid helium. After warming the manifold at least 40 minutes at room temperature, the triple isotopic composition of produced oxygen is injected in the mass spectrometer (MAT 253) and measured by dual inlet against a reference O₂ gas (2 runs of 20 measurements).

Every day, at least one home standard is run with the batch of samples to check the stability of the fluorination line and mass spectrometer and a series of 3 water home standards, whose $\delta^{18}\text{O}$ and $\delta^{17}\text{O}$ values are calibrated on the SMOW – SLAP scale following Schoenemann et al. (2013), is run at least every month. For this study, the SMOW – SLAP calibrated home standards have $\delta^{18}\text{O}$ values of respectively -18.64 ‰, -33.56 ‰ and -54.05 ‰, hence bracketing the $\delta^{18}\text{O}$ values of the measured samples. The comparison of the measured and SMOW-SLAP calibrated $\delta^{18}\text{O}$ and $\delta^{17}\text{O}$ values then enable calibrating the $\delta^{18}\text{O}$ and ^{17}O -excess values of the NGRIP samples of this study following the method described in (Schoenemann et al., 2013; Landais et al., 2014). The resulting mean uncertainty is of 5 ppm (1 σ) for the ^{17}O -excess measurements of this study and we note that the use of the MAT 253 mass spectrometer gave much more stable results than a Delta V+ instrument used for previous studies at LSCE (e.g. Landais et al., 2012).”

This last point is in particular critical. Sample purification is something that for a long time went overlooked because there was no need to push precision, but to get down to sub-10 ppm with any hope of being similarly accurate seems to require GC purification of the gas. This includes CoF3 lines. If there is no GC purification to remove residual impurities, then I think some in the community would be inclined to not trust the ^{17}O -excess results to the detail needed for the submitted study. These impurities may be things that would clearly cause issues with the measurement, such as the mass 33 isobar generating NF_3 but also non-isobar impurities which can generate pressure baseline type effects via scattering. The errors in ^{17}O -excess induced by these impurities roughly scale with the $\delta^{18}\text{O}$ (or $\delta^{17}\text{O}$) so correcting for these can not be done by simply shifting the results but can be accounted for by applying a scale compression correction such as VSMOW-SLAP assuming that the impurity is a constant.

Related to this, how was, or was, the ^{17}O -excess data scaled? Is VSMOW-SLAP being used? It is stated on line 106 that “home” standards, which I guess should be “in-house” standards, spanning the SMOW-SLAP scale are run on a regular basis, but it is not clear if these have been calibrated to VSMOW-VSLAP2, or any other scaling. It might be this was intended by the Schoenemann et al. 2013 citation, which with the current language seems out of place. The reported values do seem like they plot in the correct region for being calibrated to VSMOW-SLAP.

>> See answer above.

Line 139-140 This is true enough that seawater ^{17}O -excess remains constant, at least in the recent past. However, this is an artifact of the 0.528 slope in the definition (and the value assigned to SLAP) in addition to the logarithmic form. There should be a citation here that amounts to essentially saying average glacial water falls on a 0.528 slope from modern seawater. Zach Sharp and company had a recent study in *Geochemical Perspectives Letters* which has data this could be calculated from.

>> We actually cannot claim for sure that ^{17}O -excess of seawater remains constant over the last deglaciation, no data can demonstrate it with the sufficient accuracy yet. What we want to emphasize is that the water cycle processes will not create an artificial ^{17}O -excess signal linked to the seawater $\delta^{18}\text{O}$ change of 1 permil over the last deglaciation as it is the case for $\delta^{18}\text{O}$ -excess. If ^{17}O -excess of seawater is modified, this modification will be conserved in the meteoric water ^{17}O -excess. This was indeed not very clear and we propose to clarify as follows:

“Because of its logarithmic definition, ^{17}O -excess is not sensitive to changes in $\delta^{18}\text{O}_{\text{sea water}}$ given that the ^{17}O -excess of global sea water remains constant with time. As a consequence, a change in sea water isotopic composition will only be transmitted to the ^{17}O -excess of the precipitation if the ^{17}O -excess of the evaporated sea-water is modified.”

Line 197 “Rhodes 2915” is cited.

>> Oops... Indeed, it is not correct and the “9” should be changed in “0”, thank you for pointing it.

As a minor point: In a general sense, I feel there is too much interpretation in the Result part:

>> We tried to better equilibrate this part and it goes along with more details in the methodology part, cf revised manuscript below.

Ice core evidence for decoupling between mid-latitude atmospheric water cycle and Greenland temperature during the last deglaciation

Amaëlle Landais^{1,*}, Emilie Capron^{2,3}, Valérie Masson-Delmotte¹, Samuel Toucanne⁴, Rachael Rhodes⁵, Trevor Popp², Bo Vinther², Bénédicte Minster¹, Frédéric Prié¹

¹ Laboratoire des Sciences du Climat et de l'Environnement, IPSL, UMR 8212, CEA-CNRS-UVSQ-UPS, Gif sur Yvette, France

² Centre for Ice and Climate, Niels Bohr Institute, University of Copenhagen, Juliane Maries Vej 30, DK-2900, Copenhagen, Denmark;

³ British Antarctic Survey, High Cross, Madingley Road, Cambridge, CB3 0ET, UK

⁴ IFREMER, Laboratoire Géophysique et enregistrement Sédimentaire, CS 10070, 29280 Plouzané, France

⁵ Department of Earth Sciences, University of Cambridge, Downing Street, Cambridge, CB2 3EQ
UK

* Corresponding author: phone : +33 (0)169084672, email : amaelle.landais@lsce.ipsl.fr, ORCID : 0000-0002-5620-5465

Keywords: Deglaciation, Water isotopes, ¹⁷O-excess, d-excess, Heinrich event, Mystery interval

Abstract

The last deglaciation represents the most recent example of natural global warming associated with large-scale climate changes. In addition to the long-term global temperature increase, the last deglaciation onset is punctuated by a sequence of abrupt changes in the Northern Hemisphere. Such interplay between orbital- and millennial-scale variability is widely documented in paleoclimatic records but the underlying mechanisms are not fully understood. Limitations arise from the difficulty in constraining the sequence of events between external forcing, high- and low- latitude climate and environmental changes.

Greenland ice cores provide sub-decadal-scale records across the last deglaciation and contain fingerprints of climate variations occurring in different regions of the Northern Hemisphere. Here, we combine new ice δ -excess and ^{17}O -excess records, tracing changes in the mid-latitudes, with ice $\delta^{18}\text{O}$ records of polar climate. Within Heinrich Stadial 1, we demonstrate a decoupling between climatic conditions in Greenland and those of the lower latitudes. While Greenland temperature remains mostly stable from 17.5 to 14.7 ka, significant change in the mid latitudes of northern Atlantic takes place at ~ 16.2 ka, associated with warmer and wetter conditions of Greenland moisture sources. We show that this climate modification is coincident with abrupt changes in atmospheric CO_2 and CH_4 concentrations recorded in an Antarctic ice core. Our coherent ice core chronological framework and comparison with other paleoclimate records suggests a mechanism involving two-step freshwater fluxes in the North Atlantic associated with a southward shift of the intertropical convergence zone.

Introduction

The last deglaciation (~19 thousand to 11 thousand years before present, ka) is the most recent major reorganization of global climate and is thus extensively documented by proxy records from natural climate archives. The wealth of high-resolution records from well-dated archives and data synthesis obtained over the past decades show two modes of climate variability during this period (e.g. Denton et al., 2010, Clark et al., 2012). The first is a long-term increase in global surface temperature and atmospheric CO₂ concentration between 18 and 11 ka. Superimposed on this is a sequence of centennial-scale transitions between three quasi-stable intervals documented in Northern Hemisphere temperature, namely (i) the Heinrich Stadial 1 (~17.5-14.7 ka), that encompasses the massive rafting episode known as Heinrich event 1 (from ~16 ka); (ii) the Bølling-Allerød warming phase (~14.7 to 12.9 ka) and (iii) the Younger Dryas cold phase (~12.9 to 11.7 ka). This three-step sequence coincides with rapid variations in the Atlantic Meridional Oceanic Circulation (AMOC) (Mc Manus et al., 2004), with evidence for a weak meridional overturning in the North Atlantic during the cold period encompassing Heinrich Stadial 1 and the Younger Dryas.

Our understanding of the mechanisms at play during these North Atlantic cold phases remains limited. First, recent studies challenge the earlier attribution of the AMOC slowdown during Heinrich Stadial 1 to the impact of the Iceberg Rafted Debris (IRD) from the Laurentide ice sheet through Hudson Strait (Alvarez-Solas et al., 2011). In particular, meltwater releases from the European ice sheet occurring as early as 19 or 20 ka may have played an important role in this AMOC slowdown (Toucanne et al., 2010; Stanford et al., 2011; Hodell et al., 2017).

Second, major global reorganizations of the hydrological cycle have been demonstrated during Heinrich Stadial 1. They can be separated in two phases. In North America, a first time interval characterized by low lake levels (referred to as the “big dry”, 17.5 to 16.1 ka) was followed by a second time interval with high lake levels (referred to as the “big wet”, 16.1 to 14.7 ka) (Broecker et al., 2012),

both apparently occurring during a stable cold phase in Greenland temperature. The second phase of Heinrich Stadial 1 is also associated with a weak East Asian monsoon interval (Zhang et al., 2014), understood to reflect a southward shift of the Inter-tropical Convergence Zone (ITCZ). While there is growing evidence for large-scale reorganizations of climate and low- to mid- latitude atmospheric water cycle within Heinrich Stadial 1, the exact sequence of events is not known with sufficient accuracy to understand the links between changes in North Atlantic climate, AMOC, and the lower latitude water cycle.

Linking changes in the high latitudes of the North Atlantic and the mid- to low- latitudes requires precise absolute chronologies such as those obtained from annual layer counting of Greenland ice (e.g. Andersen et al., 2006) or U/Th dating of speleothems (e.g. Zhang et al., 2014). Unfortunately, absolute dating uncertainties increase above a hundred years during the last deglaciation, precluding a direct comparison of proxy records at the centennial scale. In this study, we circumvent this difficulty by using proxy records measured on Greenland ice cores that represent both Greenland temperature and mid-latitude moisture source conditions.

Analytical method

Here, we present new water isotope records ($\delta^{18}\text{O}$, $\text{d-excess} = \delta\text{D} - 8 \times \delta^{18}\text{O}$, $^{17}\text{O-excess} = \ln(\delta^{17}\text{O} + 1) - 0.528 \times \ln(\delta^{18}\text{O} + 1)$) from the NGRIP ice core (NGRIP et al., 2004), reported on the annual-layer counted Greenland Ice Core Chronology 2005 (hereafter GICC05, Rasmussen et al., 2006; Svensson et al., 2008), and associated with relatively small absolute uncertainties over the last deglaciation (maximum counting 1σ error of 100-200 yr). Other Greenland and Antarctic ice cores have been aligned on the GICC05 chronology, with a maximum relative dating uncertainty of 400 years over the last deglaciation (Rasmussen et al., 2008; Bazin et al. 2013; Veres et al., 2013).

The new NGRIP $\delta^{18}\text{O}$ and δD dataset was obtained at Laboratoire des Sciences du Climat et de l'Environnement (LSCE) using a PICARRO laser cavity ring-down spectroscopy (CRDS) analyzer. The accuracy for $\delta^{18}\text{O}$ and δD measurements displayed here is about 0.1‰ and 1‰ respectively. This new

dataset completes the NGRIP high-resolution isotopic dataset published over the time period 11.5 to 14.7 ka with $\delta^{18}\text{O}$ and δD measured respectively at the University of Copenhagen and at the Institute of Arctic and Alpine Research (INSTAAR) Stable Isotope Lab (SIL) (University of Colorado), respectively. $\delta^{18}\text{O}$ analyses were performed at the Niels Bohr Institute (University of Copenhagen) using a CO_2 equilibration technique (Epstein et al., 1953) with an analytical precision of 0.07‰. δD measurements at INSTAAR were made via an automated uranium reduction system coupled to a VG SIRA II dual inlet mass spectrometer (Vaughn et al., 1998). Analytical precision for δD is $\pm 0.5\text{‰}$ or better. Both series show similar $\delta^{18}\text{O}$ values, in agreement with the reference $\delta^{18}\text{O}$ series for NGRIP over the last climatic cycle (NGRIP community members, 2004) within error bars. However, while both LSCE and INSTAAR SIL d-excess series display the same 3.5‰ decrease over the onset of Bølling-Allerød, the mean d-excess level differs by 2.5‰ between the two timeseries. Despite several home standard intercalibrations between the two laboratories, this difference remains unexplained and prevents any further discussion on the absolute NGRIP d-excess levels. The new and published NGRIP d-excess dataset are combined after a shift of the INSTAAR SIL d-excess series by -2.5‰.

In order to perform ^{17}O -excess measurements on water samples at LSCE, we follow the method described in details in (Barkan and Luz, 2005). In short, for each sample, 2 μL of water are injected in a helium flow purified by passing through a trap immersed in liquid nitrogen. Water vapor then reacts with CoF_3 (producer Sigma-Aldrich) in a nickel tube heated at 370°C to produce oxygen and fluorhydric acid which is trapped in liquid nitrogen at the outlet of the nickel tube. Oxygen is first trapped in a molecular sieve tube immersed in liquid nitrogen and then separated from helium and purified through 2 cycles of warming (+30°C) and cooling (-196°C) of the tube with molecular sieves. The oxygen is finally trapped in a manifold immersed in liquid helium. After warming the manifold at least 40 minutes at room temperature, the triple isotopic composition of produced oxygen is injected in the mass spectrometer (MAT 253) and measured by dual inlet against a reference O_2 gas (2 runs of 20 measurements).

Every day, at least one home standard is run with the batch of samples to check the stability of the fluorination line and mass spectrometer. In addition, a series of 3 water home standards, whose $\delta^{18}\text{O}$ and $\delta^{17}\text{O}$ values are calibrated on the SMOW – SLAP scale following Schoenemann et al. (2013), is run at least every month. For this study, the SMOW – SLAP calibrated home standards have $\delta^{18}\text{O}$ values of respectively -18.64 ‰, -33.56 ‰ and -54.05 ‰, hence bracketing the $\delta^{18}\text{O}$ values of the measured samples. The comparison of the measured and SMOW-SLAP calibrated $\delta^{18}\text{O}$ and $\delta^{17}\text{O}$ values then enables calibrating the $\delta^{18}\text{O}$ and ^{17}O -excess values of the NGRIP samples following the method described in (Schoenemann et al., 2013; Landais et al., 2014). The resulting mean uncertainty is of 5 ppm (1σ) for the ^{17}O -excess measurements of this study. Note that the use of the MAT 253 mass spectrometer gave more stable results than a Delta V+ instrument used for previous studies at LSCE (e.g. Landais et al., 2012).

Results: water isotopic records at NGRIP

Our 1518 new measurements of $\delta^{18}\text{O}$ and d-excess on the NGRIP ice core cover the time period 14.5 to 60 ka (Figure 1) and we present 454 duplicate measurements of ^{17}O -excess over the time period ranging from 9.6 to 20 ka (Figure 2). $\delta^{18}\text{O}$ is a qualitative proxy for local surface temperature. Comparisons between ice core $\delta^{18}\text{O}$ data and paleotemperature estimates from borehole temperature profile inversion and abrupt temperature changes inferred from isotopic measurements on trapped air showed that the $\delta^{18}\text{O}$ -temperature relationship at NGRIP varies from 0.3 to 0.5 ‰.°C⁻¹ during glacial-interglacial periods (Buizert et al., 2014; Dahl-Jensen et al., 1998; Kindler et al., 2014).

The second-order parameter d-excess (Dansgaard, 1964) is used in Greenland ice cores to track past changes in evaporation conditions or shifts in moisture sources (Johnsen et al., 1989; Masson-Delmotte et al., 2005a). Evaporation conditions affect the initial vapor d-excess through the impact of surface humidity and sea surface temperature on kinetic fractionation (Jouzel et al., 1982). Recent vapour monitoring and modelling studies show that the d-excess signal of the moisture source can be preserved in polar vapour and precipitation after transportation towards polar regions (Bonne et al.,

2015; Pfahl and Sodemann, 2014). This signal can however be altered during distillation due to the sensitivity of equilibrium fractionation coefficients to temperature, leading to alternative definitions using logarithm formulations for Antarctic ice cores (Uemura et al., 2012; Markle et al., 2016). Finally, changes in $\delta^{18}\text{O}_{\text{sea water}}$ also influence $\delta^{18}\text{O}$ and d-excess in polar precipitation. Summarizing, d-excess in Greenland ice core is a complex tracer: interpreting its past variations in terms of changes in evaporation conditions (sea surface temperature or humidity) requires deconvolution of the effects of glacial-interglacial changes in $\delta^{18}\text{O}_{\text{sea water}}$ and condensation temperature.

^{17}O -excess provides complementary information to d-excess (Landais et al., 2008; Landais et al., 2012). At evaporation, d-excess and ^{17}O -excess are both primarily influenced by the balance between kinetic and equilibrium fractionation, itself driven by relative humidity at the sea surface. During transport, while d-excess is influenced by distillation effects during atmospheric cooling, ^{17}O -excess is largely insensitive to this effect, except at very low temperatures in Antarctica (Winkler et al., 2012). Conversely, ^{17}O -excess is affected by recycling or mixing of air masses along the transport path from low to high latitudes (Risi et al., 2010), and by the range over which supersaturated conditions occur, itself affected for instance by changes in sea-ice extent or temperature along the transport path (Schoenemann et al., 2014). Because of its logarithmic definition, ^{17}O -excess is not sensitive to changes in $\delta^{18}\text{O}_{\text{sea water}}$ given that the ^{17}O -excess of global sea water remains constant with time. As a consequence, a change in sea water isotopic composition will only be transmitted to the ^{17}O -excess of the precipitation if the ^{17}O -excess of the evaporated sea-water is modified.

As previously reported for the central Greenland GRIP ice core (Masson-Delmotte et al., 2005b; Jouzel et al., 2005), the NGRIP $\delta^{18}\text{O}$ and d-excess records exhibit a systematic anti-correlation during the abrupt Dansgaard-Oeschger (DO) events of the last glacial period and last deglaciation (Bølling-Allerød and Younger Dryas), with d-excess being higher during cool Greenland Stadials and lower during warm Greenland Interstadials.

The origin of moisture may be different at GRIP and NGRIP. While both sites are expected to receive most of their moisture from the North Atlantic and North America (Werner et al., 2001, Landais et al., 2012, Langen and Vinther, 2009) with modulation partly linked to sea ice extent (Rhines et al., 2014), the northwestern NGRIP site may also receive moisture from North Pacific (Langen and Vinther, 2009). Nevertheless, the two sites depict similar amplitudes of d-excess variations across DO events (Figure 1). We note that this contrasts with a slightly lower amplitude (typically by 1‰) of abrupt $\delta^{18}\text{O}$ changes at NGRIP compared to GRIP.

The fact that d-excess increases (by 3.5 ± 1 ‰) when $\delta^{18}\text{O}$ decreases (by 4 ± 1 ‰) during Greenland stadials relative to interstadials may at least partly reflect the influence of local temperature changes on d-excess, challenging a simple interpretation in terms of changes in source conditions. We note one exception, the Heinrich Stadial 1 cold phase preceding the onset of the Bølling-Allerød at 14.7 ka. In this case, $\delta^{18}\text{O}$ remains almost stable from 17.5 to 14.7 ka on the three Greenland ice cores NGRIP, GRIP and GISP2 displayed on Figure 2. Over this period, $\delta^{18}\text{O}$ variations are smaller than 1 ‰, i.e. less than one fourth of the average amplitude in $\delta^{18}\text{O}$ changes across DO events, suggesting no large temperature change in Greenland during this period. The link between flat $\delta^{18}\text{O}$ and minimal temperature variability can be challenged since a mean temperature signal can be masked by a change in seasonality of moisture source origin on the $\delta^{18}\text{O}$ record (Boyle et al., 1994; Krinner et al., 1997). However, our assumption of stable temperature is supported by constant $\delta^{15}\text{N}$ of N_2 values in the GISP2 and NGRIP ice cores (Buizert et al., 2014), $\delta^{15}\text{N}$ of N_2 being an alternative paleothermometry tool in ice core that is not affected by processes within the water cycle (Severinghaus and Brook, 1999). In contrast to this almost stable $\delta^{18}\text{O}$ signal, d-excess depicts an average 2.2 ‰ increase at 16.1 ka (more than 60% of the average amplitude during DO events) with a larger amplitude at GRIP (2.7 ‰) than at NGRIP (1.7 ‰) (Figure 2). In this case, the increase in d-excess cannot be explained by any Greenland temperature change, and therefore demonstrates a decoupling between cold and stable Greenland temperatures and changing climatic conditions at lower latitudes during Heinrich Stadial 1 (see also SOM).

While the ^{17}O -excess level is similar at the Last Glacial Maximum (i.e. before 19 ka on Figure 2) and the Early Holocene (40 ppm), it also shows significant variations during the last deglaciation. Most of these variations co-vary with those of $\delta^{18}\text{O}$ such as the four main oscillations during the Bølling-Allerød and the onset and end of the Younger Dryas. They can be interpreted as parallel variations in the Greenland temperature and lower latitude climate with a possible contribution of local temperature through kinetic effects. Again, a major difference occurs during Heinrich Stadial 1. While the $\delta^{18}\text{O}$ record is relatively stable, ^{17}O -excess exhibits a decreasing trend (strongest between 17.5 and 16.1 ka) before a minimum level is reached between 16.1 to 14.7 ka. We therefore observe a clear and synchronous signal in both d-excess and ^{17}O -excess dated around 16.2 ka from statistical analysis (cf. section statistical analyses in SOM).

Discussion

The ^{17}O -excess and d-excess transitions at 16.2 ka do not have any clear counterpart in $\delta^{18}\text{O}$ (cf section correlation in SOM) and no temperature variation at that time was recorded in the $\delta^{15}\text{N}$ of N_2 record. We interpret these patterns as illustrating a reorganization of climatic conditions and/or water cycle at latitudes south of Greenland. A similar shift in ^{17}O -excess has already been observed during Heinrich Stadial 4 in the NEEM ice core, while the $\delta^{18}\text{O}$ record exhibits a constant low level (Guillevic et al., 2014). This pattern was also attributed to a change in the water cycle and/or climate at lower latitudes.

The Greenland water stable isotope records demonstrate a change in the water cycle and/or climate at lower latitudes at 16.2 ka when Greenland conditions were relatively stable and cold. This change at low latitudes is confirmed by the high resolution atmospheric CH_4 concentration record from the WAIS Divide ice core (Rhodes et al., 2015) (Figure 2). At 16.2 ka, the CH_4 record indeed exhibits a 30 ppbv peak hypothesized to reflect more CH_4 production in Southern Hemisphere wetlands, driven by wetter conditions due to a southward shift of the tropical rainbelts associated with the ITCZ (Rhodes et al., 2015). The parallel increase of atmospheric CO_2 concentration by 10 ppm in ~ 100 years (Marcott et al., 2013) is understood to result from increased terrestrial carbon fluxes or enhanced air-sea gas

exchange in the Southern Ocean (Bauska et al., 2014). We also highlight an unusual characteristic of the bipolar seesaw pattern in Antarctic ice core $\delta^{18}\text{O}$ records at 16.2 ka. As observed during all Greenland Stadials of the last glacial period, Antarctic $\delta^{18}\text{O}$ also increases during Heinrich Stadial 1 (e.g. EPICA community members, 2006), through the warming phase of Antarctic Isotopic Maximum 1. The EPICA Dronning Maud Land (EDML) ice core, drilled in the Atlantic sector of Antarctica, shows an associated two step $\delta^{18}\text{O}$ increase. The first step, marked by a strong increasing trend, is followed by a change in slope at 16.2 ka. The second step is characterized by a slower increasing trend from 16.2 to 14.7 ka (EPICA community members, 2006; Stenni et al., 2011) (Figure 2). The EDML $\delta^{18}\text{O}$ variations are expected to be closely connected to changes in AMOC due to the position of the ice core site on the Atlantic sector of the East Antarctic plateau and a link between EDML $\delta^{18}\text{O}$ record and low latitude signal over Heinrich Stadial 1 has already been suggested by Zhang et al. (2016). For other Antarctic sites, the change of slope around 16.2 ka is less clear, probably due to the damping effect of the Southern Ocean or because other climatic effects linked to atmospheric teleconnections with the tropics affect the Pacific and Indian sectors of Antarctica (Stenni et al., 2011, WAIS Divide members, 2013; Buiron et al., 2012). A change in the teleconnections between West Antarctic climate and tropical regions is also observed around 16.2 ka (Jones et al., 2018). Summarizing, our synthesis of ice core records clearly demonstrates a climate shift at 16.2 ka, identified in proxy records sensitive to shifts in tropical hydrology (CH_4), mid-latitude hydrological cycle changes in the Atlantic basin (Greenland second order isotopic tracers), as well as in Antarctic climate dynamics in the Atlantic basin. This suggests some reorganization of water cycle in the Atlantic region (possibly involving AMOC) related to surface shifts in the ITCZ at 16.2 ka. This does not appear to affect the high latitudes of the North Atlantic as Greenland temperatures stay uniformly cold. Uniformly cold conditions in Greenland are generally observed during Heinrich Stadials of the last glacial period with temperature and $\delta^{18}\text{O}$ levels that are not significantly lower than temperature levels observed during Greenland Stadials (Kindler et al., 2014; Guillevic et al., 2014). Because Greenland is surrounded by large sea ice during Greenland Stadials and Heinrich Stadials (Hoff et al., 2016), an explanation may be that central

Greenland temperatures are saturated during cold periods so that AMOC modifications occurring south of the sea ice edge are not influencing significantly Greenland temperatures. During Heinrich 1 occurring during the last deglaciation, the story may be more complicated because of the CO₂ concentration and insolation increases. In this case, the occurrence of Heinrich Stadial 1 may counteract the increase in Greenland temperature records induced by CO₂ and insolation forcing through winter cooling driven by AMOC weakening as suggested by Buizert et al. (2018).

At low latitudes, an ITCZ shift at 16.2 ka is clearly expressed through a weak monsoon interval in East Asian speleothem records and through change in hydrology in the low-latitude Pacific region, Cariaco Basin and Brazil (Partin et al., 2007; Deplazes et al., 2013; Russell et al., 2014; Strikis et al., 2015). Since we have ruled out a local temperature signal at 16.2 ka in Greenland, the origin of the Greenland δ-excess and ¹⁷O-excess changes around 16.2 ka is also linked to changes in the climate of the source evaporative regions. When evaporation conditions change, they affect the proportion of kinetic versus equilibrium fractionation, and cause similar trends in both δ-excess and ¹⁷O-excess. Both of them indeed increase when kinetic fractionation is more important, i.e. when relative humidity decreases, or when a change in sea ice modifies the evaporative conditions (Klein et al., 2015; Kopec et al., 2016). However, δ-excess in the atmospheric vapor is affected by distillation toward higher latitudes, and strongly depends on the source-site temperature gradient, while ¹⁷O-excess preserves better the initial fingerprint of relative humidity of the evaporative region.

As a result, the opposing trends observed in δ-excess and ¹⁷O-excess at 16.2 ka can most probably be explained by an increase of both the relative humidity and the sea surface temperature of the evaporative source regions for Central and North Greenland. Despite known limitations (Winkler et al., 2012, Schoenemann and Steig, 2016), the classical approach for inferring changes in source relative humidity and surface temperature from δ-excess and ¹⁷O-excess in Greenland (Masson-Delmotte et al., 2005a; Landais et al., 2012) suggests respective increases of the order of 3°C and 8% for temperature and relative humidity of the source evaporative regions respectively. The larger δ-excess

increase at the transition between Phase 1 and Phase 2 of Heinrich Stadial 1 observed at GRIP compared to NGRIP is compatible with a larger proportion of GRIP moisture provided by the mid-latitude North Atlantic for this site. A larger increase in the sea surface temperature of the source of moisture for GRIP compared to NGRIP would also reduce the source-site temperature gradient and is fully compatible with the 2 ‰ less depleted level of $\delta^{18}\text{O}$ at GRIP, compared to NGRIP, during Phase 2. The increases in both temperature and relative humidity of the Greenland source regions suggest a more intense evaporative flux from lower latitudes starting at 16.2 ka. Such features could be explained either by a local climate signal of evaporative regions or by a southward shift of evaporative source regions toward warmer and more humid locations. The signal of source temperature increase is in line with earlier interpretations of Greenland d-excess changes (Steffensen et al., 2018; Masson-Delmotte et al., 2005b). The signal of source humidity increase may at least partly explained by wetter conditions in the continental North America evaporative source regions, which are known to partly affect Greenland moisture today in addition to the main source in Northern Atlantic (Werner et al., 2001; Langen and Vinther, 2009). This relative humidity signal reconstructed from Greenland ^{17}O -excess at the transition between Phase 1 and Phase 2 of Heinrich Stadial 1 coincides with the onset of the “big wet” period in North American records (Broecker and Putnam, 2012). This transition to a big wet period can be explained by a southward shift of the storm tracks and polar jet stream over North America shift during this period (Asmerom et al., 2010).

We now explore paleoceanographic records to search for a fingerprint of climate and/or AMOC reorganization at 16.2 ka in the North Atlantic region and possible implications for our ice core records. Such comparison of ice core and marine sediment records appears insightful despite existing limitations attached to relative chronologies. First, high resolution proxy records of surface sea temperature in the East Atlantic, near Europe, depict a clear warming in the middle of Heinrich Stadial 1 (Bard et al., 2000; Matrat et al., 2014, Figure 3). This signal is coherent with our interpretation of Greenland d-excess increase at 16.2 ka. In the deep Western Atlantic, no specific feature emerges

between Phase 1 and Phase 2 of Heinrich Stadial 1 from the multi-centennial resolution record of Pa/Th, a proxy of AMOC strength (McManus et al., 2004).

Heinrich Stadial 1 is associated with at least two major Iceberg Rafted Debris (IRD) discharges first identified near the Iberian margin (Bard et al., 2000). They may reflect either the impact of changes in ocean conditions on ice shelf and ice sheet stabilities (Alvarez-Solas et al., 2011). Alternatively, the iceberg discharges themselves may have affected the AMOC, which is known to have major impacts on patterns of sea surface temperature, sea ice, atmospheric circulation, and climate over surrounding continents. The first IRD phase originated from ice sheet discharges from Northern Europe and Iceland, causing strong reorganizations in deep circulation of the North East Atlantic (Stanford et al., 2011, Grousset et al., 2001; Peck et al., 2006) while the second IRD phase is caused by discharges from the Laurentide ice sheet. Recent studies (e.g. Hodell et al., 2017, Toucanne et al., 2015) suggest that all IRD phases occur after 16.2 ka, during Heinrich Stadial 1 Phase 2. Before that, Heinrich Stadial Phase 1 is associated with a strong increase of sediment fluxes due to meltwater arrival through terrestrial terminating ice streams originating from both European and American sides of the North Atlantic as a response to the beginning of the deglaciation (Toucanne et al., 2015, Ullman et al., 2015, Leng et al., 2018) (Figure 3). During the first slowdown of AMOC during Phase 1 of Heinrich Stadial 1, the associated warming of subsurface water would hence enable the destabilization of marine ice-shelves occurring during Phase 2 (Alvarez-Solas et al., 2011; Marcott et al., 2011). This second phase of Heinrich Stadial 1 is also associated with an extensive sea ice production, south of Greenland (Hillaire-Marcel and De Vernal, 2008). The increase of North Atlantic sea ice extent and major iceberg discharges during the second phase of Heinrich Stadial 1 are coherent with a southward shift of the evaporative region providing moisture to Greenland supported by δ -excess data, and a southward shift of tropical rainbelts (Chiang and Bitz, 2005), affecting southern hemisphere CH₄ sources (Rhodes et al., 2015).

Conclusions

Combined measurements of d-excess and ^{17}O -excess along the NGRIP ice core demonstrate a decoupling between a cold and stable Greenland climate and changes in hydroclimate at lower latitudes during the Heinrich Stadial 1, also referred to as the “Mystery Interval” (Denton et al., 2006).

These new measurements hence confirm the previous studies of Zhang et al. (2014, 2016) and Rhodes et al. (2015). While Greenland temperature remains mostly stable from 20 to 14.7 ka, a large-scale climatic reorganization takes place at 16.2 ka, associated with warmer and wetter conditions at the location of Greenland moisture sources. Based on a coherent temporal framework linking the different ice core records, we show that this event coincides with changes in the characteristics of the bipolar seesaw pattern as observed in the Atlantic sector of Antarctica, and has a fingerprint in global atmospheric composition through sharp changes in atmospheric CO_2 and CH_4 concentrations.

Based on these new ice core records and the comparison with marine and terrestrial records, we propose the following sequence of events during the last deglaciation. First, the initiation of Heinrich Stadial 1 occurs at 17.5 ka or earlier, with meltwater arrival from the terrestrial terminating ice-streams synchronous with a decrease in the North Atlantic sea surface temperature off-shore Europe, a first AMOC slowdown, drier conditions in North America, and an increase in Antarctic temperature as well as in atmospheric CO_2 and CH_4 concentrations. No fingerprint of this first phase of Heinrich Stadial 1 is identified in Greenland water stable isotope records: $\delta^{18}\text{O}$ (and thus local temperature), ^{17}O -excess and d-excess remain stable. A possible explanation for such stability is that the high-latitude warming induced by the increase in the summer insolation at high latitude over the beginning of the deglaciation is counterbalanced in Greenland by regional changes in e.g. increased albedo due to sea ice extent or reduced transport of heat by the atmospheric circulation towards central Greenland, which both can result from a reduced AMOC strength. The global event occurring at 16.2 ka marks the onset of the second phase of Heinrich Stadial 1. It is associated with (i) strong iceberg discharges due to dynamical instability of the Laurentide ice sheet, probably induced by the accumulation of subsurface ocean heat due to a slowdown of AMOC during Phase 1, (ii) a widespread reorganization of the atmospheric water cycle in the Atlantic region, with significant changes in d-excess and ^{17}O -excess in Greenland, as well

as (iii) the initiation of weak monsoon interval in East Asia and (iv) the transition from a “big dry” episode to a “big wet” episodes in North America. We note that this sequence of events within Heinrich Stadial 1 is invisible in all available Greenland temperature proxy records, which only display an abrupt warming at the onset of the Bølling-Allerød (14.7 ka).

Attached to a bipolar synchronised chronological framework, our new ice core data provide a unique benchmark to test the ability of Earth system models to correctly resolve the sub-millennial mechanisms at play during the last deglaciation, and especially the relationships between meltwater fluxes, the state of the North Atlantic ocean circulation, the Laurentide ice sheet instability, changes at the moisture sources of Greenland ice cores, the response of hydroclimate at low and high latitudes, as well as the net quantitative effects on global methane and carbon budgets.

Acknowledgements

The research leading to these results has received funding from the European Research Council under the European Union’s Seventh Framework Programme (FP7/2007-2013) / RC agreement number 306045. E. C. is funded by the European Union's Seventh Framework Programme for research and innovation under the Marie Skłodowska-Curie grant agreement no 600207. R.H.R. received funding from a European Commission Horizon 2020 Marie Skłodowska-Curie Individual Fellowship (no. 658120, SEADOG). We thank Myriam Guillevic and Jean Jouzel for very useful discussions.

References

Álvarez-Solas, J., Montoya, M., Ritz, C., Ramstein, G., Charbit, S., Dumas, C., Nisancioglu, K., Dokken, T. and Ganopolski, A.: Heinrich event 1: an example of dynamical ice-sheet reaction to oceanic changes, *Clim. Past*, 7(4), 1297–1306, 2011

Andersen, K.K., Svensson, A., Johnsen, S.J., Rasmussen, S.O., Bigler, M., Roethlisberger, R., Ruth, U., Siggaard-Andersen, M.L., Steffensen, J.P., Dahl-Jensen, D., Vinther, B.M. and Clausen, H.B., The Greenland Ice Core Chronology 2005, 15-42 ka. Part 1: constructing the time scale, *Quat. Sci. Rev.*, 25(23–24), 3246–3257, 2006

Asmerom, Y., Polyak, V.J. and Burns, S.J., Variable winter moisture in the southwestern United States linked to rapid glacial climate shifts, *Nature Geosci.*, 3, 114-117, 2010

- Bard, E., Rostek, F., Turon, J.L. and Gendreau, S.: Hydrological Impact of Heinrich Events in the Subtropical Northeast Atlantic *Science*, 289, 1321–1324, 2000
- Barkan, E. and Luz, B.: High precision measurements of $17\text{O}/16\text{O}$ and $18\text{O}/16\text{O}$ ratios in H_2O , *Rapid Commun., Mass Spectrom.*, 19(24), 3737–3742, 2005
- Bauska, T.K., Baggenstos, D., Brook, E.J., Mix, A.C., Marcott, S. A. and Petrenko, V., Carbon isotopes characterize rapid changes in atmospheric carbon dioxide during the last deglaciation, *PNAS*, 113(13), 3465-3470, 2016
- Bazin, L., Landais, A., Lemieux-Dudon, B., Toyé Mahamadou Kele, H., Veres, D., Parrenin, P., Martinerie, P., Ritz, C., Capron, E., Lipenkov, V., Loutre, M.F., Raynaud, D., Vinther, B., Svensson, A., Rasmussen, S.O., Severi, M., Blunier, T., Leuenberger, M., Fischer, H., Masson-Delmotte, V., Chappellaz, J. and Wolff, E.: An optimized multi-proxy, multi-site Antarctic ice and gas orbital chronology (AICC2012): 120-800 ka, *Clim. Past*, 9(4), 1715–1731, 2013
- Bonne, J.-L., H. C. Steen-Larsen, C. Risi, M. Werner, H. Sodemann, J.-L. Lacour, X. Fettweis, G. Cesana, M. Delmotte, O. Cattani, P. Vallelonga, H. A. Kjaer, C. Clerbaux, A. E. Sveinbjörnsdóttir and V. Masson-Delmotte: The summer 2012 Greenland heat wave: In situ and remote sensing observations of water vapor isotopic composition during an atmospheric river event; *Journal of Geophysical Research- Atmospheres*, 120(7), 2970-2989, 2015
- Boyle, E. A.: Cool tropical temperatures shift the global $\delta^{18}\text{O}$ -T relationship: An explanation for the ice core $\delta^{18}\text{O}$ -borehole thermometry conflict?, *Geophysical Research Letters*, 24(3), 273-276, 1994
- Broecker, W. and Putnam, A.E.: How did the hydrologic cycle respond to the two-phase mystery interval?, *Quat. Sci. Rev.*, 57(C), 17–2, 2012
- Buiron, D., Stenni, B., Chappellaz, J., Landais, A., Baumgartner, M., Bonazza, M., Capron, E., Frezzotti, M., Kageyama, M., Lemieux-Dudon, B., Masson-Delmotte, V., Parrenin, F., Schilt, A., Selmo, E., Severi, M., Swingedouw, D. and Udisti, R.: Regional imprints of millennial variability during the MIS 3 period around Antarctica, *Quat. Sci. Rev.*, 48, 99-112, 2012
- Buizert, C., Gkinis, V., Severinghaus, J.P., He, F., Lecavalier, B. S., Kindler, P., Leuenberger, M., Carlson, A.E., Vinther, B., Masson-Delmotte, V., White, J.W.C., Liu, Z., Otto-Bliesner and B., Brook, E.J.: Greenland temperature response to climate forcing during the last deglaciation, *Science*, 345(6201), 1177–1180, 2014
- Buizert, C., Keisling, B.A., Box, J.E., He, F., Carlson, A.E., Sinclair, G. and DeConto, R.M., Greenland-Wide Seasonal Temperatures During the Last Deglaciation. *Geophys. Res. Lett.* 45, 1905-1914, 2018
- Chiang, J. C. H. and Bitz, C.M.: Influence of high latitude ice cover on the marine Intertropical Convergence Zone, *Clim. Dyn.*, 25(5), 477–496, 2005
- Clark, P.U., Shakun, J.D., Baker, P.A., Bartlein, P.J., Brewer, S., Brook, E., Carlson, A.E., Cheng, H., Kaufman, D.S., Liu, Z., Marchitto, T.M., Mix, A.C., Morrill, C., Otto-bliesner, B.L., Pahnke, K., Russell, J.M., Whitlock, C., Adkins, J.F., Blois, J.L., Clark, J., Colman, S. M., Curry, W. B., Flower, B. P., He, F., Johnson, T.C., Lynch-Stieglitz, J., Markgraf, V., Mcmanus, J., Mitrovica, J.X., Moreno, P.I. and Williams, J.W.: Global climate evolution during the last deglaciation, *PNAS*, 109(19), E1134-E1142, 2012
- Dahl-Jensen, D., Mosegaard, K., Gundestrup, N., Clow, G.D., Johnsen, S.J., Hansen, A.W. and Balling, N.: Past Temperatures Directly from the Greenland Ice Sheet, *Science*, 282(5387), 268–271, 1998
- Dansgaard, W.: Stable isotopes in precipitation, *Tellus*, 16, 436–468, 1964

Denton, G. H., Broecker, W.S. and Alley, R.B.: The mystery interval 17.5 to 14.5 kyrs ago PAGES news, 14(20), 14–16, 2006

Denton, G.H., Anderson, R.F., Toggweiler, J.R., Edwards, R.L., Schaefer, J.M. and Putnam, A.E.: The Last Glacial Termination, *Science*, 328, 1652–1656, 2010

Deplazes, G., Luckge, A., Peterson, L.C., Timmermann, A., Hamann, Y., Hughen, K.A., Rohl, U., Laj, C., Cane, M.A., Sigman, D.M. and Haug, G.H., Links between tropical rainfall and North Atlantic climate during the last glacial period, *Nature Geosci.*, 6, 213–217, 2013

Epica community members: One-to-one coupling of glacial climate variability in Greenland and Antarctica, *Nature*, 444(7116), 195–198, 2006

Epstein, S., Buchsbaum, R., Lowenstam, H.A. and Urey, H.: Revised carbonate-water isotopic temperature scale *GSA bulletin*, 64(11), 1315–1325, 1953

Gherardi, J.-M., Labeyrie, L., McManus, J.F., Francois, R., Skinner, L.C. and Cortijo, E.: Evidence from the Northeastern Atlantic basin for variability in the rate of the meridional overturning circulation through the last deglaciation, *Earth Planet. Sci. Lett.*, 240(3), 710–723, 2005

Groote, P. M., Stuiver, M., White, J. W. C., Johnsen, S. and Jouzel, J.: Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores, *Nature*, 366, 552–554, 1993

Grousset, F.: Les changements abrupts du climat depuis 60 000 ans / Abrupt climatic changes over the last 60,000 years, *Quaternaire*, 12(4), 203–211, 2001

Guillevic, M., Bazin, L., Landais, A., Stowasser, C., Masson-Delmotte, V., Blunier, T., Eynaud, F., Falourd, S., Michel, E., Minster, B., Popp, T., Prié, F. and Vinther, B. M.: Evidence for a three-phase sequence during Heinrich Stadial 4 using a multiproxy approach based on Greenland ice core records, *Clim. Past*, 10, 2115–2133, 2014

Hillaire-Marcel, C. and De Vernal, A.: Stable isotope clue to episodic sea ice formation in the glacial North Atlantic, *Earth Plan. Sci. Lett.*, 268, 143–150, 2008

Hoff U, Rasmussen TL, Stein R, Ezat MM, Fahl K. Sea ice and millennial-scale climate variability in the Nordic seas 90 kyr ago to present. *Nat Commun.*, 7, 12247, 2016.

Hodell, D.A., Nicholl, J.A., Bontognali, T.R.R., Danino, S., Dorador, J., Dowdeswell, J.A., Einsle, J., Kuhlmann, H., Martrat, B., Mlenceck-vautravars, M.J., Rodríguez-tovar, F.J. and Röhl, U.: Anatomy of Heinrich Layer 1 and its role in the last deglaciation, *Paleoceanography and Paleoclimatology*, 32(3), 10.1002/2016PA003028, 2017

Johnsen, S. J., Dansgaard, W. and White, J.W.C.: The origin of Arctic precipitation under present and glacial conditions, *Tellus B*, 41B(4), 452–468, 1989

Jones, T. R., Roberts, W. H. G., Steig, E.J., Cuffey, K.M., Markle, B.R. and White, J.W.C.: Southern Hemisphere climate variability forced by Northern Hemisphere ice-sheet topography, *Nature*, 554, 351–355, 2018

Jouzel, J., Merlivat, L. and Lorius, C.: Deuterium excess in an East Antarctic ice core suggests higher relative humidity at the oceanic surface during the last glacial maximum, *Nature*, 299, 688–691, 1982

Jouzel, J., Masson-Delmotte, V., Stiévenard, M. and Landais, A.: Rapid deuterium-excess changes in Greenland ice cores : a link between the ocean and the atmosphere, *Comptes Rendus - Geosci.*, 337, 957–969, 2005

- Kindler, P., Guillevic, M., Baumgartner, M., Schwander, J., Landais, A. and Leuenberger, M.: Temperature reconstruction from 10 to 120 kyr b2k from the NGRIP ice core, *Clim. Past*, 10(2), 887–902, 2014
- Klein, E.S., Cherry, J.E., Young, J., Noone, D., Leffler, A.J. and Welker, J.M.: Arctic cyclone water vapor isotopes support past sea ice retreat recorded in Greenland ice Sci. Rep.5: 10295-10300, 2015
- Kopec, B.G., Feng, X., Michel, F.A. and Posmentier, E.S., Influence of sea ice on Arctic precipitation *Proc. Natl. Acad. Sci.*, 113(1), 46–51, 2016
- Krinner, G., Genthon, C. and Jouzel, J.: GCM analysis of local influences on ice core δ signals. *Geophysical Research Letters*, 24(22), 2825-2828, 1997
- Landais, A., Barkan, E. and Luz, B.: Record of $\delta^{18}\text{O}$ and ^{17}O -excess in ice from Vostok Antarctica during the last 150,000 years, *Geophys. Res. Lett.* 35(2): L02709, 2008
- Landais, A., Steen-Larsen, H.C., Guillevic, M., Masson-Delmotte, V., Vinther and B., Winkler, R.: Triple isotopic composition of oxygen in surface snow and water vapor at NEEM (Greenland) *Geochim. Cosmochim. Acta*, 77, 304–316, 2012
- Landais, A., Winkler, R. and Prié, F., Triple Isotopic Composition of Oxygen in Water from Ice Cores, application note 30287, Thermo, 2014
- Langen, P.L. and Vinther, B.M.: Response in atmospheric circulation and sources of Greenland precipitation to glacial boundary conditions, *Clim. Dyn.*, 32(7–8), 1035–1054, 2009
- Leng W, von Dobeneck T, Bergmann F, Just J., Mulitza S., Chiessi C., St-Onge G. and Piper D.: Sedimentary and rock magnetic signatures and event scenarios of deglacial outburst floods from the Laurentian Channel Ice Stream, *Quat Sci Rev.*, 186, 27-46, doi:<https://doi.org/10.1016/j.quascirev.2018.01.016>, 2018
- Marcott, S.A., Clark, P.U., Padman, L., Klinkhammer, G.P., Springer, S.R., Liu, Z., Otto-Bliesner, B.L., Carlson, A.E., Ungerer, A., Padman, J., He, F., Cheng, J. and Schmittner, A., Ice-shelf collapse from subsurface warming as a trigger for Heinrich events, *Proc. Natl. Acad. Sci. U. S. A.*, 2011
- Marcott, S. A, Bauska, T.K., Buizert, C., Steig, E.J., Rosen, J.L., Cuffey, K.M., Fudge, T.J., Severinghaus, J.P., Ahn, J., Kalk, M.L., McConnell, J.R., Sowers, T., Taylor, K. C., White, J.W.C. and Brook, E.J.: Centennial-scale changes in the global carbon cycle during the last deglaciation, *Nature*, 514(7524), 616–619, 2013
- Markle, B.R., Steig, E. J., Buizert, C., Schoenemann, S. W., Bitz, C. M., Fudge, T. J., Pedro, J. B., Ding, Q., Jones, T.R., White, J.W.C. and Sowers, T.: Global atmospheric teleconnections during Dansgaard–Oeschger events, *Nat. Geosci.* 10(1), 36–40, 2016
- Martrat, B., Jimenez-Amat, P., Zahn, R. and Grimalt, J.O.: Similarities and dissimilarities between the last two deglaciations and interglaciations in the North Atlantic region, *Quat. Sci. Rev.*, 99, 122–134, 2014
- Masson-Delmotte, V., Landais, A., Stievenard, M., Cattani, O., Falourd, S., Jouzel, J., Johnsen, S. J. Dahl-Jensen, D., Sveinbjornsdottir, A., White, J.W.C., Popp, T. and Fischer, H., Holocene climatic changes in Greenland: Different deuterium excess signals at Greenland Ice Core Project (GRIP) and NorthGRIP, *J. Geophys. Res. D Atmos.*, 110(14), 1–13, 2005a

Masson-Delmotte, V., Jouzel, J., Landais, A., Stievenard, M., Johnsen, S.J., White, J.W.C., Werner, M., Sveinbjornsdottir, A. and Fuhrer, K.: GRIP deuterium excess reveals rapid and orbital-scale changes in Greenland moisture origin, *Science*, 309(5731), 118–121, 2005b

McManus, J. F., Francois, R., Gherardi, J.-M., Keigwin, L.D. and Brown-Leger S.: Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes, *Nature*, 428(6985), 834–837, 2004

NGRIP community members, High-resolution climate record of Northern Hemisphere climate extending into the last interglacial period, *Nature*, 431, 147–151, 2004

Partin, J. W., Cobb, K. M., Adkins, J. F., Clark, B. and Fernandez, D. P.: Millennial scale trends in west Pacific warm pool hydrology since the Last Glacial Maximum, *Nature*, 449, 452–455, 2007.

Peck, V. L., Hall, I.R., Zahn, R., Elderfield, H. and Grousset, F.: High resolution evidence for linkages between NW European ice sheet instability and Atlantic Meridional Overturning Circulation, *Earth Plan. Sci. Lett.*, 243, 476–488, 2006

Pfahl, S. and Sodemann, H.: What controls deuterium excess in global precipitation?, *Clim. Past* 10(2), 771–781, 2014

Rasmussen, S.O., Andersen, K.K., Svensson, A.M., Steffensen, J.P., Vinther, B.M., Clausen, H.B., Siggaard-Andersen, M.L., Johnsen, S.J., Larsen, L.B., Dahl-Jensen, D., Bigler, M., Roethlisberger, R., Fischer, H., Goto-Azuma, K., Hansson, M.E. and Ruth, U.: A new Greenland ice core chronology for the last glacial termination, *J. Geophys. Res. Atmos.*, 111(6), 1–16, 2006

Rasmussen, S.O., Seierstad, I.K., Andersen, K.K., Bigler, M. and Johnsen, S.J.: Synchronization of the NGRIP, GRIP, and GISP2 ice cores across MIS 2 and palaeoclimatic implications, *Quat. Sci. Res.*, 27, 18–28, 2008

Rhines, A. and Huybers, P.J.: Sea ice and dynamical controls on preindustrial and last glacial maximum accumulation in central Greenland *J. Clim.* 27(23): 8902–8917, 2014

Rhodes, R.H., Brook, E.J., Chiang, J. C. H., Blunier, T., Maselli, O.J., McConnell, J., Romanini, D. and Severinghaus, J.P.: Enhanced tropical methane production in response to iceberg discharge in the North Atlantic, *Science*, 348(6238), 1016–1019, 2015

Risi, C., Landais, A., Bony, S., Jouzel, J., Masson-Delmotte, V. and Vimeux, F.: Understanding the ¹⁷O excess glacial-interglacial variations in Vostok precipitation, *J. Geophys. Res. Atmos.*, 115(10), D10112, 2010

Russell, J., Vogel, H., Konecky, B.L., Bijaksana, S., Huang, Y., Melles, M., Wattrus, N., Costa, K. and King, J.W.: Glacial forcing of central Indonesian hydroclimate since 60,000 y BP., *Proc. Natl Acad. Sci.*, 111, 5100–5105, 2014

Schoenemann, S. W., Schauer, S. and Steig, E.J.: Measurement of SLAP2 and GISP d17O and proposed VSMOW-SLAP normalization for d17O and ¹⁷Oexcess, *Rapid Commun. Mass Spectrom.*, 27(5), 582–590, 2013

Schoenemann, S.W., Steig, E.J., Ding, Q., Markle, B.R. and Schauer, A.J.: Triple water-isotopologue record from WAIS Divide, Antarctica: Controls on glacial-interglacial changes in ¹⁷O-excess of precipitation, *J. Geophys. Res. Atmos.*, 119(14), 8741–8763, 2014

Schoenemann, S.W. and Steig, E.J.: Seasonal and spatial variations of ¹⁷O-excess and d-excess in Antarctic precipitation: Insights from an intermediate complexity isotope model, *J. Geophys. Res. Atmos.*, 121(19), 11,211-215,247, 2016

Severinghaus, J. P. and Brook, E.: Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice, *Science*, 286 (5441), 930-934, 1999

Stanford, D., Rohling, E. J., Bacon, S., Roberts, A.P., Grousset, F.E. and Bolshaw, M.: A new concept for the paleoceanographic evolution of Heinrich event 1 in the North Atlantic, *Quat. Sci. Rev.*, 30(9–10), 1047–1066, 2011

Steffensen, J. P., Andersen, K.K., Bigler, M., Clausen, H.B., Dahl-Jensen, D., Fischer, H., Goto-Azuma, K., Hansson, M., Johnsen, S.J., Jouzel, J., Masson-Delmotte, V., Popp, T., Rasmussen, S.O., Rothlisberger, R., Ruth, U., Stauffer, B., Siggaard-Andersen, M.-L., Sveinbjörnsdóttir, A. E., Svensson, A. and White, J.W.C., High-Resolution Greenland Ice Core Data Show Abrupt Climate Change Happens in Few Years *Science*, 321(5889), 680-684, 2008

Stenni, B., Buiron, D., Frezzotti, M., Albani, S., Barbante, C., Bard, E., Barnola, J.M., Baroni, M., Baumgartner, M., Bonazza, M., Capron, E., Castellano, E., Chappellaz, J., Delmonte, B., Falourd, S., Genoni, L., Iacumin, P., Jouzel, J., Kipfstuhl, S., Landais, A., Lemieux-Dudon, B., Maggi, V., Masson-Delmotte, V., Mazzola, C., Minster, B., Montagnat, M., Mulvaney, R., Narcisi, B., Oerter, H., Parrenin, F., Petit, J.R., Ritz, C., Scarchilli, C., Schilt, A., Schüpbach, S., Schwander, J., Selmo, E., Severi, M., Stocker, T.F. and Udisti, R.: Expression of the bipolar see-saw in Antarctic climate records during the last deglaciation, *Nat. Geosci.*, 4(1), 46-49, 2011

Stríkis, N.M., Chiessi, C.M., Cruz, F.W., Vuille, M., Cheng, H., De Souza Barreto, E.A., Mollenhauer, G., Kasten, S., Karmann, I., Edwards, R.L., Bernal, J.P. and Sales, H.D.R.: Timing and structure of Mega-SACZ events during Heinrich Stadial 1, *Geophys. Res. Lett.*, 42 (13), 5477–5484, 2015.

Svensson, A., Andersen, K.K., Bigler, M., Clausen, H.B., Dahl-Jensen, D., Davies, S.M., Johnsen, S.J., Muscheler, R., Parrenin, F., Rasmussen, S.O., Röthlisberger, R., Seierstad, I., Steffensen, J.P. and Vinther, B.M.: A 60 000 year Greenland stratigraphic ice core chronology, *Clim. Past*, 4 (1), 47–57, 2008

Toucanne, S., Zaragosi, S., Bourillet, J.-F., Marieu, V., Cremer, M., Kageyama, M., Van Vliet-Lanoë, B., Eynaud, F., Turon, J.-L. and Gibbard, P.-L.: The first estimation of Fleuve Manche palaeoriver discharge during the last deglaciation: Evidence for Fennoscandian ice sheet meltwater flow in the English Channel ca 20-18 ka ago, *Earth Planet. Sci. Lett.*, 290(3–4), 459–473, 2010

Toucanne, S., Soulet, G., Freslon, N., Silva Jacinto, R., Dennielou, B., Zaragosi, S., Eynaud, F., Bourillet, J.-F. and Bayon, G.: Millennial-scale fluctuations of the European Ice Sheet at the end of the last glacial, and their potential impact on global climate, *Quaternary Science Reviews*, 123, 113-133, 2015

Touzeau, A., Landais, A., Stenni, B., Uemura, R., Fukui, K., Fujita, S., Guilbaud, S., Ekaykin, A., Casado, M., Barkan, E., Luz, B., Magand, O., Teste, G., Le Meur, E., Baroni, M., Savarino, J., Bourgeois, I. and Risi, C.: Acquisition of isotopic composition for surface snow in East Antarctica and the links to climatic parameters, *The Cryosphere*, 10, 837–852, 2016

Uemura, R., Masson-Delmotte, V., Jouzel, J., Landais, A., Motoyama, H. and Stenni, B.: Ranges of moisture-source temperature estimated from Antarctic ice cores stable isotope records over glacial–interglacial cycles, *Clim. Past* 8(3), 1109–1125, 2012

Ullman, D.J., Carlson, A.E., Anslow, F.S., LeGrande, A.N. and Licciardi, J.M.: Laurentide ice-sheet instability during the last deglaciation, *Nat. Geosci.*, 8(7), 534-537, 2015

Vaughn, B., H., White, J.W.C., Delmotte, M., Trolier, M., Cattani, O. and Stievenard, M.: An automated system for hydrogen isotope analysis of water, *Chemical Geology Including Isotope Geoscience*, 152, 309-319, 1998

Veres, D., Bazin, L., Landais, A., Toyé Mahamadou Kele, H., Lemieux-Dudon, B., Parrenin, P., Martinerie, P., Blayo, E., Blunier, T., Capron, E., Chappellaz, J., Rasmussen, S.O., Severi, M., Svensson, A., Vinther, B. and Wolff, E.W., The Antarctic ice core chronology (AICC2012): an optimized multi-parameter and multi-site dating approach for the last 120 thousand years, *Clim. Past*, 9(4), 1733–1748, 2013

Wais Divide Members: Onset of deglacial warming in West Antarctica driven by local orbital forcing *Nature*, 500(7463), 440–444, 2013

[Werner et al., 2001](#)

Winkler, R., Landais, A., Sodemann, H., Dümbgen, L., Prié, F., Masson-Delmotte, V., Stenni, B. and Jouzel, J.: Deglaciation records of ^{17}O -excess in East Antarctica: Reliable reconstruction of oceanic normalized relative humidity from coastal sites, *Clim. Past*, 8(1), 1–16, 2012

Zhang, W., Wu, J., Wang, Y., Wang, Y., Cheng, H., Kong, X. and Duan, F. A: detailed East Asian monsoon history surrounding the ‘Mystery Interval’ derived from three Chinese speleothem records, *Quat. Res.*, 82(1), 154–163, 2014

Zhang, H., Griffiths, M.L., Huang, J., Cai, Y., Wang, C., Zhang, F., Cheng, H., Ning, Y., Hu, C. and Xie, S., Antarctic link with East Asian summer monsoon variability during the Heinrich Stadial–Bølling interstadial transition. *Earth Planet. Sci. Lett.* 453, 243-251, 2016

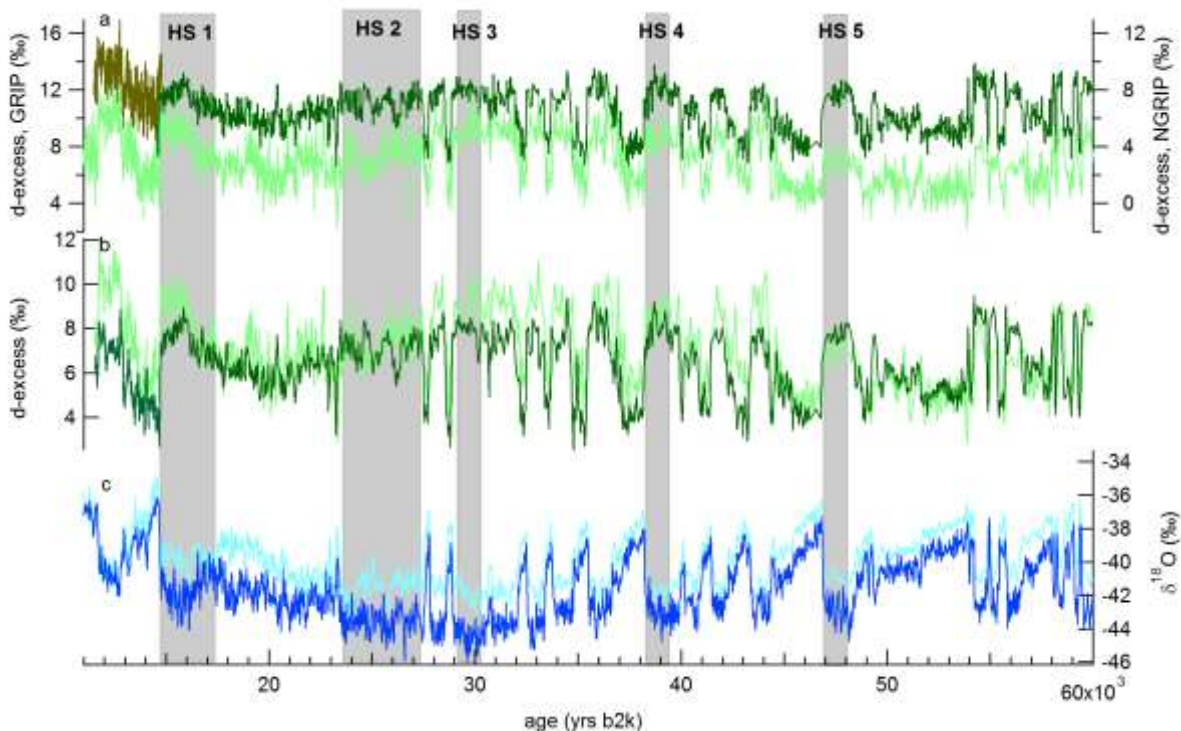


Figure 1: water stable isotope records ($\delta^{18}\text{O}$ and d-excess, in ‰) from GRIP and NGRIP ice cores reported on the GICC05 chronology (in thousands of years before year 2000 CE). a- d-excess from the NGRIP ice core (khaki: data obtained at INSTAAR SIL, Steffensen et al., 2008; dark green: data obtained at LSCE, this study); d-excess from the GRIP ice core (light green, Masson-Delmotte et al., 2005); b- d-excess from the NGRIP ice core after correction of the shift between INSTAAR SIL and LSCE (dark green) dataset, and d-excess from the GRIP ice core (light green); c- $\delta^{18}\text{O}$ from the NGRIP ice core (dark blue) datasets, $\delta^{18}\text{O}$ from the GRIP ice core (light blue). Grey intervals display Heinrich Stadials (HS).

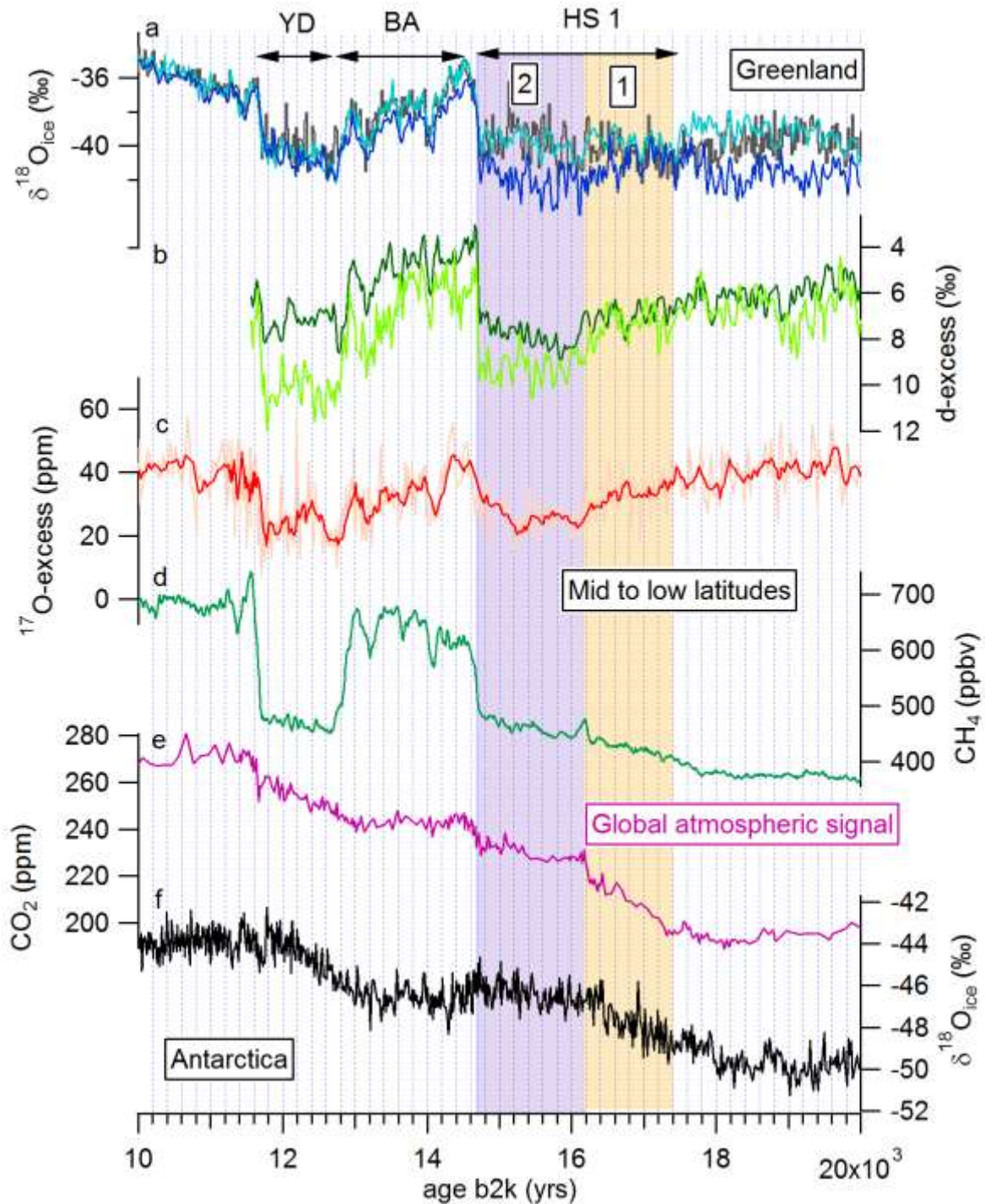


Figure 2: A synthesis of ice core records over the last deglaciation **on the respective GICC05 and AICC2012 timescales** with an identification of two phases (1, orange box and 2, purple box) within Heinrich Stadial 1 (HS1) as discussed in the text: we locate the transition between phases 1 and 2 at the timing of the sharp increase in CO₂ and CH₄ concentrations, both being global atmospheric composition signals. The Younger Dryas (YD) and Bølling-Allerød (BA) periods are also indicated.

a-GRIP, NGRIP and GISP2 $\delta^{18}\text{O}$ (light blue, dark blue and black respectively (Grootes et al., 1993; NGRIP community members, 2004) interpolated at a 20 years resolution; b-GRIP and NGRIP d-excess (light and

dark green respectively: Jouzel et al., 2005, this study) interpolated at a 20 years resolution;c-NGRIP ^{17}O -excess (orange curve shows the original series and the red curve the 5 years running average, this study);d-WAIS Divide CH_4 (Rhodes et al., 2015);e-WAIS Divide CO_2 (Marcott et al., 2013);f-EPICA Dronning Maud Land (EDML) $\delta^{18}\text{O}_{\text{ice}}$ (EPICA community members, 2006)

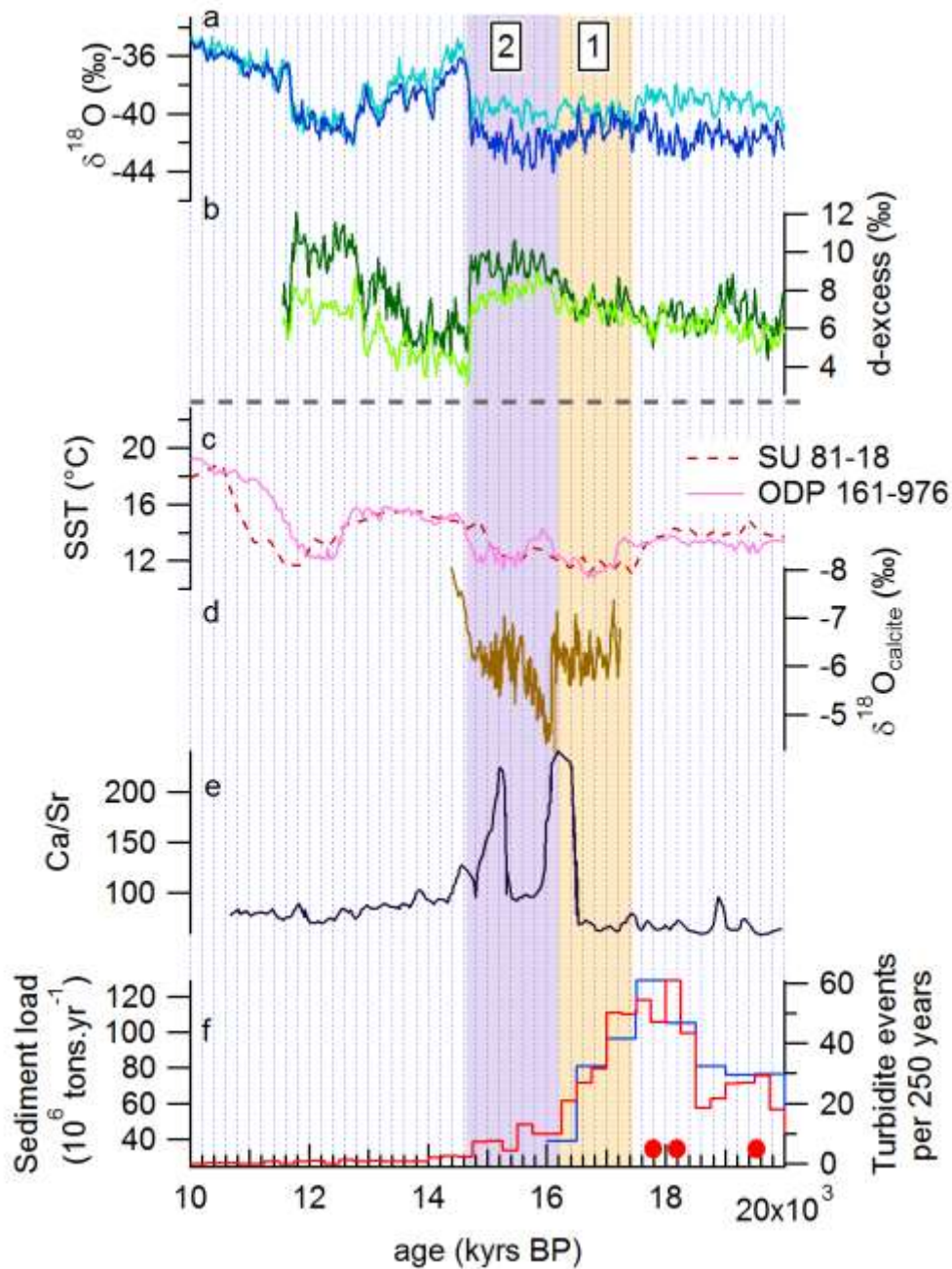


Figure 3: The sequence of Phase 1 and Phase 2 of Heinrich Stadial 1 identified in Greenland records and in proxy records of North Atlantic SST, IRD events, and changes in East Asian hydroclimate. a- NGRIP (dark blue) and GRIP (light blue) $\delta^{18}\text{O}$ records; b- NGRIP (dark green) and GRIP (light green) d-excess records; c- Sea surface temperature (SST) for North Atlantic cores SU 81-18 (Bard et al., 2000) and ODP 161-976 (Martrat et al., 2014); d- Calcite $\delta^{18}\text{O}$ of Hulu cave (China, Zhang et al., 2014); e- Ca/Sr from site U1308 in the IRD belt (Hodell et al., 2019) as signature from strong iceberg discharges from the Laurentide ice sheet; f- Indications for Channel River sediment load (blue, sediment load; red, turbidite frequency) (Toucanne et al., 2010; 2015) as signature for meltwater input from European side.

The 3 red circles indicate plumite layers resulting from outburst floods on the Eastern Canadian margin (Leng et al., 2018), i.e. meltwater arrival from the North America side in the absence of strong iceberg discharge.

The dashed horizontal line separates the ice core records on the GICC05 timescale from non ice core records on their own timescales.