Interactive comment on "Multi-proxy fingerprint of Heinrich event 4 in Greenland ice core records" by M. Guillevic et al.

Replies to Anonymous Referee #1 and #2

We thank the reviewers for taking the time to review our manuscript and for insightful and constructive comments. Hereafter, the comments from the referees are in normal text and the replies in bold text. The revised manuscript is attached at the end with major modifications in blue.

Anonymous Referee #1

This paper presents a multi-proxy dataset across a set of Dansgaard-Oeschger (DO) events. A particular novelty is that it presents the first high resolution ^{17}O-excess data across a series of D-O events. The authors attempt to make the case that they can discern different substages within GS-9, and they suggest that one of these stages is associated with iceberg discharges. If this were correct, it would allow a clearer definition of how Heinrich Events fit into the DO paradigm.

Firstly, I would like to congratulate the authors on producing the ^{17}O-excess and new $\delta^{18}O_{atm}$ data that are the basis for the paper: both are impressive datasets. However the way the authors use the data requires a chain of guesswork and circumstantial evidence that is not sufficient to support the bold claims they make, not least in the title of the paper.

>>The manuscript has been strongly rewritten as well as the title and conclusions to tone them down.

The first issue is with the way they interpret some of the data themselves. ^{17}O data can be interpreted through the use of climate models. However, as with deuterium excess, it is already obvious that there is no simple explanation for any change, as the very vague interpretation in the last paragraph of section 3.1 indicates. I realise this is not simple as we do not yet have a full understanding of what factors may control ^{17}O-excess.

However, as an example of the confusion in this interpretation, on page 1188 at line 11 the authors describe scenarios that "are in agreement with a southward shift of the vapour source ... as suggested by the high d-excess", but at line 26 a southward shift is required to cause low ^{17}O-excess but constant d-excess. As a result, the interpretations that follow are highly speculative and not certain enough to support the broad inferences in the paper.

>>It has been already shown that d-excess is very strongly influenced by local temperature or temperature along the trajectory while ^{17}O-excess is much less affected by temperature (except in cold regions of Antarctica [Winkler et al., 2012; Landais et al., J. glacio., 2012; Schoenemann et al., 2014]) but more sensitive to humidity at the source and recycling effects. As a consequence, there can be a decoupling between d-excess, $\delta^{18}O$ and ^{17}O-excess. The fact that this decoupling is evidenced here for the first time is actually a new result and we have tried to highlight this more as well as to provide more clear explanations for this effect. In particular, the decoupling between $\delta^{18}O$ and ^{17}O-excess shows that condensation temperature is not the main driver for the ^{17}O-excess changes over GS-9.

Perhaps even more worrying is the interpretation of MSA as a sea ice proxy: in support of this the authors cite a recent review (Abram et al 2013) and a paper by O'Dwyer that showed a weak but statistically significant negative correlation (less sea ice = more MSA) at Svalbard. However the O'Dwyer paper specifically pointed out that the only work (and to my knowledge still the only work) in Greenland had suggested a positive correlation with sea ice (though this is very tentative in the Legrand paper cited). The Abram review cited the same papers and certainly did not support the use of a negative correlation for Greenland. I agree that it is plausible that MSA might be a sea ice proxy in Greenland (just as it is plausible, but completely unproven for Greenland, that sea salt, which would tell a different story, may be a sea ice proxy). However this is no basis for the strong conclusions that follow; on the showing of the only published work we are aware of, the conclusion would actually be the opposite. I don't advocate the opposite conclusion, and other factors have to be taken into account (such as the effect of lower accumulation rates on concentration, of higher dust on MSA preservation, and of changing transport paths and source regions). However, it does mean that the strong interpretation in terms of sea ice is simply not currently supportable. (As an aside I wonder why the authors used GISP2 MSA, when MSA data were certainly collected at NEEM).
>>We have entirely removed all the MSA-related discussion in the revised manuscript. While the review from Abram et al., 2013 presented this proxy as potentially promising without giving any firm conclusion, we agree that we have used this light suggestion with too much enthusiasm. This is indeed a complex proxy influenced by biological production of DSM by algae in the sea-ice marginal zone, oxidation of this product and finally transport and preservation in the snow. We are not aware of any published MSA data from the NEEM ice core so far. For the Greenland ice sheet, they are a few studies presenting MSA measured in the GRIP (Legrand et al., 1997), GISP2 (Saltzman et al., 1997) and NGRIP ice core (Jonsell et al., 2007). For all these records, we have calculated the MSA flux and the pattern during GS-9 is not modified compared to the concentration variations, because of a constant accumulation rate during all GS-9 for these three cores. In the same way, the high resolution dust record from NGRIP show a stable level during the entire GS-9, and thus the MSA variations should not be influenced by dust-induced varying preservation in the snow. We agree with Reviewer #1 that using MSA would require a much longer discussion for conclusions that would not be definitive concerning this proxy. We have thus decided to remove this proxy from our study.

The second issue is simply with the way the data are described, particularly the 17O data on which the story is hung. The description in the first paragraph of section 3.1 is very leading for the reader, essentially stating that 17O is high in GI and low in GS – except that the relationship breaks down for half the record!! If one looks at Fig 3 the relationship is really not that obvious: I'd like to know what the correlation coefficient between $\delta^{18}O$ and ^{17}O-excess actually is across the whole record. Having made this rather wishful statement, the authors then try to interpret the departures from it, in particular picking out more detailed changes in GS-9. Clearly if the initial GI/GS difference is not really robust, then the discussion of departures from it becomes difficult. I can see that there are interesting signals there, but so there are in GS-8, where I could also argue for 3 stages and yet no-one proposes an HE there.
>>The reviewer is right and we have tried to strengthen this part. First, we have performed new ^{17}O-excess measurements over GS-8 and 9 to check that GS-9 has really a specific

fingerprint: during GS-8, the ^{17}O-excess level is relatively low even if some scattering is observed. Figure 3 now shows the whole ^{17}O-excess record which encompasses GI-7 to GI-13. Over this 15 000 years record, we have a positive correlation between ^{17}O-excess and $\delta^{18}O$. The calculated correlation coefficient (0.2) is still quite weak because of the scattering in the ^{17}O-excess data. The strongest exception from this general pattern is over GS-9 when ^{17}O-excess shows significant and long term variations (several centuries) that are not seen in the $\delta^{18}O$ record. We do not exclude that this behavior can be observed elsewhere but the high number of measurements performed over this period (additional measurements for this revised version) excludes that this is only noise in the data. Moreover, we have also combined the ^{17}O-excess measurements with other proxies of lower latitudes in ice core ($\delta^{18}O_{atm}$, CH_4, δD of CH_4) to confirm this signal and its interpretation. We have made this approach clearer in the revised version.

Thirdly, while I am happy to agree that HE occur only after GS have taken hold, the link from the authors's substage 2 to the iceberg discharges is completely guesswork. I can see nothing (other than that the iceberg discharges appear to be in mid-stadial) to support this. Section 3.3 seems very speculative. Finally the attempt to implicate HE5 as having a similar structure, in the absence of the 17O data, is unconvincing. As stated above I think we could find 3 phases in any section of the core if one subdivides carefully enough. In this case the classification seems to be largely on the basis of MSA, which shows numerous changes as large as the ones separating the notional GS13-1,2,3 and yet these are not commented on.
>>We have measured new ^{17}O-excess data covering GI-12 to GI-13 but removed the MSA data. Based on the combined proxies ^{17}O-excess, methane mixing ratio and $\delta^{18}O_{atm}$, it is still possible to suggest a unique time when these proxies clearly show synchronous ^{17}O-excess decrease, methane mixing ratio increase and $\delta^{18}O_{atm}$ increase, as at the onset of phase 2 during GS-9. We therefore still propose this timing as a possible onset of the iceberg delivery H5, while our data prevent us to make any tentative identification of the boundary between phase 2 and phase 3.

In summary I think this paper is a long way from being suitable for publication in CP.
The new data are fascinating and hard-won. They deserve to be published. It would be reasonable to discuss the changes within GS9 but this needs to be done with much less certainty about their significance, and the link to the iceberg discharges must be considered very tentative (indeed probably this can be no more than a suggestion, certainly not the basis for the title). I am afraid this will leave a paper that is little more than a data presentation and with a very different focus, but it's really hard to see that the present analysis is supportable. I will suggest major revisions, but I admit that I have doubts that even a major revision can produce a publishable paper with the present focus (i.e. on Heinrich events).
>>We have changed the principal focus and the title. The link between Heinrich events and Greenland Stadials is still presented as a motivation but no definitive conclusions are presented, only perspectives.

Detailed comments:
Page 1180, line 3. "coincide" is a bad word, as the whole point of the paper is that the HE and

the stadial don't coincide (implying causality), rather the HE happen to occur during some GS.
>>We have changed this word for "appear related to"

Page 1180, line 10: infirmed is not correct English. "disproven" would be a possible replacement.
>>The abstract has been rewritten and this sentence has been removed.

The abstract will need to be toned down. The three phases are nowhere near as obvious as the paper suggests, and the link to the HE is tenuous.
>>The abstract has been rewritten and toned done.

Page 1183, line 1-2. "stricto senso" is not a phrase I know, do you mean "strictu senso"?
>>This typing mistake has been corrected.

Page 1185, para 1. It would be worth here clarifying that you use the GS numbering as recommended by INTIMATE and as developed by various authors, but that this is different to the C-numbering of cold phases sometimes used in the marine community (as in Rousseau et al 2006, numbering is one different). Given that this paper is aimed partly at marine people, you need to make this clear.
>>The manuscript has been modified accordingly.

Section 3.1. A better description than the one given in item (i) is needed. Indeed the reasons why ^{17}O-excess deviates so much from δ^{18}O and deuterium-excess across the entire record might be a better focus for the paper.
>>We have provided a stronger background on ^{17}O-excess in the introduction part and much more discussed the ^{17}O-excess – δ^{18}O decoupling observed during GS-9.

Page 1189 – as explained already the use of MSA as a sea ice indicator in Greenland is not supported by the references shown, and is probably not at this point supportable. Again a much more complex discussion will be needed if this is to be used in this paper at all.
>>All the discussion implying MSA has been removed from the paper.

Page 1190, line 10. In discussing the effects of oxidation on methane between GI and GS, you might want to cite [Levine, J. G., et al (2012), Controls on the tropospheric oxidizing capacity during an idealized Dansgaard-Oeschger event, and their implications for the rapid rises in atmospheric methane during the last glacial period, Geophys. Res. Lett., 39, L12805, doi:10.1029/2012gl051866.]
>>This reference has been added to the manuscript.

Section 3.4. I think it would be better to leave this discussion of HE5 out altogether. It is not convincing.
>>Since we have now ^{17}O-excess data covering GS-13, we have kept only the observation that one time point from GS-13 presents the same features as the onset of phase 2 in GS-9: methane abrupt increase, $\delta^{18}O_{atm}$ gradual increase, abrupt ^{17}O-excess decrease and stable δ^{18}O.

I did not check all refs, but Eynaud et al 2009 has a strange bug in the author list (F., ni, : : :). I also assume the odd page numbers at the end of each ref are an artefact of the editing and will need to be removed in future versions.
>>This reference has been corrected. The bug was due to the accents on the co-author name "Sanchez Goñi". Indeed, the page numbers at the end of each citation are an artefact of the editing.

Fig 4: You use the notation GI and GS in the text so better to use that on the figure, rather tha DO (for GI) and GS.
>>This has been changed.

Fig. 6 is really hard to follow because you have made no attempt at synchronisation. What the reader is trying to do is see if the IRD is synchronous between sites to judge the importance of your observation that the onset of cold is not synchronous. However this is really hard to do on this presentation. I understand the reluctance to make a synchronisation if it is not supported by data but perhaps you could at least draw some dashed vertical lines.
>>We agree with the reviewer that this would be much more practical. Unfortunately, we have not so far found any climate independent proxy (such as beryllium isotopes or volcanic ash layers) that would be available for these cores. Since the ice core data presented in this study suggest no synchronous climate changes during GS-9 in between high and middle-low latitudes, we have on purpose made no climate-proxy-based synchronisation in between these marine sediments records that originate from different latitudes (such as based on a temperature proxy as it is usually done). Taking into account the chronology uncertainties in between marine cores and in between marine sediment records and ice core records, there is actually no evidence for synchronous changes, as there is no chronology-based evidence for non-synchronous changes. This approach is not new: for example Blaaw et al. showed that for MIS3, there were no evidence for synchronous climatic changes in between the NGRIP δ^{18}O records and the lake productivity records from Lake Les Echets, France (45°54'N, 48°56'E, 275 m above sea level):
"Even with the highest-resolution dated age models (67 radiocarbon and IRSL dates for Les Échets, multi-proxy annual layer counting for Greenland; Andersen et al., 2006), chronological uncertainties are currently too high to resolve through independent chronologies whether last glacial D-O climate events were simultaneous, or even related, between Greenland and Western Europe."

Anonymous Referee #2

General comments

There are two points, where the paper should go deeper in its discussion and beyond its current status. The first one pertains the discussion of the ^{17}O-excess. ^{17}O-excess is not yet a well established addition to the ice core parameters previously studied and deserves a more detailed introduction and detailed explanations in the manuscript. For instance on page 1184 the background given on this parameter is quite vague. Given that the ^{17}O-excess represents the central parameter in the discussion, the information hidden in the ^{17}O-excess and the underlying physics should be explained in more detail on page 1184. Again in section 3.1 the information gained from the ^{17}O-excess is just stated (and a reference to the paper by Risi et al. is given) without explaining in more detail, why exactly this information (for example on water vapor recycling) can be gleaned from this parameter. This should be explained in significantly more detail in the revised version.

>>We have substantially given more details on the physics and already published articles about ^{17}O-excess in polar environments (Greenland and Antarctica), both in the introduction and in the discussion part.

The second general comment is more a pledge than a request. The new results by Guillevic et al. suggest that the H event itself (related to a collapse of one or several ice sheets) is not the trigger for the stadial cooling as it comes later in the game. I would like to read the authors' opinion, on what initiated the Stadial in the first place in the Discussion and/or Conclusion chapter of the manuscript. I am sure they have thought about this thoroughly and may be able to make an educated guess.

>>We have added a paragraph to discuss this (Section 3.4 in the revised manuscript, 2nd paragraph).

Finally, not all of the tracers used in this study provide unambiguous information on certain aspects of the climate system. Alternative ways of interpretation (for example on sea ice extent) should be discussed and speculations made in the paper more clearly qualified as such.

>>We have suppressed all reference to the MSA proxy. We have toned down the discussion in the manuscript.

Specific comments

P1180, L10: "could, however, never"
P1191, L5: "temperature, unveiled"
P1181, L12: "conditions, while"
P1181, L27: "Meridional Overturning Circulation"
P1182, L2: "Atlantic marine sediment cores"
P1182, L11: "in a sediment core"
>>All these changes have been made in the revised manuscript.

P1183, 1st paragraph: A few more words to motivate this discussion would be helpful.

You may want to cite in this respect the recent paper by Skinner et al. 2014 in PNAS.
>>We have delayed this paragraph.

P1183 2nd paragraph: This discussion goes into a lot of detail on the accuracy of the chronologies etc. This may be justified but an introductory sentence, why the following discussion is important at this point of the paper, would be helpful. You may want to cite the recent paper by Jo et al., 2014 in Nature and an older one by Bozbiyik et al, 2011 in Climate of the Past on hydrological responses to ITCZ changes.
>>We have shortened this paragraph. We have added the suggested references.

P1183, L15: "(ITCZ), which:" **>>Manuscript modified accordingly.**
P1183, L17: "without Heinrich events." **>>Manuscript modified accordingly.**

P1184, last paragraph: Please provide more detail on ^{17}O-excess
>>A full paragraph on ^{17}O-excess has been added.

P1185, L1-2: "We have performed the first ^{17}O-excess measurements on the NEEM core (Dahl-Jensen)" **>>Manuscript modified accordingly.**
P1185 and throughout: exchange "homemade standard" with "in-house standard" or "working standard" **>>Manuscript modified accordingly.**
P1186, L3. "With δ^{18}O varying between -43 and -58 permille, we used working standards at around" **>>Manuscript modified accordingly.**
P1186, L11 and L20: "two point calibration" **>>Manuscript modified accordingly.**
P1186, L22: "necessarily" **>>Manuscript modified accordingly.**

P1187 2nd paragraph: Explain in a little more detail, how this correction is done
>>We have added more details on the $\delta^{18}O_{atm}$ corrections.

Section 3.1 Explain in sufficient detail the arguments concerning what information is hidden in the ^{17}O-excess
>>We have significantly lengthened this section, to provide more explanation and also discussion about the ^{17}O-excess variations.

P1188, L11& L26: "southward shift" **>>Manuscript modified accordingly.**
P1188 last paragraph: Again, the ^{17}O-excess argument needs more explanation.
>>We have provided more explanation.

P1189, L22 "shifts of the ITCZ" **>>Manuscript modified accordingly.**

P1189, 2nd paragraph: The use of MSA as a sea ice tracer for the Arctic is not really established yet. Accordingly, this discussion should be more clearly qualified as speculation and alternative ways to interpret this record should be explored.
>>We have removed the MSA from this study.

P1189, L29: “with a potential clathrate release” **>>Manuscript modified accordingly.**

P1190, L1: I agree with this statement, but you should explain the reader in more detail why.
>>About the potential clathrate release: we have additionally used the $\delta^{13}C-CH_4$ data from Möller et al., 2013, the $\delta D-CH_4$ data from Bock et al., 2010, as well as the model study from Boch et al., 2012. Comparing the expected signature of a clathrate release in ice cores and the observed isotopic signature, we actually cannot rule out this possibility. We have added a paragraph about these new findings. We have also strengthened the discussion on the following stable plateau that we propose to be due to new source(s) activation.

P1190, L6: “such a southward ITCZ shift” **>>Manuscript modified accordingly.**
P1191, L3: “(phase 2), which” **>>Manuscript modified accordingly.**
P1191, L10: “suggest a globally” **>>Manuscript modified accordingly.**
P1191, L19: “2008), while” **>>Manuscript modified accordingly.**

P1191, L27: Please provide a reference
>>The references corresponding to the publication of the marine core results have been added: Auffret et al., 2002 and Sanchez-Goni et al., 2008.

P1192, L1” “measurements” **>>Manuscript modified accordingly.**
P1192, L3: “As discussed above, we argue for a” **>>Manuscript modified accordingly.**
P1192 L23: “of events in phase 3: **>>Manuscript modified accordingly.**

Section 3.4: I would suggest to place 3.4 immediately after 3.2 (potentially without an extra header) and only after that go to the discussion in 3.3
>>We have changed the manuscript accordingly. Moreover, the section about the ice core proxies covering GS-13 has been significantly shortened.

P1193, P10: “ends with the MSA”
>>MSA has been removed from this manuscript.
P1193, P23: “phases 1 and 3, a hypothesis to be tested in future studies.”
>>We have shortened this section and therefore removed this discussion part.

P1194, L7: “may mark the maximum” **>>Manuscript changed accordingly**.
P1194, L25: Is “trigger” the right expression here? **>>This paragraph has been removed.**
P1195, L7: “sea ice extent” **>>Manuscript changed accordingly.**
P1196, L12-13: “: : :from warm subsurface waters in the North Atlantic Drift, to where” **>>Manuscript changed accordingly.**
P1196, L20: “state, while” **>>Manuscript changed accordingly.**

P1196 last paragraph: Can you speculate also on the role of sea ice in the Baffin Bay and Labrador Sea, which appear to be of more importance for water vapor transport to Western Greenland.

>>The MARGO project 2005 and several model simulations of sea ice cover (e.g., Li et al., 2010, Flückinger et al., 2008) suggest that the Labrador Sea and the Baffin Bay were constantly sea-ice covered during the LGM. The model study of Van Meerbeck et al., 2009, using the LOVECLIM model, produced a reduced (compared to LGM) sea ice cover in the Labrador sea in both GS and GI, while the Baffin Bay was sea ice covered. Model simulations from Li et al., 2010 for the LGM state suggested that Greenland (including West Greenland) temperature and accumulation were much more sensitive to Nordic Seas sea ice cover decrease compared to West Atlantic (including Labrador Sea) sea ice cover decrease. We therefore think that sea ice cover in the Labrador did not play an important role for West Greenland climate.

P1197, L16: "between the constantly cold Greenland"
>>Manuscript changed accordingly.
P1198, L22: "approach by matching" **>>Manuscript changed accordingly.**
P1198, L25: "using the mid-slope of each GI onset as match points" **>>Manuscript changed accordingly.**

Fig. 1: It may be helpful to indicate (e.g. by lettering) from which ice sheet which H event originated in this figure.
>>We have modified the color code for the colored areas, so that it now reflects the origin of the icebergs: grey for major icebergs deliveries of mostly Laurentide origin, yellow for the minor events of mostly Laurentide origin, pink for the events of mostly European origin, and blue for the episodes of low sea surface salinity recorded in the Celtic margin.

Fig. 4: This central figure should be two columns wide in the final print. It was hard to discern things in the CPD version.
>>This figure is now 2 columns wide.

Fig. 5: What caused the cooling into phase 1?
>>A new paragraph has been added to Section 3.4 (in the revised version) to discuss this.

Fig. 6: Add numbering of the DO events. **>>This has been done**.

Fig. 7: I don't understand the comment in line 5 on the use of a PICARRO instrument.
>>When measuring methane mixing ratio in the NEEM ice core at the drilling site, three instruments were connected in series to successively measure this: a non-destructive laser technique from LGGE (Laboratoire de Glaciologie et Geophysique de l'Environnement, France), a second non-destructive technique using a laser instrument originating from the PICARRO company, run by CIC (Centre for Ice and Climate, Denmark), and finally a gas chromatography system from the University of Bern. More details can be found in Stowasser et al., 2012 and Chappellaz et al., 2013.

Manuscript prepared for Clim. Past
with version 5.0 of the LaTeX class copernicus.cls.
Date: 20 July 2014

Evidence for a 3 phase sequence during Heinrich Stadial 4 using a multiproxy approach based on Greenland ice core records

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Abstract. Glacial climate was characterised by two types of abrupt events. Greenland ice cores record Dansgaard–Oeschger events, marked by abrupt warming in-between cold, stadial phases. Six of these stadials appear related to major Heinrich events (HE), identified from ice-rafted debris (IRD) and large excursions in carbon and oxygen stable isotopic ratios in North Atlantic deep sea sediments, documenting major ice sheet collapse events. This finding has led to the paradigm that glacial cold events are induced by the response of the Atlantic Meridional Overturning Circulation to such massive freshwater inputs, supported by sensitivity studies conducted with climate models of various complexities. This mechanism implies synchronous Greenland temperature and lower latitude hydrological changes. To investigate the sequence of events between climate changes at low latitudes and in Greenland, we provide here the first ^{17}O-excess record from a Greenland ice core during Dansgaard-Oeschger events 7 to 13, encompassing H4 and H5. Combined with other ice core proxy records, our new ^{17}O-excess dataset demonstrates that Stadial 9 consists of three phases, characterised first by Greenland cooling during 550 ± 60 years (as shown by markers of Greenland temperature δ^{18}O and δ^{15}N), followed by a specific lower latitude fingerprint as identified from several proxy records (abrupt decrease in ^{17}O-excess, increase in CO_2 and methane mixing ratio, heavier δD-CH_4 and $\delta^{18}O_{\mathrm{atm}}$), lasting 740 ± 60 years, itself ending approximately 390 ± 50 years prior to abrupt Greenland warming. We hypothesise that this lower latitude signal may be the fingerprint of Heinrich event 4 in Greenland ice cores. The proposed decoupling between stable cold Greenland temperature and low latitude climate variability provides new targets for benchmarking climate model simulations and testing mechanisms associated with millennial variability.

1 Introduction

Glacial climate is characterised by millennial variability, recorded with specific expressions in different archives and at different latitudes (Voelker, 2002; Clement and Peterson, 2008). Greenland ice core records of ice δ^{18}O, a qualitative proxy of air temperature, unveiled at high resolution the succession of cold phases (Greenland Stadials, GS)

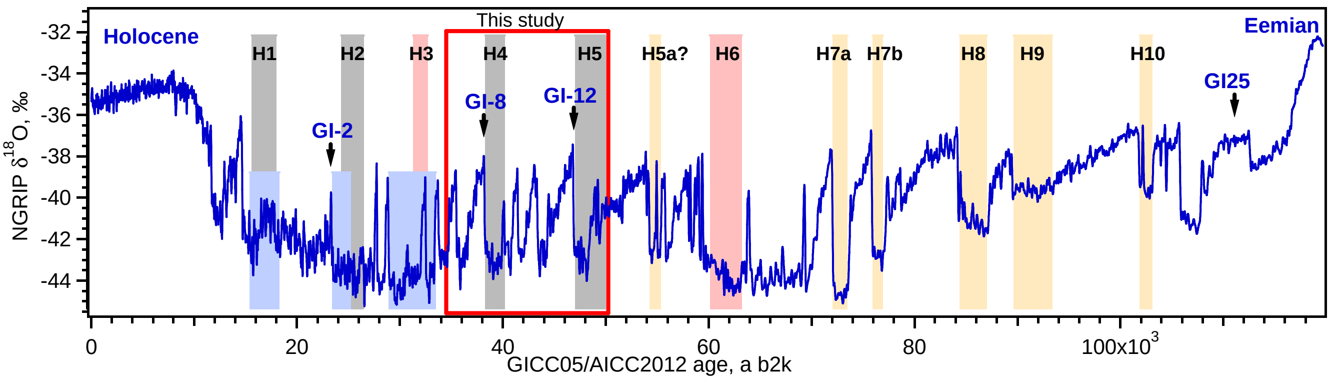

Fig. 1. Dansgaard–Oeschger (the warm phases being marked as Greenland Interstadials, GI) and Heinrich (H) events during the last glacial period. Blue line: NGRIP δ^{18}O, ‰ (NGRIP members, 2004), on the GICC05 timescale back to 60 ka b2k (Appendix A1) and AICC2012 beyond (Bazin et al., 2013; Veres et al., 2013). Coloured areas: position and duration of Heinrich events recorded in marine cores from the North Atlantic. Grey: H1, H2, H4 and H5: major events of mainly Laurentide origin; pink: H3 and H6, major events of mainly European origin (see text for references; timing on the GICC05 timescale according to Sánchez Goñi and Harrison, 2010). Blue: periods of low salinity corresponding to fresh water input on the Celtic Margin (Eynaud et al., 2012). Yellow: H5a and H7 to H10, minor IRD events of Laurentide origin recorded in the West Atlantic (Rashid et al., 2003; Rasmussen et al., 2003).

and warm phases (Greenland Interstadials, GI) forming the 25 Dansgaard–Oeschger events (DO) of the last glacial period (NGRIP members, 2004). DO events are recorded through climatic and environmental changes in other North Atlantic/European archives such as speleothems (Genty et al., 2010; Boch et al., 2011), pollen and marine bioindicators from North Atlantic marine cores (e.g. Voelker, 2002; Sánchez Goñi and Harrison, 2010), with interstadials being characterised in Western Europe by warmer and more humid conditions, while stadials are associated with a dry and cold climate. The tropics also exhibit a fingerprint of DO events as suggested e.g. by (i) variations in monsoon strength (e.g. recorded in speleothem growth rate and calcite δ^{18}O, Wang et al., 2001), (ii) changes in the atmospheric methane concentration (as measured in ice cores, Chappellaz et al., 1993, 2013) with its main source located at low latitudes during glacial periods (e.g. Baumgartner et al., 2012), or (iii) changes in the isotopic composition of atmospheric oxygen $\delta^{18}\mathrm{O}_{\mathrm{atm}}$ measured in air from ice cores, (Landais et al., 2007; Severinghaus et al., 2009), reflecting at this time scale changes in the low latitude water cycle and global biosphere productivity (Bender et al., 1994b). Such variations in the low latitude climate are probably due to shifts in the Intertropical Convergence Zone (ITCZ, e.g. Peterson et al., 2000). The bipolar seesaw identified between each Greenland DO event and its Antarctic counterpart (EPICA community members, 2006; Capron et al., 2010) supports a key role of reorganisations of the Atlantic Ocean inter-hemispheric heat transport (Stocker and Johnsen, 2003), likely associated with strong variations of the Atlantic Meridional Overturning Circulation (AMOC) intensity.

In addition to palaeoclimate records showing this DO succession, a prominent feature identified in North Atlantic marine sediment cores is the occurrence of ice-rafted debris (IRD, Ruddiman, 1977). These IRD are interpreted as the signature of massive iceberg discharges in the North Atlantic Ocean originating from the Laurentide, Icelandic, British–Irish and Fennoscandian ice sheets (e.g., Heinrich, 1988; Bond et al., 1993). Six such Heinrich events (HE, numbered H1 to H6, Fig. 1) have been unambiguously identified during GS phases of the last glacial period (e.g., Bond et al., 1992; Hemming, 2004). Minor events identified in one or a few cores have additionally been reported (H5a (Rashid et al., 2003), H7a, H7b, H8 to H10 (Rasmussen et al., 2003), Fig. 1). In addition to the IRD layers, periods of low surface salinity likely due to enhanced fresh-water fluxes have been evidenced in a sediment core from the Celtic margin (Fig. 1 blue areas, Eynaud et al., 2012). Studies on the composition of each IRD layer have demonstrated the dominant Laurentide origin of H2, H4 and H5, while H3 and H6 are mainly due to icebergs delivery from European ice sheets (Grousset et al., 1993, 2000; Gwiazda et al., 1996; Hemming et al., 1998; Snoeckx et al., 1999; Jullien et al., 2006). H1 results from the collapse of several ice sheets (Stanford et al., 2011).

The early proposed synchronicity of HEs in marine cores and GS in Greenland ice cores (e.g. Bond et al., 1993) have led to the paradigm that Greenland stadial/interstadial variability is related to freshwater-induced changes in AMOC. Indeed, the response of climate models of different complexities to freshwater perturbations bears similarities with palaeoclimate observations (e.g., Kageyama et al., 2013). Such GS associated to HEs are also called Heinrich stadials. Unfortunately, uncertainties associated with marine and ice core chronologies have so far prevented the determination of the exact timing of Heinrich events with respect to GS (Sánchez Goñi and Harrison, 2010; Austin and Hibbert, 2012). In addition, several lines of evidence suggest that HE are shorter than the corresponding GS (Peters et al., 2008; Roche et al., 2004) and occur after the AMOC entered a weakening trend (Flückiger et al., 2006; Marcott et al., 2011). Therefore the mechanisms relating iceberg discharge, low latitude climate change and Greenland temperature change during stadials remain debated (Hemming, 2004; Clement and Peterson, 2008; Mulitza et al., 2008).

The impact of HEs is not limited to the North Atlantic but has a signature in mid to low latitude palaeoclimate records. For instance speleothem growth rate from South America and Asia and their calcite isotopes indicate a southward displacement of the intertropical convergence zone (ITCZ) during stadials compared to interstadials (e.g., Jo et al., 2014), most pronounced during Heinrich stadials (Kanner et al., 2012; Wang et al., 2007; Mosblech et al., 2012). Model simulations indeed produce in a few years a southward displacement of the ITCZ over the Atlantic Ocean and its margins in response to North Atlantic cooling (e.g. Chiang et al., 2008; Bozbiyik et al., 2011; Cvijanovic and Chiang, 2013). Speleothems from low latitudes have the advantage to provide the best absolute chronology amongst palaeo archives so far, through uranium-thorium dating of the calcite. Associated uncertainties can be as low as $\sim \pm 1\,\%$ (2σ) of the absolute age for the Hulu cave record (Wang et al., 2001). Calcite δ^{18}O from South-East Asia speleothems and $\delta^{18}\mathrm{O}_{\mathrm{atm}}$ variations registered in the air entrapped from ice cores are both mostly controlled by latitudinal shifts of the ITCZ and associated changes in the isotopic composition of tropical precipitation. This common driving climatic mechanism therefore offers a direct link between speleothem and ice core archives. However, climate-independent markers have not yet been identified that would allow to synchronise speleothems to other archives from different latitudes such as marine or ice cores, preventing a precise comparison of the timing of events between high, mid and low latitudes associated with Heinrich events.

In Greenland ice cores, DO events are well dated thanks to annual layer counting. For the GICC05 timescale used here, absolute ages are estimated with $\sim 3\,\%$ uncertainty (2σ) during the glacial period (Svensson et al., 2008, and Appendix A1). A precise synchronisation between marine and ice cores during HE remains however difficult to establish due to the lack of a direct HE fingerprint within the Greenland ice core records. Neither ice δ^{18}O (a qualitative proxy of temperature) nor Greenland temperature (reconstructed from δ^{15}N measurements in the air bubbles based on firn gas gravitational and thermal fractionation) do exhibit any additional cooling during Heinrich stadials compared to Greenland stadials (Figs. 1, 4a,b and e.g., Kindler et al., 2014). However, Greenland ice cores do provide proxy records influenced by high, mid and low latitude climate changes. This archive should thus permit to explore time leads and lags in between events happening at different latitudes.

The combination of all stable water isotopes has been proposed as a useful tool to disentangle local from more remote effects recorded in ice cores (Vimeux et al., 1999; Stenni et al., 2001; Masson-Delmotte et al., 2005; Jouzel et al., 2007; Landais et al., 2012b). δ^{18}O is a qualitative proxy of local Greenland temperature showing the well-known GS-GI pattern. Deuterium excess ($d - \text{excess} = \delta D - 8\delta^{18}$O, Dansgaard, 1964) bears a signature of vapour source characteristics (sea surface temperature and relative humidity), but it is also affected by changes in condensation temperature (Masson-Delmotte et al., 2005; Jouzel et al., 2007; Steen-Larsen et al., 2013; Bonne et al., 2014). Temperature influences similarly the fractionation coefficients associated with δ^{17}O and δ^{18}O while the variations of fractionation coefficients associated with δD follow a different link with temperature (Barkan and Luz, 2005; Majoube, 1971; Luz et al., 2009). As a consequence, ^{17}O-excess defined as $\ln(\delta^{17}O+1) - 0.528\ln(\delta^{18}\text{O}+1)$ is less sensitive than d-excess to the distillation history, hence to local temperature (Masson-Delmotte et al., 2005; Landais et al., 2012b). ^{17}O-excess therefore better reflects the climatic conditions during evaporation or moisture recharge of Greenland vapour source, in particular the relative humidity that controls kinetic fractionation during such processes. At NEEM, a simple water isotopic model tuned on seasonal data has enabled one to show that ^{17}O-excess increases by 1 permeg for a 1 ‰ decrease in relative humidity of the source evaporative region (Landais et al., 2012b).

Still, because kinetic fractionation is also important during snow formation and as already observed for d-excess (Jouzel et al., 2007), the influence of condensation temperature on ^{17}O-excess increases when temperature decreases. While this effect is of second order for Greenland sites or Antarctic sites with a δ^{18}O level higher than -40 ‰ (Winkler et al., 2012; Landais et al., 2012b), it is dominant for colder sites in Antarctica (Landais et al., 2012a; Schoenemann et al., 2014). This is the reason why ^{17}O-excess and d-excess alone should be combined with other ice core proxies of the lower latitudes (e.g. $\delta^{18}O_2$ of O_2, CH_4) to faithfully establish the relationship between climate and environmental changes in both polar and lower latitudes.

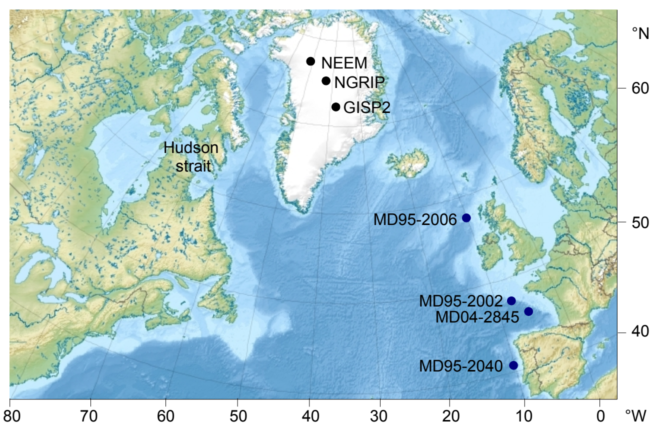

Fig. 2. Map of the Northern Hemisphere palaeoclimate archives used in this study. Blue dots: marine cores. Black dots: ice cores. Most of the IRD for H4 and H5 were deposited by icebergs originating from the Laurentide ice sheet and delivered through Hudson Strait (Hemming, 2004). All marine cores on the map contain Laurentide IRD for H4 and H5. Background map from Uwe Dedering.

In this study, we investigate multiple climate proxies registered in Greenland ice cores (Fig. 2): NEEM (North Greenland Eemian Ice Drilling), NGRIP (North Greenland Ice Core Project) and GISP2 (Greenland Ice Sheet Project 2), on the same chronology (Greenland Ice Core Chronology 2005 or GICC05, Appendix A), giving information about the local as well as the remote climate on exactly the same timescale. The aim is to test the synchronicity of climate events at high and low latitudes during Greenland stadials. With this purpose, we present the first ^{17}O-excess record over a sequence of Dansgaard-Oeschger events, with a specific focus over GS-9 (following GI-9 and preceding GI-8 according to the INTIMATE labelling scheme that we use in this study (Björck et al., 1998), 38 220–39 900 years before AD 2000 (a b2k) on the GICC05 timescale). GS-9 is characterised by the occurrence of the major H4 IRD event of mostly Laurentide ice sheet (LIS) origin (Hemming, 2004; Jullien et al., 2006). This event occurs during Marine Isotope Stage 3 (MIS3), a period consisting of short-lived and frequent DO events (NGRIP members, 2004). In the following we will argue that the climatic fingerprint of H4 can be identified in multiple proxy records sensitive to climate and environmental changes at high, mid and low latitudes, archived in the ice and air of Greenland ice cores.

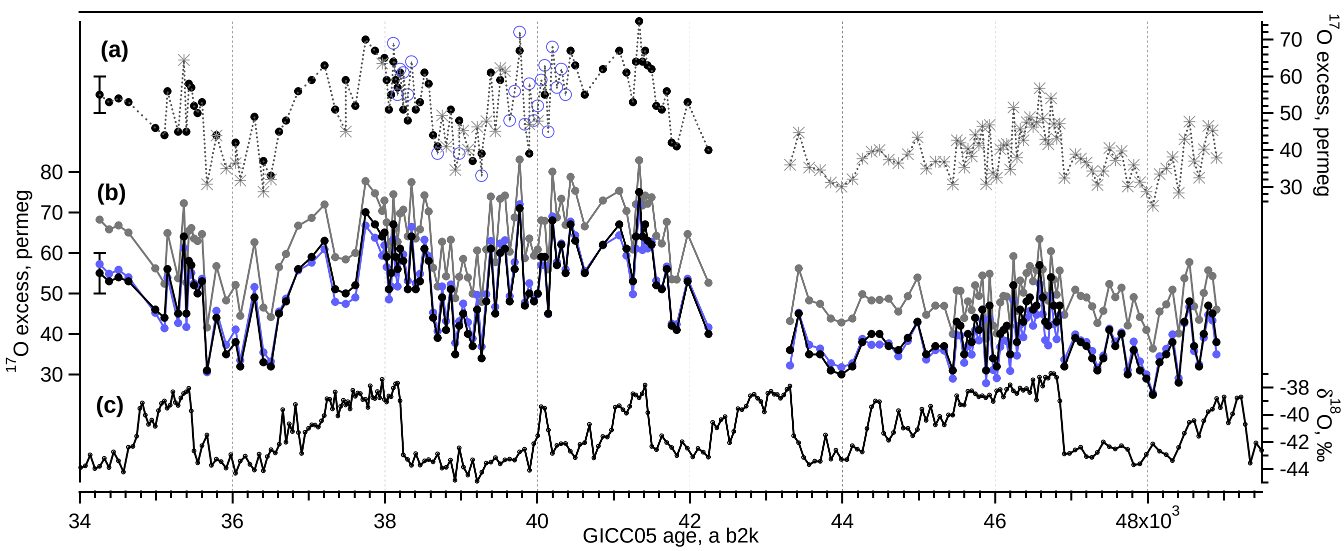

Fig. 3. NEEM ^{17}O excess (permeg) and δ^{18}O (‰) data. **(a)** Comparison of the ^{17}O-excess data measured in spring 2011 (black dots), at the beginning of 2012 (blue circles) and in winter 2013 – spring 2014 (grey stars). **(b)** Calibration of the ^{17}O-excess data. Grey dots: measured data; blue dots: one point calibration; black dots: two point calibration. **(c)** δ^{18}O, ‰.

2 Measurement method

2.1 ^{17}O-excess

We have performed the first ^{17}O-excess measurements on the NEEM ice core (Dahl-Jensen and NEEM community members, 2013, localisation on Fig. 2), spanning a sequence of 7 DO events during MIS 3, from GI-7 to GI-13. The method for ^{17}O-excess measurements was first described in Luz and Barkan (2005). 2 microlitres of water are fluorinated under a helium flow and the resulting oxygen O_2 is purified on a molecular sieve before being trapped in a manifold immersed in liquid helium. O_2 is then measured on a DELTA V isotope ratio mass spectrometer vs.

pure oxygen. Dual inlet measurements last 75 min for each converted oxygen sample. Each water sample has been converted and measured at least twice. Every day, we use 2 in-house water standards in our fluorination line to check the stability of the measurements and to perform calibrations. The resulting pooled standard deviation for ^{17}O-excess measurements is 5 permeg.

For this study with δ^{18}O varying between -43 and -38 ‰, we mainly used 3 in-house standards at around -32, -40 and -58 ‰. These in-house standards are calibrated vs. V-SMOW (Vienna Standard Mean Ocean Water, δ^{18}O=0 ‰; ^{17}O excess=0 permeg) and SLAP (Standard Light Antarctic Precipitation, δ^{18}O=-55.5 ‰; ^{17}O excess=0 permeg, Luz et al., 2009; Schoenemann et al., 2013). The calibration of our raw ^{17}O-excess data can then be performed in two different ways. In a first attempt, we have simply subtracted from all raw ^{17}O-excess data the ^{17}O-excess difference between the measured and the calibrated standard at -40 ‰. In a second attempt, we have used a two point calibration between measured and accepted values of V-SMOW and SLAP so that the ^{17}O-excess correction increases with the δ^{18}O difference between O_2 obtained from the sample conversion and O_2 obtained from V-SMOW conversion. We show in Fig. 3b the comparison of ^{17}O-excess evolution after these two corrections as well as the raw data. The general evolution of the ^{17}O-excess profile and the particular separation in phases over GS-9 is not affected by the different corrections.

Finally, note that the measurements were done in three rounds, with a first series measured in spring 2011, a second one at the beginning of 2012 and a third one in spring 2014. Comparisons of the profiles are displayed in Fig. 3a after correction with the two point calibration (V-SMOW vs. SLAP). The measurements were not necessarily performed over the same exact depth levels, which makes the inter-comparison less accurate. Both the mean levels and the variability are very coherent between the three ^{17}O-excess profiles.

2.2 $\delta^{18}O_{atm}$

Our new $\delta^{18}O_{atm}$ (δ^{18}O of O_2) dataset measured on the NEEM ice core consists of 95 data points with replicates. A melt-refreeze technique has been used to extract the air from the ice samples. Our isotope ratio mass spectrometer is equipped with 10 Faraday cups to measure simultaneously the isotopic ratios δ^{18}O and δ^{15}N as well as the elemental ratios O_2/N_2 (Landais et al., 2010).

To reconstruct the past $\delta^{18}O_{atm}$ signal, δ^{18}O measurements have been first corrected for eventual gas loss during ice storage using the $\delta O_2/N_2$ ratio measured in the same samples (more details can be found in Landais et al., 2010):

$$\delta^{18}O_{\text{gas loss corrected}}=\delta^{18}O_{\text{measured}}+0.01\,(\delta O_2/N_2+10) \quad (1)$$

The obtained data are then corrected for thermal and gravitational fractionation occurring in the firn. To do so, we use the isotopic composition of nitrogen δ^{15}N measured in the very same ice samples as well as firn modelling, to reconstruct the gravitational ($\delta^{15}N_{grav}$) and thermal ($\delta^{15}N_{therm}$) signals contributing to the measured δ^{15}N data (Guillevic et al., 2013; Landais et al., 2010, and references therein). The gravitational effect being proportional to the mass difference in between the two isotopes (Craig et al., 1988; Schwander, 1989), $\delta^{18}O_{grav}$ is therefore twice as large as $\delta^{15}N_{grav}$. Temperature gradients in the firn create thermal fractionation of nitrogen and oxygen isotopes, and this effect is 1.6 times larger for δ^{18}O compared to δ^{15}N due to differences in diffusivity coefficients (Severinghaus et al., 2001). The final $\delta^{18}O_{atm}$ is therefore obtained as follow:

$$\delta^{18}O_{atm}=\delta^{18}O_{\text{gas loss corrected}}-1.6\,\delta^{15}N_{therm}-2\,\delta^{15}N_{grav} \quad (2)$$

The resulting $\delta^{18}\mathrm{O}_{\mathrm{atm}}$ pooled standard deviation is 0.03 ‰ (Landais et al., 2010). The $\delta^{18}\mathrm{O}_{\mathrm{atm}}$ dataset is reported on the GICC05 timescale (Fig. 4f, black dots) using the NEEM gas age scale from Guillevic et al. (2013). The obtained profile is in agreement with previously published results from Severinghaus et al. (2009) measured on the Siple Dome ice core, Antarctica (Fig. 4f, grey dots and Appendix A2, Fig. 7). The advantage of using the NEEM rather than the Siple Dome $\delta^{18}\mathrm{O}_{\mathrm{atm}}$ data resides in the smaller NEEM ice-gas synchronisation uncertainty during GS-9 (Appendix A2).

3 Results and discussion

3.1 ^{17}O-excess record

The entire ^{17}O-excess record covering GI-7 to GI-13 (Fig. 4d) shows a general correlation with the δ^{18}O record and an anticorrelation with d-excess, while δ^{18}O and d-excess are quite well anticorrelated: GI corresponds to high δ^{18}O and ^{17}O-excess levels and low d-excess levels, GS corresponds to low δ^{18}O and ^{17}O-excess levels and high d-excess levels. Following the current understanding of the ^{17}O-excess proxy as given in introduction, such a general correlation with δ^{18}O can be explained by (i) a synchronous change of high latitude temperature and lower latitude hydrological cycle (relative humidity at the oceanic source of evaporation or change in the water mass trajectory / recharge) and/or (ii) an influence of local temperature on ^{17}O-excess through kinetic effect at snow formation. In addition to this general correlation with δ^{18}O, the high resolution ^{17}O-excess record reveals periods where ^{17}O-excess is decoupled from δ^{18}O. The most obvious decoupling between δ^{18}O and ^{17}O-excess is observed during GS-9: ^{17}O-excess decreases after the δ^{18}O decrease at the beginning of GS-9 and increases several centuries before the δ^{18}O increase at the beginning of GI-8. In fact, within GS-9 of constantly low δ^{18}O, we can infer 3 distinct phases from ^{17}O-excess variations (Fig. 4d, Table 2): phase 1 (39 900–39 350 a b2k), corresponding to the beginning of GS-9, is marked by high ^{17}O-excess values (mean = 55 permeg); phase 2 (39 350–38 610 a b2k) corresponds to low ^{17}O-excess values (down to 34 permeg, mean = 42 permeg); finally during phase 3 (38 610–38 220 a b2k) ^{17}O-excess reaches high interstadial levels again (mean = 57 permeg).

The decoupling observed between δ^{18}O and ^{17}O-excess implies that variations of ^{17}O-excess during GS-9 cannot be due to local (Greenland) temperature changes. This conclusion is strengthened by an independent temperature reconstruction based on firn gas fractionation (δ^{15}N, Fig. 4b and Guillevic et al., 2013) confirming a cold and stable temperature over the entire GS-9. The ^{17}O-excess variations during GS-9 thus necessarily reflect changes in the organisation of the hydrological cycle of the lower latitudes, either a change in the relative humidity of the oceanic source of evaporation or a change in the recharge of water before reaching Greenland. It is counter-intuitive that such changes in the evaporation conditions or moisture source do not affect the d-excess signal. While we do not have a definite explanation for the absence of signal in the d-excess record, we already know that d-excess in Greenland is much more sensitive to local temperature changes than ^{17}O-excess (Winkler et al., 2012; Landais et al., 2012b). Moreover, because of its linear definition, d-excess is not sensitive to mixing of different water masses along the trajectory. A simple mathematical calculation (Risi et al., 2010) shows that the mixing of two water vapour masses with similar d-excess and ^{17}O-excess but different δ^{18}O produces a stable d-excess but a decrease in ^{17}O-excess.

Because of the different possible influences on ^{17}O-excess, it is thus important to confront our finding to other low latitudes proxies registered in Greenland ice cores, to independently confirm the decoupling between Greenland and lower latitudes during GS-9 as highlighted by the ^{17}O-excess record.

3.2 Multi-proxy identification of a 3-phase sequence during GS-9

Our new high resolution $\delta^{18}O_{atm}$ measurements of air entrapped in the NEEM ice core confirm a decoupling between Greenland temperature and lower latitude water cycle over GS-9. Indeed, earlier studies have shown that millennial $\delta^{18}O_{atm}$ variations are global responses to water cycle changes in the northern tropics (Severinghaus et al., 2009; Landais et al., 2010), where ITCZ shifts drive changes in precipitation isotopic composition (Lewis et al., 2010; Pausata et al., 2011). Our 95 new measurements of $\delta^{18}O_{atm}$ depict a stable, low $\delta^{18}O_{atm}$ during phase 1, followed by an increase over phase 2 (+0.14 ‰), and finally a stable high plateau during phase 3, in agreement with the Siple Dome record (Fig. 4f). The onsets of changes of $\delta^{18}O_{atm}$ and ^{17}O-excess occur synchronously at the beginning of phase 2 (ice-gas synchronisation uncertainty less than ±100 a for the NEEM record, Appendix A2), but the increase of $\delta^{18}O_{atm}$ is much slower than the observed changes in ^{17}O-excess; this is expected because of the long residence time of oxygen in the atmosphere (1000–2000 a, Bender et al., 1994a; Hoffmann et al., 2004). Second, in addition to the $\delta^{18}O_{atm}$ record, records of NEEM CH_4 (Fig. 4g Chappellaz et al., 2013) and its hydrogen isotopic composition (Fig. 4e, NGRIP $\delta D-CH_4$, Bock et al., 2010) in the air trapped in ice core also exhibit significant variations during GS-9 that are synchronous with the 3 phases identified from the ^{17}O-excess variations. Phase 1 is characterised by a minimum in methane mixing ratio, whereas the onset of phase 2 is marked by a 20 ppb abrupt increase accompanied by a 8 ± 4.8 ‰ increase in $\delta D-CH_4$. At the onset of phase 2, the concomitant changes in CH_4, $\delta D-CH_4$, $\delta^{18}O_{atm}$ and ^{17}O-excess in the absence of any Greenland temperature change would again well be explained by a low latitude change in water cycle as induced by a southward shift of the ITCZ.

Concerning the measured increase in $\delta D-CH_4$ at the onset of phase 2, such a shift would also be in agreement with model simulations of the clathrate release signature in ice cores (Bock et al., 2012), and is also supported by the synchronously lighter $\delta^{13}C-CH_4$ anomaly recently measured in the EDML ice core (Möller et al., 2013). We thus cannot exclude a punctual clathrate release at the onset of phase 2 that might be caused by e.g. less dense water masses in the North Atlantic Ocean following e.g. a massive iceberg discharge, as modelled by Flückiger et al. (2006). Higher resolution methane isotopic data would be necessary to investigate in details this possibility. However, the long duration of the $\delta D-CH_4$ anomaly (phase 2, 740 a) as well as the stable plateau of methane mixing ratio and $\delta^{13}C-CH_4$ throughout phase 2 and 3 (1130 a) call for an additional mechanism, such as source mix change or/and heavier isotopic composition of tropical precipitation (Möller et al., 2013) at the onset of phase 2. The mixing ratio increase and the following plateau are consistent with the activation of new methane sources, in e.g. South America as simulated by Hopcroft et al. (2011), associated with a southward ITCZ shift at the onset of phase 2. Moreover, such a southward ITCZ shift is expected to produce more depleted precipitation in the SH (Southern Hemisphere) tropics and heavier precipitation in the NH (Northern Hemisphere) tropics (Pausata et al., 2011; Lewis et al., 2010). The increase in both $\delta D-CH_4$ and $\delta^{18}O_{atm}$ at the onset of phase 2 are consistent with this mechanism, provided that NH tropics remain the main source of methane and oxygen. A minor part of the $\delta D-CH_4$ increase can be due to oxidation of methane in the troposphere, consuming preferentially the light methane isotopologues, if the mean atmospheric temperature decreases at the onset of phase 2 (Lewis et al., 2010; Bock et al., 2010; Sowers, 2006; Levine et al., 2012).

Finally, two abrupt ~ 20 ppm increases in the atmospheric CO_2 concentration recently unveiled from high reso-

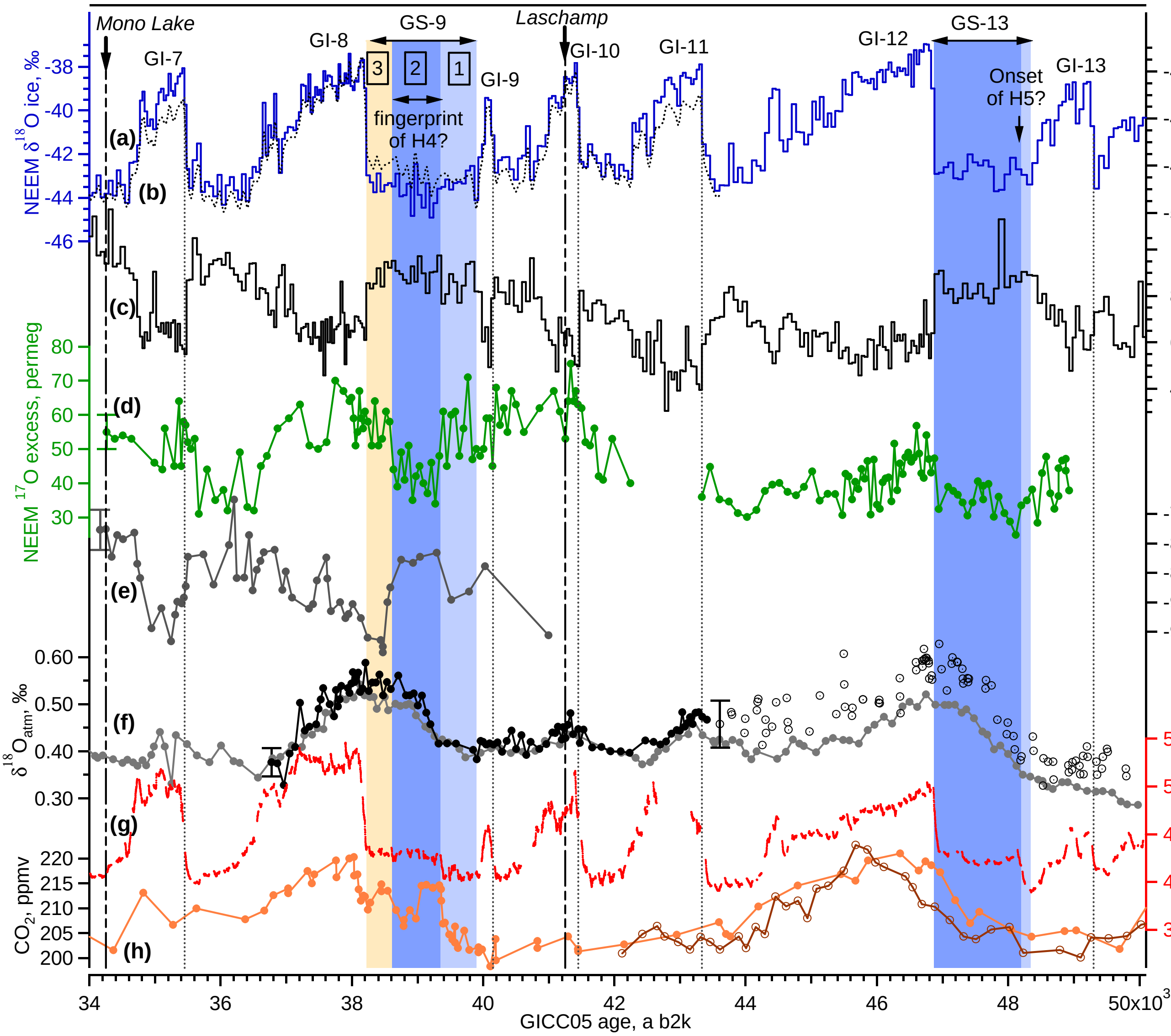

Fig. 4. Greenland and Antarctic ice core records surrounding Heinrich event 4 and 5, synchronised to the GICC05 timescale (Appendix A). Position of phase 1 (light blue area), phase 2 (blue area) and phase 3 (yellow area) as given in Table 2. Dashed-dotted vertical lines: position of the geomagnetic excursion events Mono Lake and Laschamp (Svensson et al., 2006, 2008). **(a)** Blue line: NEEM δ^{18}O ice, ‰ (precision: 0.07 ‰) (Guillevic et al., 2013). **(b)** Black dotted line: NEEM temperature reconstructed using δ^{15}N data and firn modeling (Guillevic et al., 2013). **(c)** Black: NEEM deuterium excess, ‰ (±0.7 ‰), this study, measured at LSCE (France). **(d)** Green dots: NEEM ^{17}O-excess, permeg (±5 permeg), this study, measured at LSCE. Each dot corresponds to the average over 55 cm of ice. **(e)** Dark grey: NGRIP δD-CH_4, ‰ (±3.4‰) (Bock et al., 2010). **(f)** Black dots: NEEM $\delta^{18}O_{atm}$, ‰ (±0.03 ‰), this study, measured at LSCE. Black circles: NGRIP $\delta^{18}O_{atm}$ (Huber et al., 2006). Grey dots: Siple Dome $\delta^{18}O_{atm}$ (Severinghaus et al., 2009). **(g)** Red: NEEM methane mixing ratio, ppbv (±5 ppbv), record measured by the CIC instrument (Chappellaz et al., 2013). **(h)** CO_2 mixing ratio, ppmv. Orange dots: Byrd ice core, Antarctica (Ahn et al., 2012). Brown open circles: TALDICE ice core, Talos Dome, Antarctica (Bereiter et al., 2012).

lution Antarctic ice core records occur at the onsets of phase 2 and phase 3 (Fig. 4h and Ahn et al., 2012). After Antarctic and Greenland gas record synchronisation through CH_4 (Appendix A2), our study evidences that the CO_2 rise at the onset of phase 2 is synchronous ($\pm$100 a) with a southward shift of the ITCZ, as suggested by the other NEEM ice core proxy variations. Different mechanisms are proposed to explain this CO_2 rise, as already discussed in Ahn et al. (2012): most probably upwelling in the Southern Ocean around Antarctica as suggested by increased opal burial rates in marine cores (Anderson et al., 2009), with a potential minor contribution from upwelling off the NW African coast (Jullien et al., 2007; Itambi et al., 2009; Mulitza et al., 2008). The causes of such modifications of the oceanic carbon storage may involve strengthening and/or southward shift of the SH westerlies, and/or AMOC slow down, and therefore remain debated (e.g., Toggweiler et al., 2006; Völker and Köhler, 2013; Menviel et al., 2014).

Altogether, our multi-proxy ice core dataset reveals a 740 a long period starting 550 a after the onset of GS-9 (phase 2), which is marked by southward shifts in North Atlantic storm tracks and ITCZ. The transition to phase 3 is characterised by reversed variations, with ^{17}O-excess and δD$-$$CH_4$ reaching interstadial values 480$\pm$50 a prior to the increase in δ^{18}O marking the onset of GI-8. We explain this pattern by a gradual northward shift of the ITCZ associated with lighter precipitation in the NH tropics and heavier precipitation in the SH tropics. The flat methane and $\delta^{18}O_{atm}$ records suggest a globally stable methane budget and biosphere productivity, possibly due to compensating effects at different latitudes.

Some fingerprints of the decoupling between high and low latitudes identified during GS-9 can also be evidenced during GS-13 even though less proxy data are currently available. We observe synchronous $\delta^{18}O_{atm}$ and methane increases about 140$\pm$40 a after the beginning of GS-13, at 48 200 a b2k on the GICC05 chronology, 1340 $\pm$80 a before the onset of GI-12. This shift is accompanied by a small ^{17}O-excess decrease. The CO_2 increase that was clear at the onset of phase 2 during GS-9 is however much less pronounced or perhaps even absent at the onset of phase 2 during GS-13 (+5 ppm in TALDICE, less clear in the Byrd CO_2 record, Bereiter et al., 2012; Ahn et al., 2012, and Fig. 4h). Finally, we do not observe any decoupling between Greenland and lower latitudes toward the end of GS-13 as was observed over the phase 2/phase 3 transition during GS-9.

3.3 Comparison with marine core records from the North Atlantic

Low to mid latitude palaeoclimate archives have already provided information on climate variations during Greenland stadials and especially over Stadial 9 (e.g., Voelker et al., 2006). In particular, several phases have already been identified in sediment cores off the European margin and especially off Iberia, based on pollen assemblages and charcoal records (Naughton et al., 2009; Daniau et al., 2009): in the first phase, before the IRD layer and until the maximum IRD content is reached, the climate follows a progressively cooler and wetter trend. In the second phase after the IRD peak response, the climate is on the contrary progressively drier, with more frequent fire regime episodes.

Hereafter, we combine North Atlantic marine cores from different latitudes along the European margin, latitudinally situated close to each other between 5 and 10° W (Figs. 2 and 5).

As for the multi-proxy study of GS-9 in ice cores, the multi-proxy study of several marine cores in the European margin shows some decoupling during stadial 9: (i) prior to the Laurentide IRD delivery, cold conditions are already depicted in the eastern part of the so-called Ruddiman belt (Ruddiman, 1977), at 45–55° N (e.g. Fig. 5a, core MD95–2006, Peters et al., 2008; Dickson et al., 2008), while interstadial conditions persist around the Iberian

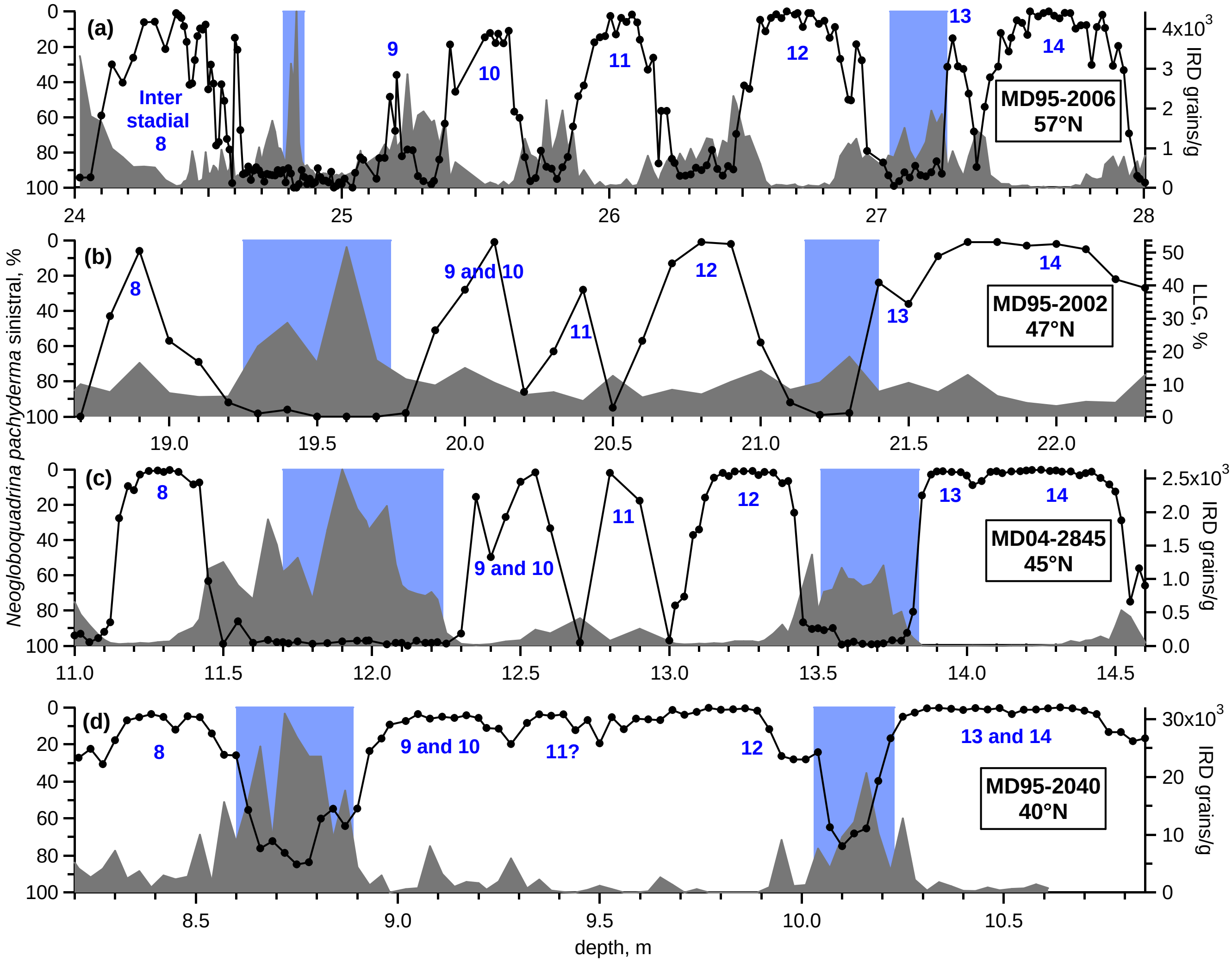

Fig. 5. H4 and H5 in marine core records from the European margin. Core localisations are shown in Fig. 2. Note that each core is on its own depth scale. **(a)** Core MD95-2006, Barra Fan, 57° 01′82 N, 10° 03′48 W (Peters et al., 2008; Dickson et al., 2008). **(b)** Core MD95-2002, Celtic margin, 47° 27′12 N, 8° 27′03 W (Auffret et al., 2002). **(c)** Core MD04-2845, Bay of Biscay, 45° 21' N, 5° 13' W (Sánchez Goñi et al., 2008). **(d)** Core MD95-2040, Portugese margin, 40° 34′91 N, 9° 51′67 W (de Abreu et al., 2003; Eynaud et al., 2009). Grey shaded area: IRD or LLG (large lithic grains), counted on the $> 150\,\mu$m fraction. Black line and dots: abundance of the polar foraminifer *N. pachyderma* sinistral, %; a high percentage denotes cold sea surface temperature. Blue shaded areas: IRD/LGG from the Laurentide ice sheet. Origin identification of IRD for core MD95-2006 based on mineral magnetic measurements (for H4 and H5, Peters et al., 2008) and an additional unpublished count of detritial carbonate grains (for H4, W. Austin, personal communication, 2014), on ϵ neodym and magnetic susceptibility for core MD95-2002 (Auffret et al., 2002) and on magnetic susceptibility (as proposed by Thouveny et al., 2000, for the Iberian Margin) for core MD04-2845 (Sánchez Goñi et al., 2008) and MD95-2040 (de Abreu et al., 2003).

Peninsula at 35–42° N (e.g. Fig. 5d, core MD95-2040, de Abreu et al., 2003); (ii) the onset of the Laurentide IRD delivery coincides with a possible further cooling at high latitudes, while Iberia enters into cold stadial conditions; (iii) once low Laurentide IRD levels are registered again, mild to interstadial conditions are recorded around the Iberian Peninsula, while cold temperatures persist at 45–55° N. In-between 40 and 57° N, cores MD04-2845 and MD95-2002 suggest that the situation might be intermediate, with a slight lag of the Laurentide IRD layer behind the polar species *Neogloboquadrina pachyderma* sinistal increase (Fig. 5b, c, Auffret et al., 2002; Sánchez Goñi et al., 2008). Higher resolution data as well as climate-independent synchronisation tools in between marine cores (such as measurements to detect specific palaeomagnetic events and/or volcanic ash layers) are necessary for a better description.

Even if the chronological uncertainties prevent us to link ice core to other archives at a centennial year resolution over Stadial 9 (e.g., see discussion in Blaauw et al., 2009), we note that a same southward ITCZ shift can explain both the fingerprints in $\mathrm{CH_4}$, $\delta\mathrm{D{-}CH_4}$, ^{17}O-excess and $\delta^{18}\mathrm{O_{atm}}$, as well as the colder conditions registered in the Iberian Peninsula that correspond to the purge of the LIS, i.e. the beginning of H4. Indeed, the H4 iceberg discharge could lead to European cooling as well as southward ITCZ shift, as simulated by climate models (Chiang et al., 2008). With this assumption, the inferred start of H4 would be $39\,350 \pm 1520$ a b2k on the GICC05 chronology.

Finally, during phase 3, we note a progressive shift towards interstadial conditions first indicated by lower latitude ice core proxies (Fig. 4, $\delta\mathrm{D{-}CH_4}$, ^{17}O-excess). This would be consistent with a progressive northwards progression of the climate recovery towards interstadial climate conditions. This is in agreement with marine cores from low to mid latitudes suggesting the start of a progressive AMOC recovery before full interstadial climate conditions are reached (e.g., Vautravers et al., 2004). The combination of ice and marine core data therefore supports the following sequence of events in phase 3: end of the iceberg delivery in the North Atlantic, northwards shift of the ITCZ and progressive AMOC restart, and finally the abrupt Greenland warming.

3.4 How realistic is a decoupling between Greenland temperature and lower latitudes changes in climate and hydrological cycle?

Our study shows that a change in the mid to low latitudes of the North Atlantic ocean happens after the beginning of GS-9, with a time lag of 550 ± 60 a, i.e. when the North Atlantic region has already entered a cooling trend. Some features of our study suggest that the same may be true for GS-13 with a decoupling between Greenland temperature and lower latitudes climate 140±40 a after the beginning of GS-13. We have hypothesised that this change may be linked to iceberg deliveries, i.e. H4 and H5. This assumption is actually supported by recent studies investigating the triggering mechanisms of HE. Indeed, it has been suggested that the Northern Atlantic temperature decrease (beginning of GS in our case) is concomitant with the onset of a progressive AMOC slow down and shoaling accompanied by sea ice extent that isolate the sub-surface ocean from the cold atmosphere and produce a sub-surface warming (Rasmussen and Thomsen, 2004; Jonkers et al., 2010). In addition, this sub-surface warming could lead to a sea-level rise of ~ 0.5 to $1\ \mathrm{m}$ by reorganisation of the North Atlantic dynamic topography and by oceanic thermal expansion (as modelled in Shaffer et al., 2004; Flückiger et al., 2006). Both sub-surface warming and sea-level rise then contribute to destabilising ice shelves from the massive Laurentide ice sheet, causing the massive iceberg delivery of Heinrich events occurring several centuries after the onset of the Greenland stadials (Alvarez-Solas et al., 2010; Alvarez-Solas and Ramstein, 2011; Marcott et al., 2011).

The suggestion that Heinrich events 4 and 5 occur after the beginning of Greenland stadials call for another mechanism to induce climate change towards stadial conditions. Recent studies listed hereafter have proposed a local (Arctic) control on Dansgaard-Oeschger events, without the need of a fresh water flux into the North Atlantic Ocean. (i) The studies of Norwegian Seas marine sediment cores from Rasmussen and Thomsen (2004) and Dokken et al. (2013) proposed that stadials were induced by progressively increasing winter sea-ice cover in the Nordic Seas and the formation of a halocline isolating warm sub-surface waters from the North Atlantic drift from the atmosphere. (ii) These studies also showed IRD from the Fennoscandian ice sheet during the warm phases of DO events. The authors proposed that the corresponding fresh water flux helped to establish a halocline and hereby facilitated sea-ice formation, inducing a cooling towards a stadial climate. A model study from Roche et al. (2010) has moreover shown that the overturning cell in the Nordic Seas is very much sensitive to fresh water perturbations, even more than the Ruddiman belt in the North Atlantic. (iii) Finally, a third mechanism (that could be combined with the two previous ones) has been proposed by Petersen et al. (2013): the onset of GI could be induced by the collapse of an ice shelf east of Greenland extending towards Denmark Strait, due to sub-surface melt. The slow regrowth together with the expansion of regional sea-ice cover would create the progressive cooling towards GS. Moreover, these three mechanisms could explain why the Nordic Seas (and Greenland) remain in a stable stadial state, while low to mid latitudes are gradually shifting towards interstadial conditions (as suggested by the increased ^{17}O excess during phase 3): during stadials, the heat advected along the North Atlantic Drift accumulates in sub-surface, below the sea ice in the Nordic Seas or below the ice shelf east of Greenland, until the low-density warm sub-surface water abruptly reaches the surface, thereby melting the sea-ice and/or ice shelf, which results in a rapid warming of the Nordic Seas (and Greenland).

These different mechanisms calling for a local Arctic control on DO events are supported by models studies. Indeed, atmospheric simulations (Li et al., 2010) suggest that Greenland temperature is sensitive to changes in sea ice cover anomalies in the Nordic Seas because of “a strengthening of the easterly flow over the Nordic Seas” impacting Central Greenland. On the contrary, in these simulations, NW Atlantic sea ice cover anomalies (supposed to increase during Heinrich events originating from the Laurentide) only affect the western flank of the Icelandic low and the associated atmospheric circulation anomaly does not noticeably impact Central Greenland. Supporting these results, the modelled regional pattern of Greenland warming in response to Nordic Seas sea-ice retreat is in agreement with regional quantitative reconstruction of Greenland temperature and accumulation rate during MIS3 (Guillevic et al., 2013).

From our data, we found that central Greenland temperature is not sensitive to lower latitude climate changes during stadials. This is actually not a unique case. Other palaeoclimate archives actually show the same insensitivity, e.g. no specific sea ice change is depicted during the entire stadial 9 as inferred from deep sea core records obtained from the Irminger Sea and south of the Faeroe Islands (Cortijo et al., 2005; Zumaque et al., 2012).

However, bio-indicators in marine cores from the Norwegian Sea robustly record warmer temperature (e.g. based on foraminifers, Rasmussen and Thomsen, 2004; Dokken et al., 2013) and reduced sea ice cover (based on dinocysts, Eynaud et al., 2002) during stadials than during interstadials. To reconcile these conflicting model-data results, Rasmussen and Thomsen (2004) and Dokken et al. (2013) proposed a perennial sea-ice cover in the Norwegian Sea during all stadials of MIS3, separated by a halocline from warm subsurface waters in the North Atlantic Drift, to where foraminifers might have moved their depth habitat. Dinocysts on the contrary may have stayed at the surface, and their observed assemblages could be due to productivity anomalies in zones of polynya. According to this scenario, the lack of specific sea ice anomaly in the Nordic Seas during GS-9 and GS-13 may explain why no cold anomaly is recorded in Greenland during the same periods.

The proposed decoupling between cold, stable Greenland temperature and lower latitude climatic changes during stadial 9 has implications for synchronisation of climate archives from different latitudes. For the last glacial period, chronologies of sediment cores from the North Atlantic and Europe are usually constructed by wiggle matching of DO-like variations to the NGRIP δ^{18}O record on its GICC05 chronology. While it has been underlined by several studies that there were no independent correlation to confirm that this hypothesis was correct (Blaauw et al., 2009), the decoupling between ^{17}O excess and δ^{18}O variations during GS-9 suggests that this would actually not be correct for this specific time period. We therefore would like to emphasis the need of high resolution climate independent synchronisation tools, such as volcanic ash layers, as already widely encouraged by the INTIMATE working group (e.g., Austin and Hibbert, 2012).

4 Conclusions and perspectives

Here, we have presented the first ^{17}O-excess records from a Greenland ice core covering a sequence of Dansgaard-Oeschger events during the last glacial period, and encompassing two major Heinrich events of Laurentide origin. The ^{17}O-excess profile in generaly high during GI and low during GS, but also display additional variability that cannot be identified nor in the d18O neither in the d-excess records, in particular during GS-9. The decoupling observed between d18O and 17Oexcess implies that variations of 17Oexcess during GS-9 cannot be due to local (Greenland) temperature changes.

Our new multi-proxy record from Greenland ice cores has revealed a 3 phase sequence of GS-9. While Greenland temperature remains uniformly cold along GS-9, synchronous changes in ^{17}O-excess, $\delta^{18}O_{atm}$, $\delta D-CH_4$, methane and CO_2 are interpreted as a polar ice core fingerprint of the lower latitude climate and hydrological cycle changes, most likely due to a southward shift of the ITCZ delayed by 550±60 a after the beginning of the cold period in Greenland (GS-9). We hypothesise that this lower latitude change dated at $39\,350 \pm 1\,520$ a b2k on the GICC05 chronology may be associated with the strong Heinrich event 4. This is in agreement with recent data and modelling studies suggesting that the iceberg discharge only occurs several centuries after the cooling of surface ocean in North Atlantic and decrease of AMOC intensity.

The fingerprint of the ITCZ southward shift observed in ice core proxies ends 740 ± 60 a later, when ^{17}O-excess, $\delta^{18}O_{atm}$ and $\delta D-CH_4$ have shifted back to interstadial climate values, 390 ± 50 a before the onset of GI-8. Preliminary investigations on GS-13 encompassing H5, based on the ice core proxies ^{17}O-excess, $\delta^{18}O_{atm}$ and CH_4 suggest the same ITCZ southward shift fingerprint, 140±40 a after the beginning of GS-13. Our findings obviously call for systematic and high resolution investigations of the ice core multi-proxy fingerprints of lower latitude climate changes in general and Heinrich events in particular.

The decoupling between the constantly cold Greenland temperature during GS-9 and the climate variability associated with H4 at lower latitudes challenges the use of Greenland ice core temperatures as a single target for benchmarking climate simulations focused on HEs. Our multi-proxy study opens new paths for parallel investigations of different marine, terrestrial and ice core climate archives.

Appendix A: Synchronisation of the used ice cores to the GICC05 chronology

A1 Ice age scales

The GICC05 (Greenland Ice Core Chronology 2005) has been constructed for the NGRIP ice core from present back to 60 ka b2k based on annual layer counting in ice cores of parameters featuring seasonal scale variations (Vinther et al., 2006; Andersen et al., 2006; Rasmussen et al., 2006; Svensson et al., 2008). The GRIP, GISP2 (Seierstad et al., 2014) and NEEM ice cores (Rasmussen et al., 2013b) have been synchronised to the NGRIP ice core using match points based on peaks of electrical conductivity and dielectrical properties of the ice. The synchronisation uncertainty between NGRIP, NEEM, GRIP and GISP2 is estimated to be $\sim 10\,\mathrm{cm}$ (1 sigma, Rasmussen et al., 2013), resulting in $\sim 10\,\mathrm{a}$ synchronisation uncertainty for GS-9. The GICC05 uncertainty is estimated by the maximum counting error in years (Rasmussen et al., 2006), with each uncertain year counted as $0.5 \pm 0.5\,\mathrm{a}$. This can be considered as a $2\,\sigma$ uncertainty. We give the duration uncertainty of each of the phases of GS-9 and GS-13 as the sum of the uncertain years counted within each phase.

A2 Gas age scales

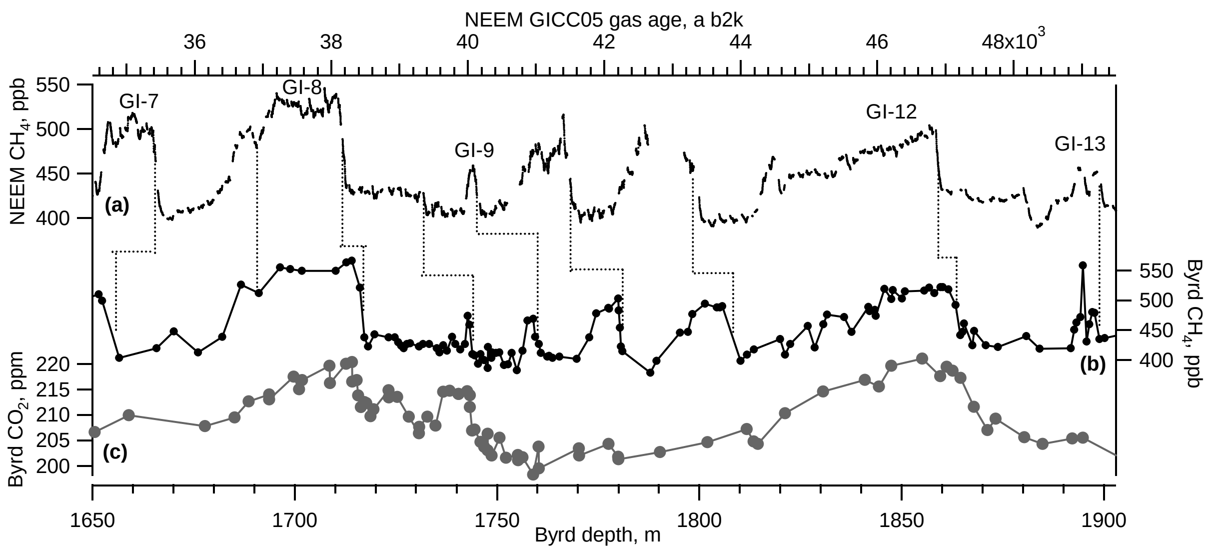

Fig. 6. Construction of a gas age scale for the Byrd ice core. **(a)** NEEM methane data (Chappellaz et al., 2013) on the NEEM GICC05 gas age scale (Guillevic et al., 2013). **(b)** CH_4 and **(c)** CO_2 data from the Byrd ice core on a depth scale (Ahn et al., 2012). The dotted black lines indicate the match points (Table 1) used to tie the Byrd record to the NEEM GICC05 gas age scale.

For the NGRIP gas records ($\delta D{-}CH_4$, $\delta^{18}O_{atm}$), we use the gas age scale from Kindler et al. (2014), initially constructed on the NGRIP ss09sea06bm timescale (NGRIP members, 2004; Wolff et al., 2010), and we transfer it to the GICC05 chronology. We use the NEEM gas age scale from Guillevic et al. (2013) for GS-8 to GS-12, compatible with the later release from Rasmussen et al. (2013b), and the one from Rasmussen et al. (2013b) for GI-12 to GI-14. The NGRIP and NEEM gas age scale covering GS-9 are well constrained (ice-gas synchronisation uncertainty less than 100 years for NEEM, less than 160 years for NGRIP) thanks to numerous measured $\delta^{15}N$ data (Kindler et al., 2014; Guillevic et al., 2013; Rasmussen et al., 2013b).

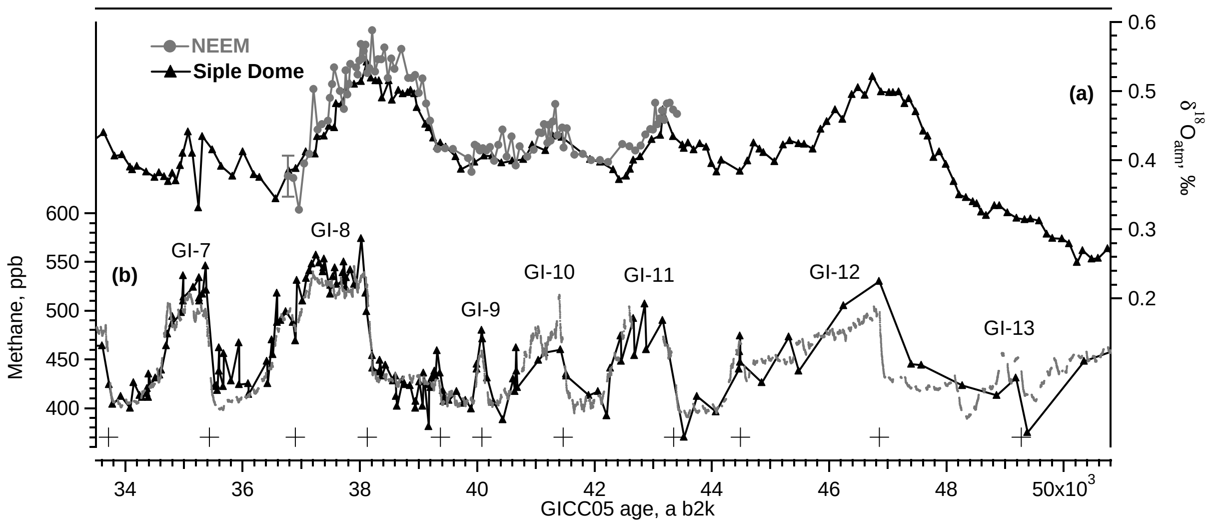

Fig. 7. Comparison of Siple Dome and NEEM $\delta^{18}O_{atm}$ records. **(a)** NEEM (grey dots, this study) and Siple Dome (black triangles, Severinghaus et al., 2009) $\delta^{18}O_{atm}$ data on the GICC05 timescale. **(b)** NEEM (grey line, Chappellaz et al., 2013) and Siple Dome (black triangles, Brook et al., 2005; Ahn et al., 2012) methane data on the GICC05 timescale (Guillevic et al., 2013). The Siple Dome methane record is matched to the NEEM methane record using the match points (Table 1) indicated by the black crosses. Light blue, blue and yellow areas: phases 1, 2 and 3 of GS-9, respectively.

Table 1. Match points between methane depths from the Byrd ice core (Ahn and Brook, 2008; Ahn et al., 2012), the Siple Dome ice core (Brook et al., 2005; Ahn et al., 2012) and NEEM methane (Chappellaz et al., 2013) gas age according to the GICC05 chronology (Guillevic et al., 2013; Rasmussen et al., 2013b).

Event	Byrd depth, m	Siple Dome depth, m	NEEM methane gas age, a b2k
GI-6	1617.15	795.37	33 716
GI-7	1654.05	809.97	35 435
GI-8, dip	1691.10	818.13	36 901
GI-8	1716.10	825.66	38 125
GS-9, plateau	1743.55	833.81	39 372
GI-9	1759.20	837.90	40 121
GI-10	1780.30	845.97	41 467
GI-11	1807.95	855.28	43 351
GI-12	1863.05	870.43	46 859
GI-13	1898.30	882.41	49 276

To synchronise the Byrd gas records to the NEEM ice core, we use a traditional approach by matching methane records from both ice cores (Blunier et al., 1998). We first use the high resolution NEEM methane record on its GICC05 gas age scale (Guillevic et al., 2013). We then match the Byrd methane record (Ahn et al., 2012) to the NEEM methane record, using the mid-slope of each GI onset as match points (Fig. 6 and Table 1). We perform a linear interpolation in between the match points. We then apply the obtain depth-gas age scale for Byrd to the Byrd CO_2 record.

The same method is applied to the Siple Dome methane record in order to place the Siple Dome $\delta^{18}O_{atm}$ record (Severinghaus et al., 2009) on the GICC05 timescale, to compare with our NEEM $\delta^{18}O_{atm}$ record (Fig. 7 and Table 1).

Appendix B: Timing and duration of the phases identified during GS-9 according to the GICC05 timescale

Table 2. Timing and duration of the phases identified during GS-9 in ice core proxies, on the GICC05 timescale. Column 2: onsets of GI and GS are given according to Rasmussen et al. (2013a). Column 3: maximum counting error (MCE), reflecting the number of uncertain annual layers compared to year AD 2000 (Rasmussen et al., 2006, and Appendix A1). Column 5: MCE of the phase duration, calculated as the MCE difference between start and end of each phase. Column 6: total duration uncertainty for each phase: MCE of the phase duration, synchronisation uncertainty of the NEEM ice core to the NGRIP GICC05 timescale (~ 20 a, Rasmussen et al., 2013b) and data resolution (± 35 a in average).

	Start time		Duration		
	a b2k	MCE, a	a	MCE, a	uncertainty (2σ), a
GS-13					
Phase 1	48 340	1988	140	7	40
Phases 2 & 3	48 200	1981	1340	69	80
GI-12	46 860	1912			
GS-9					
Phase 1	39 900	1569	550	51	60
Phase 2	39 350	1518	740	47	60
Phase 3	38 610	1471	390	22	50
GI-8	38 220	1449			

Acknowledgements. We thank S.O. Rasmussen and M. Bock for fruitful discussions. W. Austin kindly provided the data of marine core MD95-2006 and S. Zaragosi the magnetic susceptibility data of core MD04-2845. NEEM is directed and organised by the Centre for Ice and Climate at the Niels Bohr Institute and US NSF, Office of Polar Programs. It is supported by funding agencies and institutions in 14 countries: Belgium (FNRS-CFB and FWO), Canada (GSC), China (CAS), Denmark (FIST), France (IPEV, CNRS/INSU, CEA and ANR), Germany (AWI), Iceland (RannIs), Japan (NIPR), Korea (KOPRI), the Netherlands (NWO/ALW), Sweden (VR), Switzerland (SNF), UK (NERC) and the USA (US NSF, Office of Polar Programs). LSCE analytical work has been funded by the ANR VMC NEEM project. M.G. thanks the University of Copenhagen, Denmark, and the Commissariat à l'Energie

Atomique Saclay, France, for funding. The publication of this article is funded by the “Fondation de France Ars Cuttoli”.

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