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

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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 d-excess and ^{17}O-excess records, tracing changes in the mid-latitudes, with ice δ^{18}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.

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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 CO_2 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),

Clim. Past Discuss., https://doi.org/10.5194/cp-2018-65
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Discussion started: 14 June 2018

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 a diverse range of 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 ($\delta^{18}O$, d-excess, ^{17}O-excess) 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.n 2013; Veres et al., 2013).

The new NGRIP $\delta^{18}O$ and δD dataset was obtained at Laboratoire des Sciences du Climat et de l'Environnement (LSCE) using a laser cavity ring-down spectroscopy (CRDS) analyzer PICARRO. The accuracy for $\delta^{18}O$ 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 $\delta^{18}O$ 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. $\delta^{18}O$ 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 ±0.5‰ or better. Both series show similar $\delta^{18}O$ values, in agreement with the reference $\delta^{18}O$ 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 INSTAR SIL d-excess series by -2.5‰.

In order to perform ^{17}O-excess measurements on water samples at LSCE, water reacts with CoF_3 to produce oxygen whose triple isotopic composition is measured by dual inlet against a reference O_2 gas resulting in a mean uncertainty of 5 ppm (1 σ) for the ^{17}O-excess measurements (Barkan and Luz, 2005). 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 water home standards whose $\delta^{18}O$ encompasses the SMOW – SLAP scale is run every month enabling to calibrate the $\delta^{18}O$ and ^{17}O-excess values (Schoenemann et al., 2013).

Results

Ice core $\delta^{18}O$ (NGRIP community members, 2004) is a qualitative proxy for local surface temperature. Comparisons between ice core $\delta^{18}O$ data and paleotemperature estimates from borehole temperature profile inversion and abrupt temperature changes inferred from isotopic measurements on trapped

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Discussion started: 14 June 2018

air showed that the δ^{18}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). In addition to δ^{18}O records already available (NGRIP community members, 2004), we provide here new d-excess data from NGRIP during the last deglaciation. The second-order parameter d-excess (δD-8xδ^{18}O) (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 $\delta^{18}O_{sea\ water}$ also influence δ^{18}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 $\delta^{18}O_{sea\ water}$ and condensation temperature.

Our dataset also encompasses new ^{17}O-excess data from NGRIP. Defined as ln(δ^{17}O+1)-0.528*ln(δ^{18}O+1)), ^{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 $\delta^{18}O_{sea\ water}$ given that the ^{17}O-excess of global sea water remains constant with time.

Our 1518 new measurements of δ^{18}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) (see methods for details). As previously reported for the central Greenland GRIP ice core (Masson-Delmotte et al., 2005b; Jouzel et al., 2005), the NGRIP δ^{18}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 (30°N to 55°N, Landais et al., 2012) 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 δ^{18}O changes at NGRIP compared to GRIP.

The fact that d-excess increases (by 3.5 ± 1 ‰) when δ^{18}O decreases (by 4 ± 1 ‰) during Greenland stadials relative to interstadials may at least partly reflects 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, δ^{18}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, δ^{18}O variations are smaller than 1 ‰, i.e. less than one fourth of the average amplitude in δ^{18}O changes across DO events, suggesting no large temperature change in Greenland during this period. The link between flat δ^{18}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 $\delta^{18}O$ record (Boyle et al., 1994; Krinner et al., 1997). However, our assumption of stable temperature is supported by constant $\delta^{15}N$ of N_2 values in the GISP2 and NGRIP ice cores (Buizert et al., 2014), $\delta^{15}N$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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). These ^{17}O-excess and d-excess transitions at 16.2 ka do not have any clear counterpart in $\delta^{18}O$ (cf section correlation in SOM) and no temperature variation at that time was recorded in the $\delta^{15}N$ 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 $\delta^{18}O$ 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.

Discussion

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), presented on the same timescale (Figure 2). At 16.2 ka, the CH_4 record indeed exhibits a 30 ppbv peak understood 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., 2915). 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 $\delta^{18}O$ records at 16.2 ka. As observed during all Greenland Stadials of the last glacial period, Antarctic $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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. 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.

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 and Brazil (Partin et al., 2007; 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 d-excess and ^{17}O-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 d-excess and ^{17}O-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, d-excess in the atmospheric vapor is affected by distillation toward higher latitudes, and strongly depends on the source-site temperature gradient, while ^{17}O-excess preserves better the initial fingerprint of relative humidity of the evaporative region.

As a result, the opposing trends observed in d-excess and ^{17}O-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 d-excess and ^{17}O-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 d-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-

Clim. Past Discuss., https://doi.org/10.5194/cp-2018-65
Manuscript under review for journal Clim. Past
Discussion started: 14 June 2018

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 $\delta^{18}O$ 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. This latter interpretation is in line with earlier interpretations of Greenland d-excess changes (Steffensen et al., 2018; Masson-Delmotte et al., 2005b). The Greenland signals may also be 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 [38]. 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).

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). By contrast, a Pa/Th record from the Iberian margin (Gerhardi et al., 2005) at shallower depth (1500 m shallower than the western Atlantic record) shows a significant increase at 16.2 ka. These records may be interpreted as follows. A first modification of the glacial oceanic ventilation occurs at deep depth as early as 18 ka. At 16.2 ka, AMOC may be further destabilized to additionally affect Pa/Th at shallower depths.

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). 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 d-excess data, and a southward shift of tropical rainbelts (Chiang and Bitz, 2005), affecting southern hemisphere CH_4 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). 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, their coherent chronology, 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: $\delta^{18}O$ (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.

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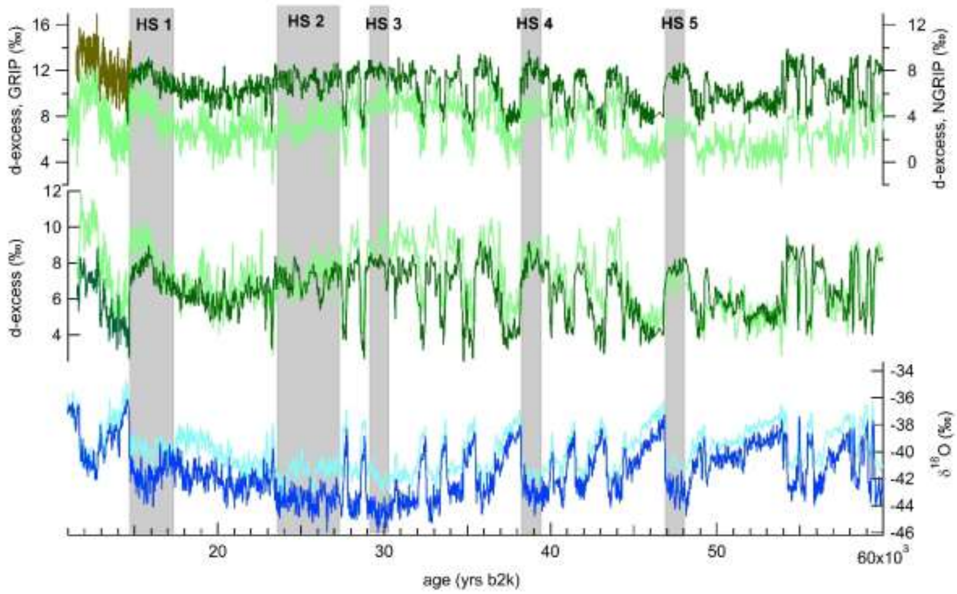


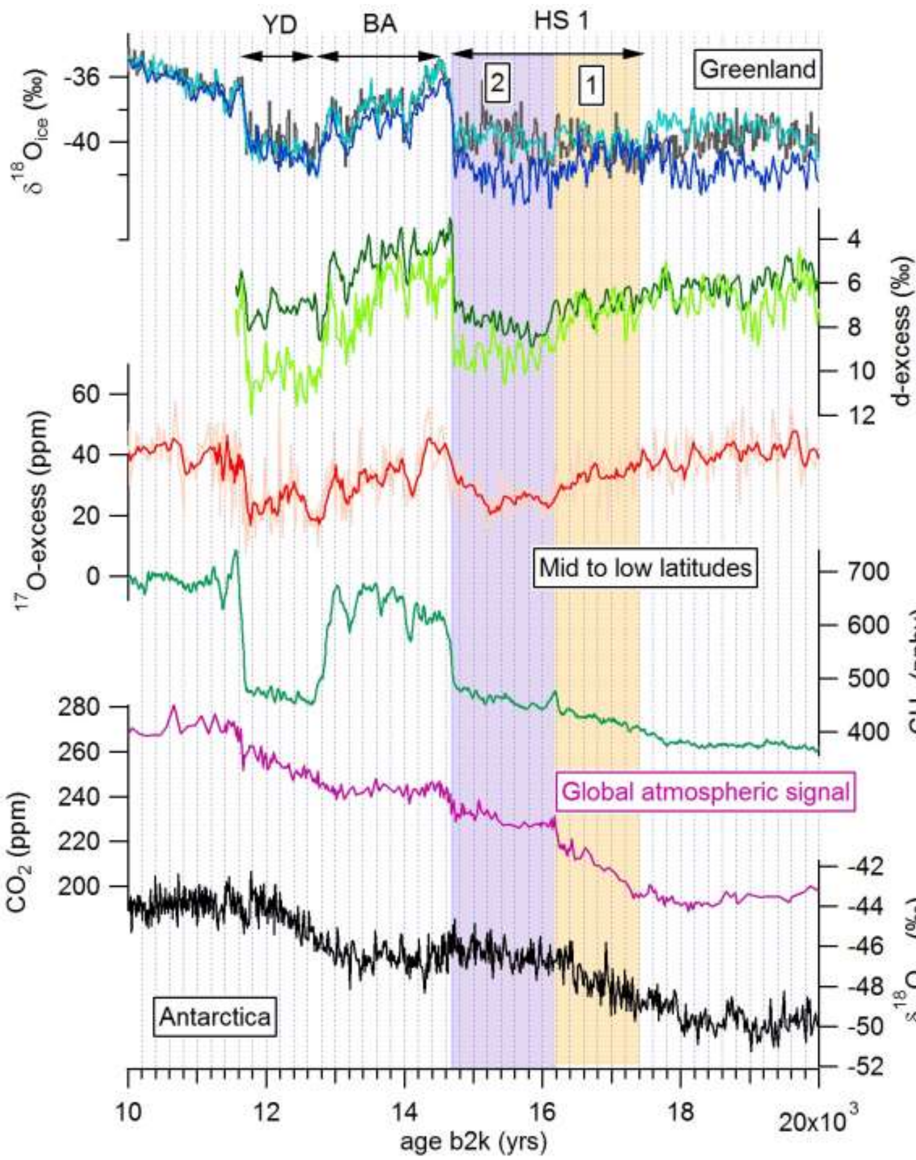
***<u>Figure 1</u>**: water stable isotope records ($\delta^{18}O$ and d-excess, in ‰) from GRIP and NGRIP ice cores reported on the GICC05 chronology (in thousands of years before year 2000 CE). From top to bottom:*

- *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)*
- *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).*
- *$\delta^{18}O$ from the NGRIP ice core (dark blue) datasets, $\delta^{18}O$ from the GRIP ice core (light blue).*

Grey intervals display Heinrich Stadials (HS).

Clim. Past Discuss., https://doi.org/10.5194/cp-2018-65
Manuscript under review for journal Clim. Past
Discussion started: 14 June 2018

Climate of the Past
Discussions

559 

560 *Figure 2: A synthesis of ice core records over the last deglaciation on the synchronized GICC05/AICC2012*
561 *timescales with an identification of two phases (1, orange box and 2, purple box) within Heinrich Stadial*
562 *1 (HS1) as discussed in the text: we locate the transition between phases 1 and 2 at the timing of the*
563 *sharp increase in CO_2 and CH_4 concentrations, both being global atmospheric composition signals. The*
564 *Younger Dryas (YD) and Bølling-Allerød (BA) periods are also indicated.*
565 *From top to bottom:*
566 - *GRIP, NGRIP and GISP2 $\delta^{18}O$ (light blue, dark blue and black respectively (Grootes et al., 1993;*
567 *NGRIP community members, 2004) interpolated at a 20 years resolution*
568 - *GRIP and NGRIP d-excess (light and dark green respectively: Jouzel et al., 2005, this study)*
569 *interpolated at a 20 years resolution*

Clim. Past Discuss., https://doi.org/10.5194/cp-2018-65
Manuscript under review for journal Clim. Past
Discussion started: 14 June 2018

Climate of the Past
Discussions

- *NGRIP* ^{17}*O-excess (orange curve shows the original series and the red curve the 5 years running average, this study)*
- *WAIS Divide* CH_4 (Rhodes et al., 2015)
- *WAIS Divide* CO_2 *(Marcott et al., 2013)*
- *EPICA Dronning Maud Land (EDML)* $\delta^{18}O_{ice}$ *(EPICA community members, 2006)*

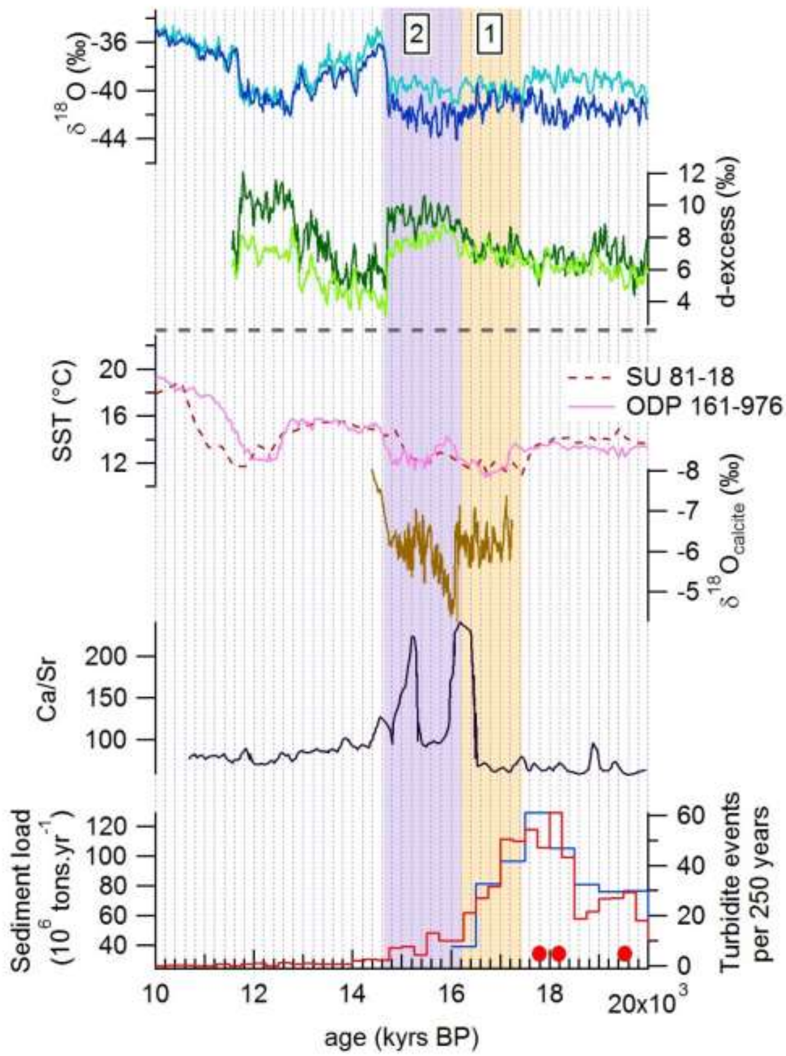


Figure 3: *The sequence of Phase 1 and Phase 2 of Heinrich Stadial 1 identified in Greenland records and in proxy records of North Atlantic SST, IRD events, and changes in East Asian hydroclimate. From top to bottom:*

- *NGRIP (dark blue) and GRIP (light blue) δ^{18}O records*
- *NGRIP (dark green) and GRIP (light green) d-excess records*
- *Sea surface temperature (SST) for North Atlantic cores SU 81-18 (Bard et al., 2000) and ODP 161-976 (Martrat et al., 2014).*
- *Calcite δ^{18}O of Hulu cave (China, Zhang et al., 2014)*
- *Ca/Sr from site U1308 in the IRD belt (Hodell et al., 2019) as signature from strong iceberg discharges from the Laurentide ice sheet.*
- *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*

Clim. Past Discuss., https://doi.org/10.5194/cp-2018-65
Manuscript under review for journal Clim. Past
Discussion started: 14 June 2018

Climate of the Past Discussions

(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.