Deep-water circulation changes lead North Atlantic climate during deglaciation

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Constraining the response time of the climate system to changes in North Atlantic Deep Water (NADW) formation is fundamental to improving climate and Atlantic Meridional Overturning Circulation predictability. Here we report a new synchronization of terrestrial, marine, and ice-core records, which allows the first quantitative determination of the response time of North Atlantic climate to changes in high-latitude NADW formation rate during the last deglaciation. Using a continuous record of deep water ventilation from the Nordic Seas, we identify a ~400-year lead of changes in high-latitude NADW formation ahead of abrupt climate changes recorded in Greenland ice cores at the onset and end of the Younger Dryas stadial, which likely occurred in response to gradual changes in temperature- and wind-driven freshwater transport. We suggest that variations in Nordic Seas deep-water circulation are precursors to abrupt climate changes and that future model studies should address this phasing.
Precise reconstructions that resolve the relative timing of changes in North Atlantic Ocean circulation, climate, and carbon cycling are necessary to anticipate the mechanisms initiating and propagating abrupt global climate changes. During the last deglaciation (~18,000-11,000-years ago), the climate system underwent numerous abrupt changes that have been attributed to variations in the strength of the Atlantic Meridional Overturning Circulation (AMOC)\(^1\). Through changes in high-latitude North Atlantic Deep Water (NADW) formation and export, AMOC exerts an important control on the global climate system by redistributing heat near the surface and regulating carbon storage at depth. In particular, the partitioning of carbon between the surface and deep ocean is thought to play a critical role in centennial-to-millennial-scale variations of atmospheric CO\(_2\) (refs \(^2\)–\(^4\)). However, reconciling the deglacial history of changes in overturning circulation as recorded in marine records with North Atlantic climate and pCO\(_2\) as inferred by Greenlandic and Antarctic ice cores, respectively, remains challenging. First, highly resolved records from deep convection sites sensitive to NADW that monitor the descending branch of AMOC are still lacking. Secondly, large uncertainties in high-latitude marine reservoir ages\(^5\) limit the precision of marine \(^{14}\)C-based chronologies. Thirdly, direct alignment of marine records to far afield Greenland ice-core stratigraphies hinders testing hypotheses of synchronicity. Lastly, precise comparisons between marine and ice-core climate records are hampered by inconsistencies between the radiocarbon and ice-core timescales\(^7\),\(^8\).

Here we present a new synchronization of high-latitude NADW, climate, and pCO\(_2\) records for the last deglaciation based on new marine and ice core data that allows us to conclude for the first time that changes in deep-water circulation in the Nordic Seas led rapid shifts in North Atlantic climate and changes in carbon cycling.

**Results**

**Site location, \(^{14}\)C ventilation and chronology.** We generated a continuous record of deep/intermediate- and surface-water \(^{14}\)C ventilation age from \(^{14}\)C measurements on planktic and benthic foraminifera (Methods) in sediment core MD99-2284 (62° 22.48 N, 0° 58.81 W, 1500 m water depth) from the Norwegian Sea (Fig. 1).

Site MD99-2284, which is characterised by exceptionally high sedimentation rates (>400 cm kyr\(^{-1}\)), is located at the gateway of the Faroe-Shetland Channel (FSC), where warm surface Atlantic water flows into the Nordic Seas and cold dense water overflows into the North Atlantic. Critically, this overflow water is one of two main NADW pathways flowing into the deep North Atlantic and a key constituent of the AMOC\(^9\). During the last glacial period and deglaciation, overflow through the FSC remained a continuous source of NADW\(^1\),\(^11\). Hence, because deep-water \(^{14}\)C activity reflects the circulation-driven exchange of carbon between the atmosphere and deep-ocean reservoir, bottom-surface water \(^{14}\)C age differences of the FSC directly inform past changes in Nordic Seas deep convection, NADW production, and its southward export\(^1\),\(^12\),\(^13\).

The age model for the core was established using a combination of tephrochronology and alignment between sea-surface temperature records from core MD99-2284 and a high-resolution hydroclimate reconstruction from a relatively closely located terrestrial sequence in southern Scandinavia (Methods; Supplementary Figs. 1–5 and Note 1–2). The approach enables us to precisely place our marine proxies on the IntCal13 timescale\(^1\) and to use the foraminiferal radiocarbon data (Methods) to calculate the marine \(^{14}\)C ventilation age. Our estimate was determined using a random walk model (RWM) (Methods and Supplementary Methods) fitted via Markov chain Monte Carlo (MCMC) that took into account uncertainty structures in both calendar age modelling and \(^{14}\)C measurements. To allow detailed comparison with Greenlandic and Antarctic ice-core records, we synchronized the ice-core GICC05 (ref. \(^1\)) and WD2014 (ref. \(^1\)), and \(^{14}\)C timescales using previously published and new \(^{10}\)Be records from GRIP\(^7\) and WAIS Divide ice cores, respectively (Methods; Supplementary Fig. 6). Ages are hereafter reported as IntCal13 years before 1950 AD ± 1σ (BP).

**Deglacial ventilation history of the deep Nordic Seas.** Surface and bottom water mass reconstructions at our site are consistent with existing paleoceanographic records of water properties, transport and exchange between the Norwegian Sea and the northern North Atlantic (Supplementary Figs. 7–8), indicating that our reconstructions are representative of regional oceanographic conditions. Benthic-planktic (B-P) ventilation ages in MD99-2284 decreased by ~700 years seemingly shortly preceding the abrupt warming transition from Greenland Stadial (GS) 2 (equivalent to Heinrich Stadial 1, HS-1, and the Last Glacial Maximum) into Greenland Interstadial (GI) 1 (equivalent to the Bolling-Allerød interstadial, BA; 14,581 ± 16 years BP) (Fig. 2a–d; Supplementary Fig. 9). Although relatively older deep/intermediate water during GS-2 (HS-1) at our site is inferred from only one B-P estimate, this is confirmed by other regional B-P
records (Supplementary Fig. 9). B-P ages gradually decreased during GI-1 (BA) by an additional ∼700 years, reaching or surpassing present-day values (0–50 years) as late as 14,150 years BP (Fig. 2d). This trend suggests that strengthening of deep convection across the onset of GI-1 (BA) was slower than previously thought, which can be attributed to a gradual northward re-initiation of NADW production in the North Atlantic. This interpretation is supported by transient simulations of the last deglaciation, which reveal a northward time-transgressive recovery of NADW formation during GI-1 (BA), with a ∼400-year long re-initiation of deep convection in the Nordic Seas relative to convection sites south of Greenland18. The century-scale northward order of NADW re-initiation is consistent with other model results19. It is also in line with paleoceanographic reconstructions suggesting a delayed resumption of deep convection in the high-latitude Nordic Seas starting midway through GI-1 (BA)20.

Prior to the onset of GS-1 (12870 ± 26 years BP), which corresponds to the Younger Dryas stadial (YD), the B-P age offset rapidly increased by ∼900 years starting at 13,250 years BP. Ahead of the termination of GS-1 (11577 ± 16 years BP), the B-P offset returned to modern values beginning at 11,890 years BP (Figs. 2d–3). GS-1 (YD) onset and termination are also preceded by surface oceanographic changes inferred from sedimentological and lipid biomarker signatures in MD99-2284. Specifically, increasing (decreasing) B-P ages are virtually synchronous with a southward (northward) migration of the polar front across the coring site inferred from faunal assemblages (Fig. 2e; Supplementary Fig. 10), relatively more (less) frequent sea-ice edge conditions inferred from PIP25 (Methods) and iceberg rafting inferred from IRD (Fig. 2f), and relatively higher (lower) inputs of terrestrial meltwater inferred from sedimentary n-alkanes concentrations (Methods; Fig. 2g), ultimately pointing at a tight link between freshwater dynamics and bottom-water ventilation. Although the benthic-atmosphere (B-Atm) offset dominates the B-P signal, the P-Atm offset also decreased prior to the onset of GS-1 (YD) and in phase with the rise in B-P values (Fig. 2c, d). Specifically, P-Atm values slowly declined by ∼500 years starting ∼13,250 years BP and approached the age of the contemporaneous atmosphere by ∼12900 years BP, suggesting that the climatic transition into GS-1 (YD) was preceded by a major slowdown of the Atlantic Inflow...
that resulted in enhanced isotopic equilibration between surface waters and the atmosphere.

**Discussion**

The timing and magnitude of changes in our B-Atm record are supported by other independent reconstructions from the FSC and the Iceland Basin, i.e., regions that monitor ventilation of Nordic Seas overflow waters\(^\text{12}\) (Fig. 3). The data are also in agreement with deep ventilation records from U–Th dated corals in the deep/intermediate northwest and equatorial Atlantic, which monitor the strength of downstream NADW transport\(^\text{22–26}\). Specifically, the slow decrease in B-Atm ages at the end of GS-2 (HS-1) and during the early GI-1 (BA) is observed in data from both the northeast and equatorial Atlantic, whereas early changes prior to the start and end of GS-1 (YD) are evident in each of the records. We hypothesise that these parallel changes in radiocarbon across sites likely reflect downstream propagation of North Atlantic deep/intermediate water \(^{14}\)C signatures associated with high-latitude NADW ventilation rates. In addition, the timing and pattern of the reconstructed NADW changes are consistent with shifts in production–corrected atmospheric \(^{14}\)C (ref. \(^\text{27}\) (Figs. 3, 4), which primarily reflects changes in ocean ventilation and high-latitude NADW formation\(^\text{28,29}\), and thus support an oceanic origin for the variations in atmospheric radiocarbon concentrations during the last deglaciation. Overall, the agreement across the range of independent deep ventilation records supports the view that our B-Atm data capture large-scale fluctuations in NADW circulation. A strong coupling between AMOC and deep ventilation in the Nordic Seas is also corroborated by transient climate model experiments (Supplementary Fig. 11 and Note 3). The simulations, which reproduce a shutdown and subsequent recovery of the AMOC, identify and validate sign and magnitude of the shifts observed in our B-P age reconstruction.

A detailed comparison between our B-P record and ice-core-based temperature records from Greenland using breakpoint analysis (Methods; Supplementary Table 1; Fig. 4), reveals a complex relationship between Nordic Seas ventilation and abrupt climate change. The analysis reveals that changes in Nordic Seas NADW formation occurred before the climate shifts into and out of GS-1 (YD) by 385 ± 32 (1σ bounds) and 447 ± 27 years, respectively, and that weakening of Nordic Seas NADW occurred 437 ± 79 years prior to the first signs of \(pCO_2\) rise near the start of GS-1 (YD). Importantly, the latter finding substantiates the hypothesised\(^\text{30,31}\) lag between AMOC reduction and \(pCO_2\) rise during early deglaciation and is observed in climate simulations as a transient response of the global efficiency of the biological pump to AMOC slowdown (Supplementary Figs. 12–14 and Note 3).

The lag between our high-latitude NADW records and Greenland temperatures across the transitions into and out of GS-1 (YD) requires further consideration. Cooling of sea-surface temperatures and sea-ice expansion in the Norwegian Sea preceding GS-1 (YD) have been documented in terrestrial temperature and isotope records\(^\text{32,33}\), and are consistent with increasing B-P offsets observed in MD99-2284 beginning at 13,250 years BP. Beyond the last deglaciation, records from the last glacial and Late Pleistocene\(^\text{34,35}\) have also implied increased...
input of ice-sheet meltwater, gradual surface cooling, and southward migration of the polar front in the Nordic Seas, preceding transitions from warm interstadials to cold stadial conditions, analogous to the oceanographic changes inferred from MD99-2284 that preceded GS-1 (YD).

On the other hand, ocean warming in the Nordic Seas, a northward diversion of the polar front, and NADW resumption during the second half of the GS-1 (YD) have been widely reported\textsuperscript{36–38}, and further support decreasing B-P offsets in MD99-2284 starting at 11,890 years BP. A lead of a few centuries of changes in high-latitude NADW formation ahead of Greenland temperatures is compatible with the timing of AMOC slowdown and recovery during deglaciation inferred from several Pa/Th circulation\textsuperscript{39} and other AMOC-sensitive proxies\textsuperscript{40,41}. Despite uncertainties with the \(^{14}C\) chronologies and coarse temporal resolution, it has been demonstrated that a century-scale response of Pa/Th to AMOC changes should be additionally accounted for when interpreting this proxy as a function of changes in ocean circulation\textsuperscript{42,43}. The existence of a significant lag time for Greenland temperature and Pa/Th records behind changes in NADW would represent an important observational constraint to account for, for example when attempting to infer the physical mechanisms of rapid climate and AMOC change from climate model simulations\textsuperscript{44,45}. Equally abrupt and synchronous changes in the AMOC, Pa/Th and Greenland temperature (i.e., to within a few centuries) should not, therefore, be expected in numerical model simulations.

On the basis of timing inferred from our \(^{14}C\) ventilation records, we propose that the lagged Greenland temperature response behind changes in Nordic Seas NADW formation can be understood as a threshold response to gradual changes in surface temperature-driven freshwater transport, which ultimately modulates the strength of the northward oceanic heat transport. Both empirical\textsuperscript{34,35,46} and climate modelling studies\textsuperscript{47} have proposed a nonlinear salt oscillator in the North Atlantic system, whereby meltwater production rates are greater (smaller) toward the end of warm (cold) episodes. This is consistent with evidence for southward (northward) migration of the polar front, more (less) extensive sea-ice cover, and increased (decreased) iceberg discharge at our coring site shortly before the termination of GI-1 (GS-1). We further hypothesise that shifts to higher (lower) rates of meltwater fluxes, although initially prompted by gradual climate warming (cooling), are amplified by concomitant changes in surface winds, which play a critical role in the rearrangement of water masses in the Nordic Atlantic. For instance, model studies suggest that sustained cold events like GS-1 (YD)\textsuperscript{48} and the Little Ice Age\textsuperscript{49,50} can be triggered by a positive feedback loop, whereby increasing export of meltwater to the subpolar gyre in response to warming leads to surface ocean cooling and consequent strengthening of atmospheric blocking over the North Atlantic—i.e. large-scale quasi-stationary anticyclonic circulation. This feedback can impede northward heat transport and suppress NADW formation for centuries. Moreover, reduced NADW can lead to an additional freshening through basin-wide subsurface warming\textsuperscript{51,52}, which causes ice-shelf thinning and iceberg discharge\textsuperscript{31}, further bolstering NADW formation weakening. However, we cannot rule out the potential amplifying role of a catastrophic drainage of freshwater from the Baltic Ice Lake into the Nordic Seas\textsuperscript{53}, which occurred precisely at the start of GS-1 (YD), and could have further impacted regional NADW formation.
Atmospheric blocking could be equally important as a mechanism for NADW resumption in the Nordic Seas at the transition out of GS-1 (YD). New studies suggest that under full GS-1 (YD) (ref. 54) and glacial conditions55, the cold Fennoscandinavian Ice Sheet induces a strong southwesterly wind flow over the Norwegian Sea associated with blocking circulation in turn gradually promotes deep-water formation in the Norwegian Sea55. Our data and other existing records all lend support to the occurrence of this proposed mechanism midway through GS-1 (YD) (i.e., stronger southwesterly winds56,37, warmer surface ocean conditions58,59, and gradually less frequent sea-ice occurrence58–59) (Supplementary Fig. 7).

In conclusion, our results suggest that gradual changes in high-latitude NADW formation are precursors of rapid climate shifts in the North Atlantic and emphasize the central role of ocean circulation in abrupt climate change, as well as its sensitivity to atmosphere and cryosphere dynamics. Given recent evidence that the current AMOC has been slowing down for several decades60,61, our findings broach the question as to whether the current decline in deep-water circulation may herald a new phase of abrupt change.

Methods

Chronology. The chronology of core MD99-2284 was established by aligning variations in downcore sea-surface temperatures (SST)62,63 with synchronous downcore δD records from the ancient lake of Atteköp, southern Sweden63 (Supplementary Figs. 1–5 and Note 1). The terrestrial site is located closely downwind of our marine core location in an area where most of the precipitation is sourced from the North Atlantic rainfalls64, and where hydroclimate shifts have a demonstrated exchange between calcite and the carbonate ions in pore waters83, which is not necessarily representative of near-surface paleoceanographic conditions in the North and Norwegian Seas65. Synchrotron-based time series and regional proxy time series were obtained using an automated stratigraphic alignment algorithm driven by a Markov chain Monte Carlo method12,66 run for 1012 iterations. The approach involves nonlinear deformation of the SST time series onto the reference δD time series to deliver an optimal alignment accounting for uneven compaction and/or expansion of sediments over time, as well as for analytical errors associated with the proxy data (Supplementary Fig. 3). An account of the mathematical formulation associated with the algorithm is presented in ref. 67.

Atteköp’s chronology was constructed using a Bayesian age model based on the Hässeldalen Tephra and 37 AMS 14C dates from selected terrestrial plant macrofossils66 (Supplementary Fig. 3) calibrated with the IntCal13 curve14, which in turn allows placing MD99-2284 records on the atmospheric IntCal13 timescale. However, the alignment between the marine and terrestrial records was done only up to 12,600 years BP due to the scarcity of datable plant macrofossils in Atteköp in this portion of the core (Supplementary Fig. 3). To secure the chronology upcore, we thus employed three well-dated tephra layers isolated in MD99-2284 sediments—the Vedde Ash, the Abernethy Tephra, and the Saksunarvatn Ash (Supplementary Note 2)—and fit the tephra-based age constraints using a Gaussian Regression model68.

Tephra analysis. Two important regional isochrones, i.e. the Vedde Ash and the Saksunarvatn Ash (both present in Greenland ice cores69), have been previously reported in core MD99-2284 (ref. 70). These horizons have been precisely and accurately radiocarbon dated in Lake Krakenes, Norway70. The markers are here supplemented by the first reported marine occurrence of the Abernethy Tephra72,73 (Supplementary Fig. 4). The finding constitutes the outcome of a targeted search for late YD tephra based on the counting of glass shards within the interval 200–360 cm in the marine core. A peak in cryptoparticles was identified in samples of 200.5 and 249.5 cm, with the latter lying above the sediment following the methods of ref. 73. Samples were treated with 10% H2O2, and heated to remove organic material, then washed with deionized water over a 63-µm sieve. A series of heavy liquid density separations using sodium polytungstate were then performed to isolate material between 2.2 and 2.5 g cm−3. The fraction was prepared on 27 × 40 cm glass slides in each replicate to expose grain interiors and analyzed at the Concord University Microanalytical Laboratory (WV, USA) using an ARL SEMQ electron microprobe equipped with six wavelength-dispersive spectrometers and a Bruker 3000 SDD energy-dispersive spectrometer. Instrument conditions included a 14 kV accelerating voltage, 10 nA beam current, and beam size of 4 to 6 µm. Results were reported as non-normalized major oxide concentrations (Supplementary Data 1).

Results from the analysis of tephra grains in samples 247.5 cm and 249.5 cm show a single geochemical composition of rhyolitic composition similar to the Abernethy Tephra. The Vedde Ash erupted from the Katla volcano in the middle of the YD. Recently, it has been shown that there was a second tephra eruption of Katla within the YD, but closer to the YD-Holocene boundary named the Abernethy Tephra72, Tephra layers from Loch Etteridge (LET-5)74 and Abernethy Forest (AF55)75, in Scotland, were attributed to this eruption, which are both similar in stratigraphy and geochemistry to tephra that we have identified (Supplementary Fig. 4). Based on petrographic counting and other lines of evidence47, it has been argued75 that the Abernethy Tephra is likely 320 ± 20 years younger than the Vedde Ash (12,064 ± 48 years BP)70. Therefore we here assigned to the layer an age of 11,744 ± 50 years BP. Radiocarbon dating of surface using on one terrestrial plant macrofossil from a sediment sequence in Abernethy Forest yields a calibrated age range of 12,380–11770 years BP, which is consistent with, or slightly underestimates our age assignment.

14C dating. Well-preserved monospecific shells of the planktic foraminifer Neogloboquadrina pachyderma (s) (calcification depth ~30 to 200 m (ref. 79)), two samples of the benthic foraminifer Cibicidoides wuellerstorfi, one shell of Pyrgo maritima, and a number of mixed benthic foraminifera were hand-picked from slices of core MD99-2284 for AMS 14C dating for a total of 88 measurements (41 planktic and 47 benthic), including 8 previously reported dates80 (Supplementary Data 2). Mixed benthic samples are mainly composed by C. neoteretis excluding Pyrgo spp and milioloid spp. Samples were mainly selected from the >150 µm fraction, except for 31 samples, which were picked from the 63-150 µm fraction. All shells were cleaned and dried prior to preparation for accelerator mass spectrometry (AMS) measurements, whereas large samples were additionally rinsed with dilute, organic-free HCl. Large samples were converted to graphite and analysed using standard AMS 14C measurement procedures within the laboratories of Beta Analytic Miami (FL), USA, and ETH Zurich, Switzerland. All the small fraction samples were processed and analysed using a newly developed method77. The approach involves direct analysis of CO2 from ultra-small amounts of carbon in a compact AMS facility equipped with a gas ion source at the Laboratory of Ion Beam Physics, ETH Zurich. To test for contamination by secondary carbonates on the small fraction samples, we performed leaching experiments on the sample material surface using HCl 0.02 M following the procedure outlined in ref. 78. All radiocarbon ages are here reported according to the standard protocol of ref. 79.

A subset of 7 samples were prepared at the University of Cambridge using methods detailed in ref. 80 and subsequently dated by AMS at the 14 Chrono Centre, University of Belfast.

The resulting 14C ages as based to generate surface/subsurface (~100 m) and bottom water (1500 m) ventilation records employing the random walk model described below, and ultimately to reconstruct benthic-planktic (B-P) ventilation ages81. We identified seven outliers that were dismissed from our bottom water ventilation record (one in GS-1 at 400.5 cm; three in GI-1 at 415, 445.5, 534.5 cm; three in GS-2 at 575.5, 581.5, 585.5 cm). The seven benthic ages based on benthic foraminifera, including one Pyrgo spp age, which has been shown to yield old 14C ages12. The other benthic dates, which are all based on small shell fragments <1 mg in weight and all associated with the 63-150 µm fraction, were either significantly younger than the corresponding planktic ages (where available), or significantly younger than the contemporaneous atmosphere. Acid leaching tests indicate that four out of the seven outliers are likely affected by modern CO2 carbon contamination, as evidenced by the high 14C content in the respective leach fractions (Supplementary Data 2). Given that sedimentation rates at our site are remarkably high (~400 cm kyr−1), we argue that complications associated with bioturbation and differential mixing effects are likely negligible. Rather, we argue that the systematically younger ages measured in these samples may be the result of preferential modern-carbon contamination among the finer carbonate fraction through secondary growth of calcite within the sediment pore waters and exchange between calcite and the carbonate ions in porewaters82.

Random walk model. Regional reservoir (AR) and reservoir (B-Atm and P-Atm) estimates were inferred using a random walk model that incorporates both the uncertainties in the calibration curve and our 14C AMS measurements (Supplementary Data 3). B-P estimates were obtained by subtracting the modelled B-Atm and P-Atm curves. This approach was preferred to the projection method, which is more influenced by atmospheric Δ14C variations in the North Atlantic84. In the following, we provide a brief overview of the random walk model and the construction of the reservoir curve.

For each of our N ocean objects, we observe an estimate \( \hat{R}_i \) of its true calendar age \( R_i \) and a radiocarbon determination \( z_i \) that is also subject to noise. This
radiocarbon determination can be decomposed into the site-general marine radiocarbon age \(m(\theta)\), and \(\Delta R(\theta)\) the local reservoir variation i.e., difference between the reservoir age \(R(\theta)\) associated with \(\theta\) with the sea surface boundary condition to estimate the Norwegian Sea reservoir age \(\Delta R(\theta)\) given these paired observations \((t_i, z_i)_{i=1,...,N}\) and our prior knowledge about the value of \(\Delta R(\theta)\) provided by the Marine13 radiocarbon calibration curve (ref. 14). We then infer the value of the absolute Norwegian Sea reservoir correction \(R(\theta)\) by comparison with the atmospheric radiocarbon curve at the same \(\theta\). Specifically we model:

\[
\Delta R(\theta) = R(\theta) = m(\theta) - \Delta R(m(\theta)) + \xi(i, t_i, z_i, \ldots, \theta),
\]

where \(\xi(i, t_i, z_i, \ldots, \theta)\) are the uncertainties in our measurement of calendar age and radiocarbon age assumed here to be independent and identically distributed (IID) with mean 0 and variances \(\sigma_i^2\) and \(\sigma_{\xi}^2\), respectively. Note in particular that the true calendar ages \(\theta\) are only subject to noise to this should be incorporated into any estimation procedure. If we ignore this additional uncertainty we are likely to introduce a bias in our estimate for \(\Delta R(\theta)\) and oversmooth. We incorporate a Bayesian approach for our estimation of \(\Delta R(\theta)\). This allows us to incorporate our priors about the value of \(\Delta R(\theta)\) and model the evolution of \(\Delta R(\theta)\) over time in a physically interpretable way; and update these prior beliefs in a consistent framework. To make our problem identifiable we require strong prior information on the site-general \(m(\theta)\). This is provided by Marine13, the marine radiocarbon calibration curve (ref. 14). At any calendar age \(\theta\) this provides pointwise estimates of the value

\[
m(\theta) = \mu_0 + \sigma_0 \theta^m,
\]

Additionally, we require a prior model for the evolution of the Norwegian Sea’s reservoir age \(\Delta R(\theta)\). While we do not expect this to be constant over time, we would expect that knowing \(\Delta R(\theta)\) at a particular \(\theta\) provides some information about the likely variation for \(\theta\). We therefore, place a Wiener process (random walk) prior on \(\Delta R(\theta)\) similar to that used in the modelling of the IntCal atmospheric calibration curve (see refs. 85–87) whereby

\[
\Delta R(\theta) = \frac{\eta(\theta)}{\sigma_{\eta}(\theta)}
\]

The above random walk model is then combined with our prior on \(m(\theta)\) and our observed \((t_i, z_i, \ldots, \theta)\) through a Metropolis-within-Gibbs sampler. This provides a posterior estimate for \(\Delta R(\theta)\) that allows us to borrow strength from the surrounding observations (i.e., neighbouring \(\theta\)). It also updates our site-general marine \(m(\theta)\). Combining both the estimate of \(\Delta R(\theta)\) and the updated \(m(\theta)\) and comparing with the atmospheric radiocarbon levels provided by IntCal13 (ref. 14) then allows us to estimate the absolute Norwegian Sea reservoir correction \(R(\theta)\), i.e., B-Atm and P-Atm. Details on the implementation of this sampler can be found in Supplementary Methods.

10Be measurements. A total of 170 samples from 1793 to 2279 m depth in the WAIS Divide 06A ice core (WD-06A) were analysed for 10Be concentrations at UC Berkeley and the Purdue University PRIME Lab (Supplementary Data 4). Samples typically represent continuous ice core sections of ~3 m length (although samples 700 g were weighed, melted, and acidified as eluting solvents hexane-saturated columns using as eluting solvents hexane-saturated columns using as eluting solvents hexane-saturated columns using an internal standard, and each added to samples before GC analysis. Similarly, IP25 and 9-OHD (over a range of isomers) were calculated based on the comparison of the adjusted peak areas relative to both C36 alkane and 5-fluorotoluene, which were quantified by gas chromatography on a Thermo-TRACE Ultra Gas Chromatography Flame Ionization Detector (GC-FID) equipped
with PTV injector operated in splitless mode. Samples were measured with a DB-1 column (60 m, 0.25 µm ID, 0.1 µm film thickness). Lastly, an in-house alkenone standard with known relative concentrations of C25-unsaturated alkenones was injected every six samples to monitor the instrument performance and analytical precision of the alkenone-unsaturated index \(IP_{25}\) (ref. 99) (1± = 0.0002).

Changes in sea-ice cover were finally reconstructed by calculating a phytoplankton-IP25 index \(IP\) (Fig. 2) following the equation

\[
IP_{25} = \frac{IP_{25}}{|P|} = \frac{|IP_{25}|}{|P|} \times C
\]

whereby \(|IP_{25}|\) and \(|P|\) represent the respective concentrations of \(IP_{25}\) and the phytoplankton biomarker \(P\), while \(C\) is a balance factor calculated from the ratio of the mean \(IP_{25}\) to \(P\) concentrations.

**Breakpoint analysis.** To estimate the timing of the transitions recorded in the proxy data, we used a Monte Carlo version of the function Segmented in the R package ‘Segmented’100, i.e., a piecewise linear fitting regression method that determines the breakpoints of two lines. To combine the analytical and age uncertainties associated with each proxy record, while at the same time enforcing monotonicity in the age estimates, the data sets were first modelled using the Random Walk Model described above, whereby the model was fitted directly to the observations. We subsequently randomly resampled (5000 times) the modelled data within its uncertainty, assuming a Gaussian error distribution. We then estimated the breakpoint structure of each Monte Carlo realization using the Segmented function run with 100 bootstrap iterations. The inferred likelihood distribution of each breakpoint allowed us to quantifying the timing uncertainty for the start and end of the transitions under investigation (Fig. 4; Supplementary Table 1).

**Code availability**

All the climate model output and R codes used for the numerical procedures are available from the main author upon reasonable request.

**Data availability**
The source data are available from the Dryad Digital Repository, doi:10.5061/dryad.7h9m0, and from the PANGAEA paleoclimate data archive.

Received: 30 September 2019 Accepted: 26 February 2019
Published online: 20 March 2019

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Acknowledgements

We are grateful to D.I. Blindheim for assistance with foraminifera sampling, H. Sadatzky for assistance with sampling for lipid biomarker analyses, and S. Rasmussen for helpful discussions. The research leading to these results has received funding from the European Research Council under the European Community’s Seventh Framework Programme (FP7/2007-2013)/ERC grant agreement 610055 as part of the ‘ice2ice’ project. The 10Be measurements in the WAIS Divide core were funded by NSF grants ANT 0839042 (to M.W.C) and 0839137 (to K.C.W.). Additional funding by the Columbia Climate Center of Columbia University (to F.M and W.J.D) and the Vetlesen Foundation (W.J.D) is gratefully acknowledged. This study is a contribution to the INTIMATE project.

Additional information

Supplementary Information accompanies this paper at https://doi.org/10.1038/s41467-019-09237-3.

Competing interests: The authors declare no competing interests.

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Journal peer review information: Nature Communications thanks Lowell Stott and Anders Svensson for their contribution to the peer review of this work. Peer reviewer reports are available.

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