

Title: Weakening of AMOC linked to past Greenland Ice Sheet retreat

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Abstract: A weakening of the Atlantic Meridional Overturning Circulation (AMOC) is predicted to occur under multiple scenarios of future warming. However, the effect of meltwater from a decaying Greenland Ice Sheet (GrIS) on AMOC is uncertain. Using a basin-wide network of North Atlantic sediment cores, we show that an interval of largescale melting of the GrIS during a previous interglacial (Marine Isotope Stage 11c (MIS 11c); 397 – 427 ka) was associated with a millennial-scale, $30 \pm 15\%$ weakening of AMOC and an associated abrupt 2 – 7°C cooling event over the subpolar North Atlantic. We reveal that this AMOC weakening occurred due to a combination of a weakening in both upper and lower North Atlantic Deep Water (respectively also termed Labrador Sea Water and the downstream Nordic Overflows). Based on available paleoclimate datasets, this re-arrangement of ocean circulation had a prominent impact on terrestrial ecosystems and atmospheric conditions in western European and the Northwest African margin while the evidence for global climate impacts is equivocal. This study suggests that modern oceanographic conditions are analogous to those immediately prior to the most extreme AMOC weakening event in MIS 11c, with modern rates of GrIS melt being comparable to those modelled for MIS 11c. Critically, these findings provide empirical evidence that suggests the potential vulnerability of AMOC weakening to sustained GrIS decay.

Main text:

Introduction

Continued global warming will push components of the Earth system closer to tipping points that, if crossed, may trigger abrupt change (1). One of the most widely discussed tipping points is the potential weakening, or even collapse to a weak state (below ~ 5 Sv (2)), of the Atlantic Meridional Overturning Circulation (AMOC), caused by increased freshwater fluxes into the high latitude North Atlantic through either high precipitation levels and/or and the melting of high latitude sea-ice and glaciers (1,3,4). Paleoclimate evidence and modelling studies suggest that an AMOC collapse would have major climatic and socioeconomic impacts (5,6), including strong cooling over much of the Northern Hemisphere, and shifts in rainfall patterns over west Africa and central Asia (7,8). Future AMOC weakening is, thus, a key concern for global policymakers and stakeholders. Researchers have largely focussed their attention on abrupt events during glacial stages (*e.g.* Dansgaard-Oeschger events), which have been related to shifts in the mode of operation of AMOC (9), but as we are presently in an interglacial stage these glacial events are less relevant as analogues for constraining modern/future AMOC behaviour. It is therefore imperative to also examine AMOC and its response to forcing during warm climates, such as previous interglacials. In this study, we investigate the response of AMOC and its key components to a period of increased freshwater flux derived from the melting of the Greenland Ice Sheet (GrIS) during Marine Isotope Stage 11c (MIS 11c).

There is uncertainty regarding both the magnitude and timing of any future weakening or collapse in AMOC (3,4,10,11). It has been suggested (and countered (12)) that AMOC could collapse to a weak state by the middle of the 21st century (3,4); however, Intergovernmental Panel on Climate Change (IPCC) projections do not forecast AMOC collapse under any scenario prior to 2100 (13)(medium confidence). Yet, further analysis of extended model simulations shows that many of these IPCC runs result in AMOC collapse by 2200 (2). Uncertainty in the robustness of the climate models used in the IPCC reports in part derives from the model's lack of accurate representation of key ocean features such as overflows (14), but also reflects that fact that models are typically biased towards stability (15,16). Furthermore, most models do not incorporate interactive ice sheets into their representation of the Earth system and thus cannot fully account for changes in meltwater runoff from the GrIS (17–19), which has been found to be melting 20% faster than previously thought (20). Our understanding of the impact of GrIS melt on AMOC also remains poorly constrained: meltwater experiments produce a wide range of estimates for the rate of GrIS melt necessary to impact AMOC strength (*e.g.* meltwater of 0.005 – 0.1 Sv), further complicated by differing impacts dependent on meltwater spreading pathways (21–24).

Analysis of empirical data derived from intervals of GrIS melt in the geological past can be used to investigate the potential response of AMOC and its components to freshwater release derived from this source. Evidence from the current interglacial suggests that the modern GrIS extent was reached ~ 7000 years ago, with minimal changes between then and the industrial era (20th century) (25,26). Contrastingly, there is evidence from sediment provenance and pollen data which indicate that a major reduction in GrIS extent relative to today occurred during MIS 11c (397,000 – 427,000 years ago) (27,28), when global surface temperatures were $\sim 1^\circ\text{C}$ warmer than the pre-industrial era but with concentrations of atmospheric CO_2 similar to the pre-industrial (~ 280 ppmv) (29,30). This warm period also has more convincing evidence for an abrupt climate event in northern Europe (31,32) and the North Atlantic (33–35) than any pre-Holocene interglacial, testifying to the presence of triggers of climatic instability at this time. A recent tephra-based correlation between a European varved lake (Marks Tey) and North Atlantic marine sediment core (ODP Site

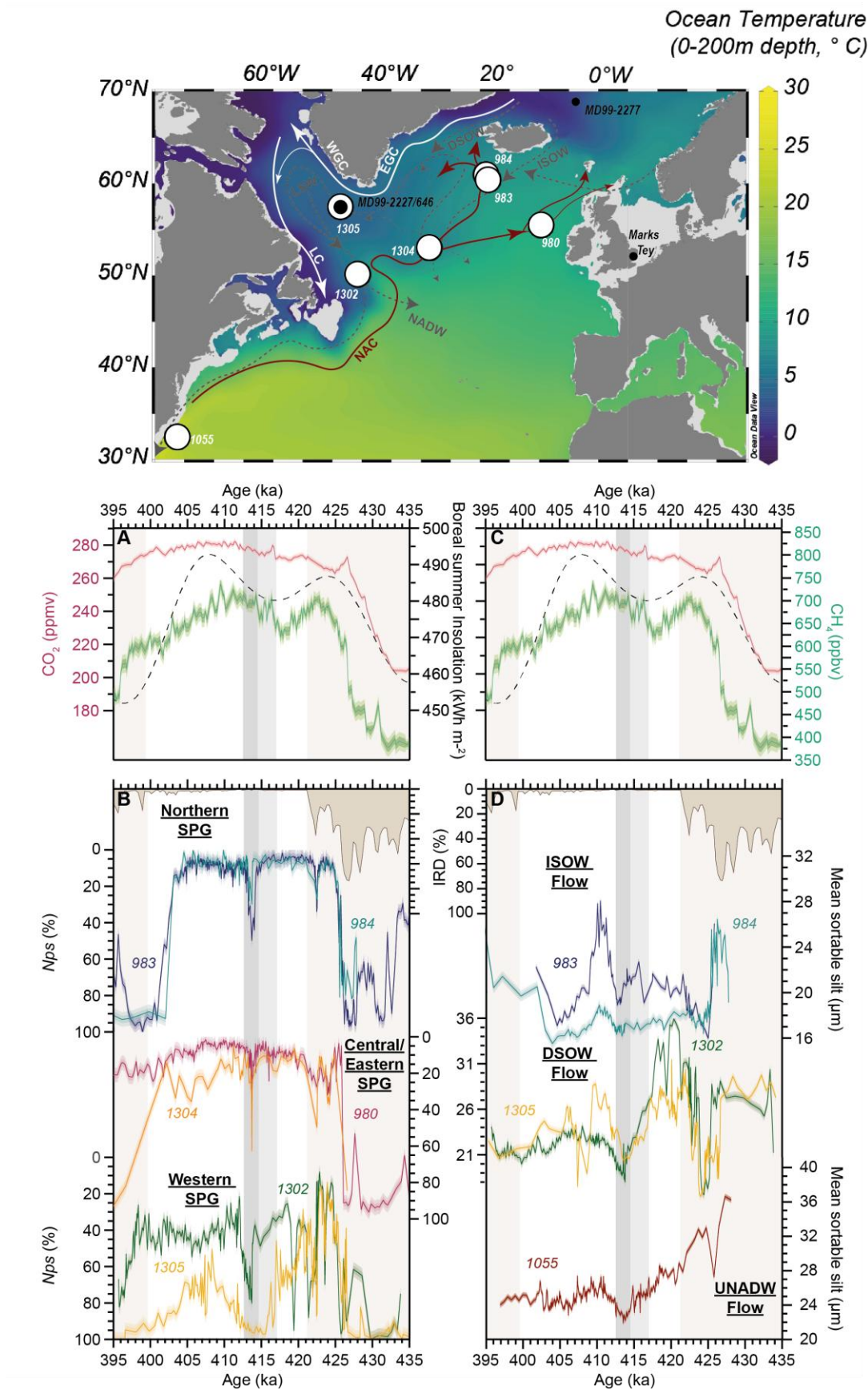


Fig. 1. Core locations and data synthesis compared to key climate forcings: (Top) IODP Site U1302 (50.2°N, 45.6°W; 3250m depth) (this study); IODP Site U1305 (57.5°N, 48.5°W; 3463m depth) (this study, (38)); ODP Site 983 (60.4°N, 23.6°W; 1983m depth) (this study, (35)); ODP Site 984 (61.0°N, 24.0°W; 1650m depth) (this study); IODP Site U1304 (53.1°N, 33.5°W; 3024m depth) (this study); ODP Site 980 (55.5°N, 14.7°W; 2170m depth) (this study, (61)); IODP Site 1055 (32.47°N, 76.2°W; 1800m depth) (this study); projected on a map of North Atlantic Ocean Temperatures (°C, 0 – 200m depth) using data from WOA23 plotted in Ocean Data View. Study core sites are depicted as white circles with black outlines. Key comparison sites in this study are depicted by black circles. Key surface (solid) and intermediate/deep (dashed) ocean currents are depicted. (A, C) Key climate forcings (duplicated above each dataset) including greenhouse gas data (CO₂ and CH₄) from EPICA Dome C (30) and summer insolation in kilowatt-hours per square metre (kWh m⁻²) at July 65°N (62). (B) % *Neogloboquadrina pachyderma* (Nps) records compared with compiled normalised IRD from study sites. (D) Mean sortable silt in microns (µm) data compared with a compiled normalised Ice Rafted Debris record from study sites. Coloured bars: pale brown at onset and end of MIS 11c = ice rafting intervals; light grey = broad mid-interglacial cooling; dark grey = abrupt mid-interglacial cooling. DSOW = Denmark Strait Overflow Water; ISOW = Iceland Scotland Overflow Water; LSW = Labrador Sea Water; NADW = North Atlantic Deep Water; NAC = North Atlantic Current; EGC = East Greenland Current; WGC = West Greenland Current; LC = Labrador Current; SPG = Sub-Polar Gyre; UNADW = Upper North Atlantic Deep

980) (36) has shown that an abrupt centennial-scale decline in air temperature over northwest Europe was synchronous with a surface ocean cooling in the northeast Atlantic. Furthermore, records from across western and central Europe show this cooling triggered transformative ecological shifts (32). Currently, it is unclear whether this interglacial cold event was caused by AMOC weakening, potentially linked to GrIS melt, and if so, how the various components of AMOC contributed to the event.

To investigate the relationship between GrIS wastage, abrupt climate change and AMOC weakening during MIS 11c, we generated a suite of palaeoceanographic proxy data, including sortable-silt and %*Neogloboquadrina pachyderma* (Nps) records from 7 International Ocean Drilling Program (IODP) sites at key locations across the subpolar North Atlantic (see Methods) (Fig. 1). All sites were placed on a common stratigraphic age model via alignment of each site's Relative PalaeoIntensity and benthic foraminiferal oxygen isotope record (see SI). Quantitatively calibrated downcore proxy changes are reported relative to the modern ocean by comparison to modern core top values.

Planktic foraminifera census counts were used to reconstruct upper ocean temperature changes across the subpolar North Atlantic (Fig. 1C). All sites reveal a prominent abrupt cooling event (of 2 – 7°C) centred at ~413 ka, with a longer duration and more severe cooling in the central Labrador Sea, consistent with previous studies (33,35,37,38). This event was not associated with ice rafted detritus, unlike an earlier cold event at ~422 ka, which was associated with the latter stages of the glacial termination (Fig. 1B). It has been proposed that the surface ocean temperature anomaly of the Subpolar Gyre (SPG) region can be used as proxy for AMOC strength (39,40): the AMOC temperature fingerprint (see Methods). Accordingly, we combine our surface ocean temperature records to generate a temperature-based record of AMOC strength (AMOC_{SPG-T}; Figure 2D). Our AMOC_{SPG-T} record reveals a long-term weakening of AMOC from ~420 – 413 ka, with a more abrupt weakening centred at ~413 ka, representing a 50± 15% (1σ) decline in AMOC strength from the strongest (+29% relative to modern) to weakest (-24% relative to modern) state during MIS 11c (~420 – 413 ka). Of this 50% decline in AMOC 25% occurred in association with an abrupt weakening at 413 ka. An earlier smaller (~17%) transient weakening also occurred at 417 ka, though this barely exceeds our uncertainty estimates. We note that when examining the 413 ka event in the raw datasets, as opposed to the smoothed averaged records, the major cooling in surface ocean temperature occurred within 100 years or less (Fig. 1A; SI Table 7).

Additional evidence for changes in AMOC during the mid-MIS 11c cold event is gained by employing the mean sortable silt ($\overline{SS}$) grain size proxy to reconstruct deep water flow speeds at our core sites (Fig. 1D). These are located under the pathway of key AMOC components: (1) the two major Nordic Overflows Iceland Scotland Overflow Waters (ISOW) and Denmark Strait Overflow Waters (DSOW), hereafter collectively referred to as Lower North Atlantic Deep Water

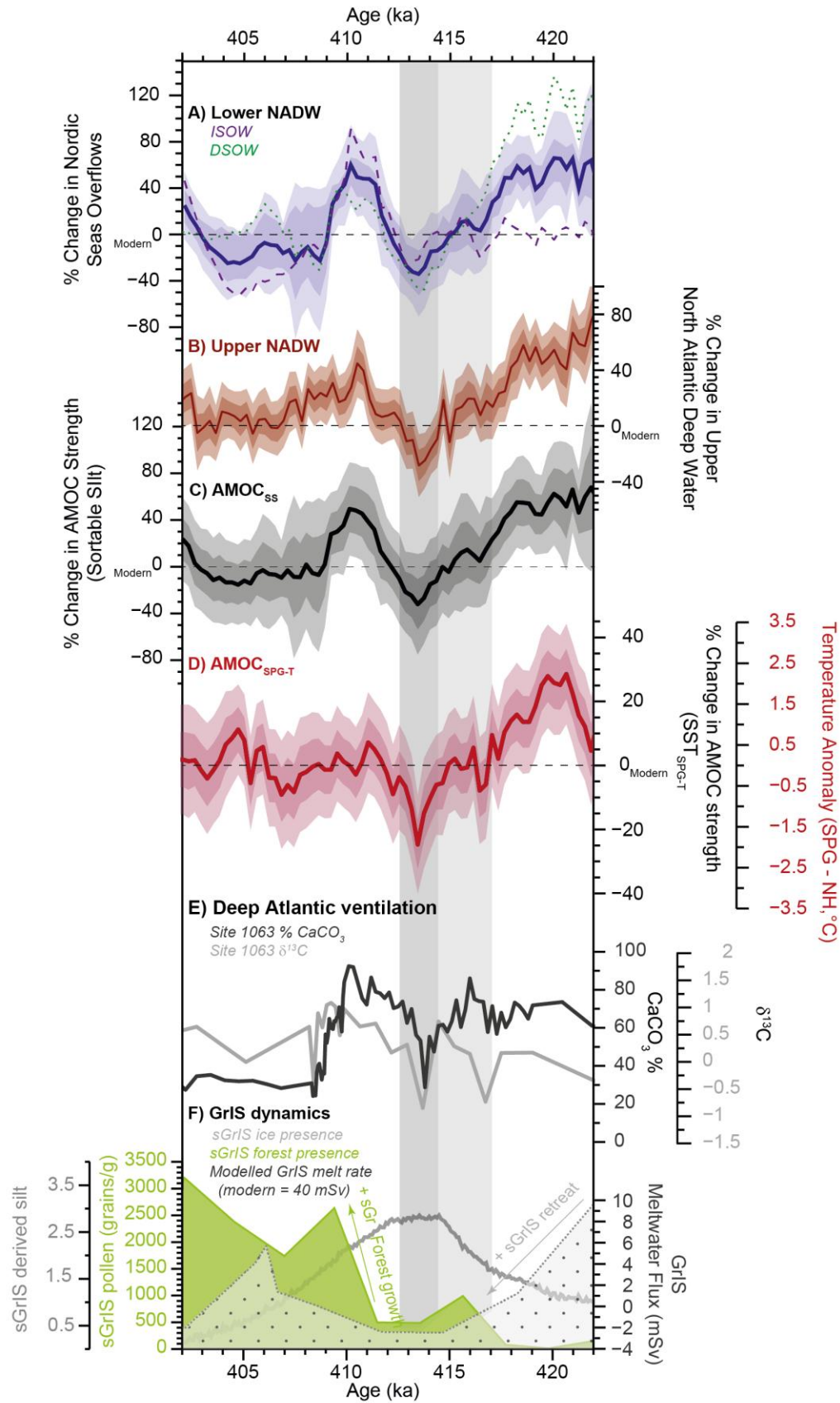


Fig.2. Physical and Index-based reconstructions of AMOC compared to potential forcing mechanisms during the main interglacial period of MIS 11c: (A): Lower North Atlantic Deep Water (blue solid line) reconstructed from grain size records of Iceland Scotland Overflow Water (ISOW – purple dashed line) (ODP Sites 983 and 984) and Denmark Strait Overflow Water (DSOW – green dotted line) (IODP Sites U1302 and U1305). (B): Upper North Atlantic Deep Water (maroon solid line) reconstructed from the grain size record of ODP Site 1055. (C): Compiled Physical (grain size) based record of AMOC across the main warm interval of MIS 11c ($AMOC_{SS}$ – black solid line) assuming equal contribution of each AMOC component; (D): Compiled Index (ocean temperature) based record of AMOC across the main warm interval of MIS 11c ($AMOC_{SPG-T}$ – red solid line); Records A-D presented with reference to modern (0) with modern indicated by a black dashed line; (E): Proxy records (Benthic $\delta^{13}C$ and % $CaCO_3$) of changes in the Deep Western Boundary Current (DWBC) from ODP Site 1063 (46,63). % $CaCO_3$ was calculated using equations from (64); (F) Greenland ice sheet (GrIS) dynamics derived from $CaCO_3$ -free silt flux (grey dotted polygon) (27) and pollen concentrations (green polygon) (28). Modelled GrIS meltwater flux was derived from (65) (grey solid line) and has been temporally adjusted within age model uncertainties so that the highest rates of meltwater flux coincide with the smallest sGrIS extent. Note this has been inverted. Coloured vertical bars: light grey = broad mid-interglacial cooling; dark grey = abrupt mid-interglacial cooling. Shading around compiled records indicate 1σ (darker) and 2σ error (lighter).

(LNADW); and (2) Labrador Sea Water, hereafter referred to as Upper North Atlantic Deep Water (UNADW). We combine these to form a physical circulation-based proxy of AMOC strength ($AMOC_{SS}$; Figure 2D, see methods) to compare to the $AMOC_{SPG-T}$ record.

Overall, the two independent AMOC reconstructions show good agreement in both the overall trends and magnitude of AMOC strength changes, particularly during intervals of weakening (see SI Fig. 3). Both $AMOC_{SS}$ and $AMOC_{SPG-T}$ indicate that AMOC during MIS 11c was often stronger than modern (Fig. 2C and 2D), consistent with previous studies that have proposed broadly stronger NADW formation during MIS 11c, based on changes in deep ocean chemical proxies (41,42). Despite the broad agreement, there are some differences in AMOC strength between the two reconstructions, such as at 410 ka and 405 ka. These differences may be due to $AMOC_{SPG-T}$ also being affected by regional heat divergence or convergence depending upon the strength of the subpolar gyre and its control on northward heat transport into the Nordic Seas and Arctic region, which continued during MIS 11c (43–45).

Figure 2 shows that the abrupt weakening recorded by our $AMOC_{SPG-T}$ proxy at 413 ka was accompanied by a 38% weakening of $AMOC_{SS}$ (reaching -32% relative to modern strength at its lowest; Fig. 2C). This was preceded by a multi-centennial $18 \pm 18\%$ weakening of $AMOC_{SS}$ at 417 ka, which occurred against a background of strong $AMOC_{SS}$ (+23% relative to modern), which was also associated with a weakening in $AMOC_{SPG-T}$. Weakening of AMOC is also suggested by reconstructed declines in benthic $\delta^{13}C$ and $CaCO_3$ preservation in the deep Northwest Atlantic at sites typically bathed by NADW during interglacials, suggesting the weakening and/or shoaling of NADW (46,47) (Fig. 2C). Therefore, there is evidence of a substantial AMOC weakening during mid-MIS 11c from both surface and deep ocean proxy records that persisted for ~1000 years before recovering.

Differing behaviour of key AMOC components

Before exploring how different individual AMOC components contributed to the minimum in AMOC strength at ~ 413 ka, we first explore and explain the behaviour of each of LNADW (Fig. 3) and UNADW (Fig. 2B) through MIS 11c, and in doing so provide the context necessary for developing a mechanistic understanding of the abrupt MIS 11c event. Figure 3 presents our $\overline{SS}$ based constraints on the two Nordic Seas overflows (LNADW), revealing that they exhibited different behaviour during MIS 11c, consistent with similar findings from the late Holocene (48) and instrumental period (49). Our data reveal that ISOW was anomalously strong (+ $91 \pm 12\%$ relative to modern) during mid-MIS 11c (412 – 409 ka) (Fig 3B). Previous Holocene work has shown that ISOW reached a maximum strength once remnant ice sheets had melted away and insolation changes reduced Arctic sea-ice export to the Nordic Seas, thus allowing strong Nordic

Seas deep convection (50). Our new results show similar behaviour during MIS 11c with a prominent maximum in ISOW strength occurring during the later MIS11 maximum in 65°N summer insolation, when proxy data from the Nordic Seas indicate changes in the water column that favour increased convection (51,52), thus promoting stronger ISOW production (Fig 3C and 3D, and Methods/SI). We therefore suggest that the decline in ISOW strength following this mid-interglacial peak was related to declining boreal summer insolation.

In contrast to ISOW, DSOW was anomalously strong ($+136\% \pm 16\%$ relative to modern) during early MIS 11c (423 – 417 ka; Fig 3F). We propose that this was due to the effects of the remnant Laurentide Ice Sheet (LIS) on (i) atmospheric circulation, and (ii) subsurface ocean density gradients. Model simulations show that a remnant LIS can promote more northerly winds in the western Nordic Seas (53). Modern physical oceanographic observations (54) demonstrate: (i) stronger northerly winds enhance onshore Ekman transport which limits the spreading of fresh surface waters and sea-ice; (ii) this facilitates severe heat loss from the ocean offshore of the ice edge which increases the density of the Return Atlantic Water that feeds DSOW and thus can affect the lower limb of AMOC. These modelling studies (53) are unlikely to fully capture these small-scale frontal processes due to its relatively low spatial resolution ($1.25^\circ \times 1.25^\circ$). Equally, model simulations have also shown that increased meltwater derived from the decaying LIS can reduce the density of the intermediate depth ocean in the Labrador and Irminger Seas, thus increasing the density gradient across the Denmark Strait and driving a stronger DSOW (55). The result of these two factors would be a substantially strengthened DSOW early in an interglacial during greater LIS extent which gradually weakens to modern values as the LIS retreats, in agreement with the proxy data (Fig. 3G and 3H).

The remnant LIS likely also strengthened UNADW during the early interglacial (Fig. 2B). The presence of ice sheet topography leads to changes in wind stress curl which promote a stronger SPG, greater doming and outcropping of isopycnals that both limits the spread of freshwater into the centre of the gyre, and preconditions the water column for convection (53,55). The strengthened SPG circulation caused by a remnant LIS (53) may also have contributed to the stronger DSOW during early MIS 11c by driving an increased Atlantic Inflow to the Nordic Seas, which is densified upon its transit around the boundary rim of the Nordic Seas as Return Atlantic Water, feeding DSOW (56). Furthermore, we propose that the cold, dry continental winds coming off the LIS would have triggered intense wintertime ocean cooling and densification in the subpolar region, increasing the formation of UNADW.

It has been inferred that a relatively strong AMOC is a recurrent feature of early interglacial climates (35). Because this early interglacial AMOC ‘anomaly’ persists for millennia, it is not attributable to a multi-centennial scale AMOC-overshoot mechanism caused by preceding salinity build-up; instead, it was associated with an abrupt switch of the AMOC to an interglacial mode while the broader climate system had not yet re-equilibrated following deglaciation (35). Our new reconstruction of the individual deep AMOC components, and inferred mechanisms of change linked to the impact of the remnant LIS, provide further insight into this early interglacial AMOC behaviour.

Contributions to AMOC weakening

Next, we explore the contribution of each AMOC component to AMOC weakening during mid-MIS 11c. Figure 3B shows that there were two transient ISOW weakening events at 417 – 415 and 414 – 412 ka (each of approximately 30%). In isolation, neither event was sufficient to drastically weaken overall AMOC strength; rather, it is because these events occurred in addition to a long

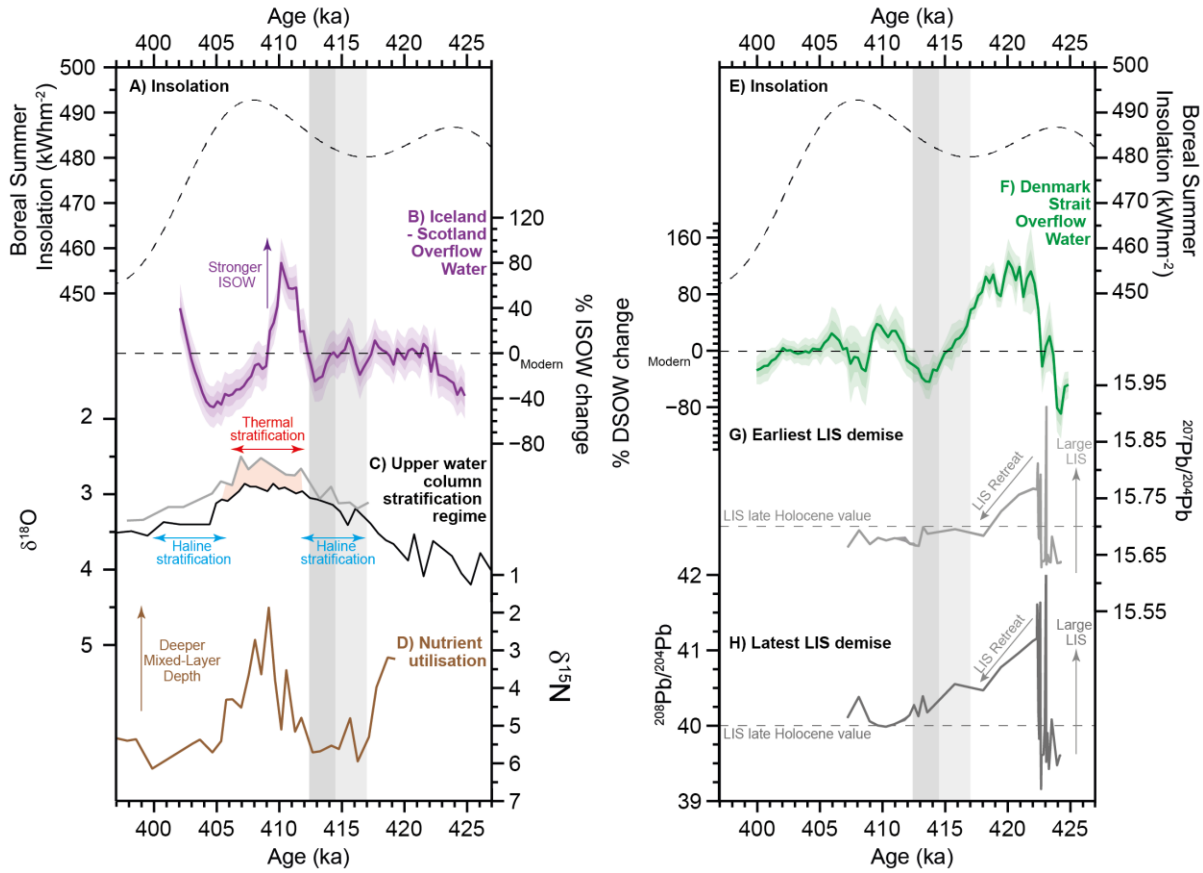


Fig. 3. Records of Nordic Seas overflows compared to key forcings across MIS 11c: (66). (A-D) Compiled Iceland Scotland Overflow Water (ISOW) record compared to forcing mechanisms. (A and E): Boreale summer insolation was obtained from (62) in kilowatt-hours per square metre (kWh m^{-2}) at July 65°N . (B): Compiled ISOW record (purple) derived from ODP Sites 983 and 984 (C): Oxygen isotopes derived from planktic foraminifera *Neoglobobulimina pachyderma* (black) and *Turborotalita quinqueloba* (grey) (43). Highlighted in red are consistent (multi-datapoint) offsets of $\geq 0.4\text{‰}$, indicating a thermal regime favouring increased convection and stronger ISOW. (D): Nitrate utilisation ($\delta^{15}\text{N}$) indicates reduced utilisation during the mid-late interglacial, indicating increased mixed-layer depth, favouring convection (50); (E-H) Compiled Denmark Strait Overflow Water (DSOW) record compared to forcing mechanisms; (F): Compiled DSW record from IODP Sites U1302 and U1305 (green); (G-H): Lead isotope data indicates the earliest (G) and latest (H) timing of the Laurentide Ice Sheet retreating to the late Holocene extent (Canadian Archipelago) on the age model of this study (late Holocene levels = dashed line). The larger Laurentide Ice Sheet coincides with strongest DSW. Panels B and F are presented with reference to modern (0), indicated by a dashed black line for ISOW DSW. Coloured bars: light blue = broad mid-interglacial AMOC weakening; dark blue = abrupt mid-interglacial AMOC weakening. Shading around compiled records indicate 1σ (darker) and 2σ error (lighter).

term (420 – 413 ka) decrease of $136 \pm 16\%$ in DSW strength (Fig. 3F) and $65 \pm 10\%$ UNADW strength (Fig. 2B), which culminated in the overall minimum ($-32 \pm 22\%$ compared to modern) in combined AMOC_{SS} strength during the latter ISOW event at 413 ka (Fig. 2C).

Another clear difference in the climate state between the two ISOW weakening events is the spatial extent of surface ocean cooling. An abrupt, large ($7 \pm 1.5^{\circ}\text{C}$), and persistent cooling of the north-western SPG (Labrador Sea; Fig 1B), to temperatures comparable to those seen in the late Holocene, occurred at 417 ka. Based on the high % *Nps* values, this was likely associated with increased sea-ice cover; this 417 ka cooling was spatially restricted to the western SPG region, as southern-SPG and central/eastern subpolar North Atlantic sites maintained low abundance of the polar foraminifera *Nps* at this time. In contrast, cooling extended across the whole of the subpolar North Atlantic from 414 – 412 ka (*i.e.*, during the second interval of ISOW weakening).

Consequently, the broader AMOC weakening interval from 417 ka to 412 ka can be subdivided into: (i) an initial stage (417 – 415 ka) involving a transient weakening of ISOW and cooling and

expansion of sea-ice extent limited to the Labrador Sea region; and (ii) a second stage (414 – 412 ka) characterized by widespread cooling of the entire subpolar North Atlantic linked to a further, more severe weakening of AMOC caused by the combined effect of a now much weaker DSOW and UNADW, and a transient ISOW weakening.

Weakening of AMOC during enhanced GrIS melt

Our records not only reveal a decline in both reconstructions of AMOC strength at 414–412 ka (Fig. 2 and 2D) but can also be directly aligned with proxy records of southern GrIS melt and retreat (Fig. 2F) using co-occurring turbidite layers at Site U1305 (see SI section 1.5). Previous work has used pollen records from Atlantic marine cores to indicate GrIS deglaciation through the rapid development of spruce forest in southern Greenland at this time (28). This has been supported by silt provenance data from similar archives to reconstruct shifts in erosion patterns and, consequently, to infer a retreat of the southern GrIS. Stratigraphic alignment of the respective cores enables us to show that the reconstructed period for AMOC weakening coincided with this inferred melting of the Southern GrIS, and the subsequent rapid expansion of woodland in the region.

The maximum modelled rate of GrIS meltwater flux during MIS 11c was 8–10 mSv (Fig. 2F), which is at the lower end of the estimated flux required to impact AMOC, according to modelling studies investigating AMOC weakening (21–24). Furthermore, the GrIS flux estimates from MIS 11c are substantially lower than estimates of present-day GrIS freshwater flux (40 mSv) (57). Previous work has suggested that a meltwater flux of 5 mSv is sufficient to impact Labrador Sea convection if it persisted over a sustained period (23), and present day meltwater fluxes are exceeding this and have been argued to be impacting Labrador Sea convection (58). Furthermore, recent modelling work supports a significant role for the GrIS in AMOC weakening in the future, with up to 22% weakening to occur over 200 years under a high-melt scenario (59). Under that scenario, this does not happen abruptly nor to the extent reconstructed during MIS 11c; however, our study reconstructs a GrIS melt period substantially longer than the model run in question. What is less certain is the impact of GrIS melt on DSOW in the future as this is not as routinely investigated by ocean models. This is critical as, during MIS 11c, the weakening of DSOW appears to have been a key contributor to the occurrence of the severe AMOC weakening at 413 ka.

Despite model experiments suggesting that there would have been continued, prolonged melting of the GrIS, our data show that AMOC did not remain in a weakened state for the rest of the interglacial, but instead the weak AMOC event persisted for ~1000 years before recovering to pre-event levels (and stronger for the Nordic Overflows; Fig 2B)). This AMOC recovery is consistent with model results from abrupt CO₂ forcing scenarios where AMOC recovers on multi-centennial to millennial timescales (60) as well as the transient nature of the Holocene 8.2 kilo-year event (47).

Are we poised to experience an AMOC event similar to during MIS 11c? Through a comparison of the downcore data to modern (core-top) conditions, our data reveal that conditions today generally resemble those that occurred prior to the weak AMOC event at 413 ka, namely: (i) ISOW has weakened throughout the mid-late Holocene (50) and it is at a similar strength now as at ~414 ka (Fig. 3A); (ii) the relative abundance of polar planktic foraminifera in the northwest SPG (Labrador Sea) region (and the broader basin-wide spatial pattern of upper ocean temperatures), today, is comparable to that recorded after the SPG shift at 417 ka and prior to the AMOC weakening event at 413 ka (SI Fig. 18); (iii); UNADW strength is at a similar level now as at ~414 ka (Fig. 2B); and (iv) the present-day GrIS surface mass-balance anomaly is comparable to

modelled meltwater flux rates during MIS 11c, and are increasing (20). One key difference between conditions prior to the abrupt mid-MIS 11c AMOC event and today is that, to our knowledge, there is no evidence reporting a significant long-term (i.e. beyond inter-annual to decadal) weakening of DSOW, which our data suggests was a key component of the MIS 11c AMOC weakening.

Climatic impacts of AMOC weakening during MIS 11c

In modelling studies, the projected AMOC weakening of 12-54% by 2100 (61) is associated with global climate consequences (5,6). Beyond its effect on North Atlantic conditions, simulations predict that an AMOC slowdown would significantly impact temperatures and precipitation patterns across Europe and West Africa (6). These shifts would trigger substantial socioeconomic impacts, including changes in wildfire magnitude and frequency, as well as agricultural productivity (6). By combining our two independent measures of AMOC (Fig. 4B), we have reconstructed a 30-45% AMOC weakening event during MIS 11c, comparable in magnitude to the AMOC reduction in model projections. We therefore might expect to observe detectable global climate impacts of our MIS 11c AMOC weakening event in appropriate palaeoclimate archives.

A review of palaeoenvironmental archives that span MIS 11c indicate that evidence for climatic instability coincident with our evidence for AMOC weakening is widespread across a range of locations but is only most clearly seen at locations proximal to the North Atlantic. This includes evidence for cooling and aridification across northwestern Europe (Fig. 4C and 4D; see also (32)), with increased continentality towards Eastern Europe; and reductions in SST from marine cores of the Atlantic margins of southwest Europe and west Africa (Figure 4E and 4F). Much of the evidence from western, central and eastern Europe comes from pollen records, often in annually laminated, or *varved*, sediments. These archives allow the duration of vegetation response in these settings to be quantified and, typically, these show ecological shifts that persist for a few hundred years, in contrast to the millennial-scale weakening of AMOC recorded here. We attribute this discrepancy to the fact that ecosystem do not respond in a simple linear manner to climate change and may show little change until a threshold is crossed. This suggestion is supported by the observations that: 1) where direct tephra correlation between lake and marine archives is possible the ecological turnover occurs at ca 413ka, the period of most extreme AMOC weakening, and 2) where varved records contain independent climate proxies, i.e. lake carbonate $\delta^{18}\text{O}$, these show millennial-scale episodes of climate instability with the vegetation response occurring in response to the most prolonged and extreme isotopic shift (31). More equivocal evidence for mid-interglacial climatic instability is present in a range of locations ranging from southern Europe to Lake Baikal to eastern China (see SI Fig. 17). Development of well dated paleoclimate records across this interval to better constrain the magnitude and spatial extent of the climate impact of this AMOC weakening is a clear target for future efforts.

A notable feature of the MIS 11c greenhouse gas record is an abrupt short-lived increase in atmospheric CO_2 concentrations (a carbon dioxide jump, or CDJ, here CDJ-11c) (Fig. 4G); previous work has found these CDJ- events often coincide with AMOC weakening (30). The same study suggested that CDJ-11c and subsurface ocean cooling at Site 983 may have been concurrent, with temporal offsets attributed to age model uncertainties. As our age model is on the same timescale as the CDJ event, we are able to conclude that this is not the case. Although we do identify widespread evidence for AMOC weakening at 413 ka, including the cooling at Site 983, it did not coincide with CDJ-11c. Instead, CDJ-11c occurred at ~ 417 ka, at the onset of cooling

in the northwestern subpolar gyre. The link between AMOC changes and CDJ-11c therefore remains unclear.

MIS 11c AMOC weakening: a lesson for the future?

AMOC is a crucial component of the climate system, but its response to ongoing and projected GrIS melting is uncertain. Our study has shown that a $32 \pm 11\%$ mid-interglacial AMOC weakening occurred during MIS 11c, with associated shifts in the climate system across Europe, the North Atlantic and beyond. Moreover, our data reveal that this AMOC weakening event occurred during the most substantial wastage of the GrIS of the past 500,000 years. We suggest that the sustained GrIS melt contributed directly to the AMOC weakening, an interpretation supported by our reconstruction of individual components of the AMOC and the potential impact of GrIS melt upon them. There is wide debate about whether, and under what conditions, ongoing GrIS melting may cause AMOC weakening. Our study indicates that AMOC may respond to persistent GrIS wastage even when melt rates are relatively low (≤ 0.01 Sv).

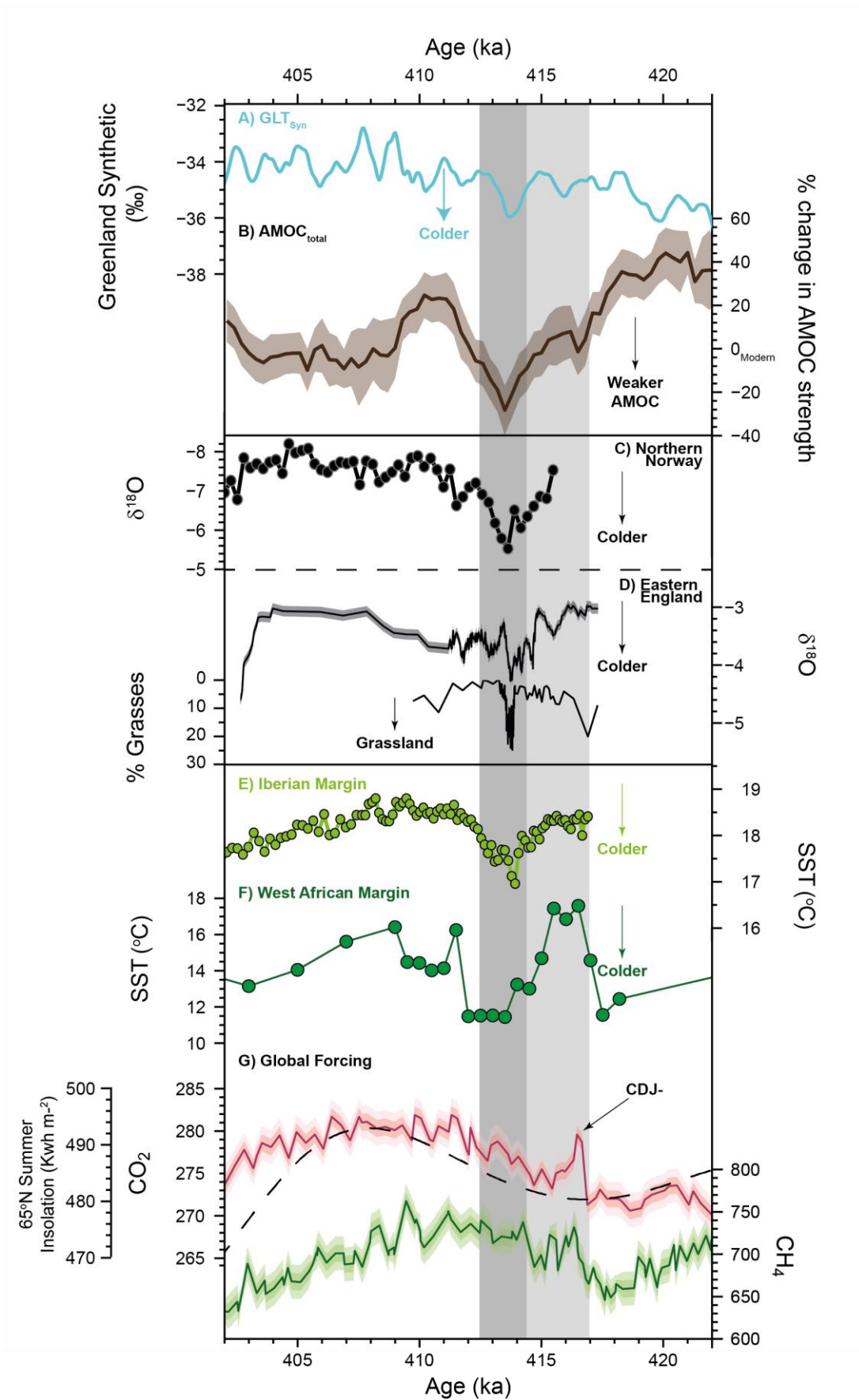


Fig. 4. Wider comparison of key MIS 11c records that show evidence for mid-interglacial instability associated with AMOC weakening: (A): The Synthetic Greenland $d^{18}O$ (‰) record used to generate the age models in this study (67) shown for context; (B): our best-estimate of MIS 11c AMOC strength ($AMOC_{Total}$ – brown solid line) compiled by averaging $AMOC_{SS}$ and $AMOC_{SPG-T}$; (C): $d^{18}O$ (‰) Speleothem record from Søylegrotta Cave, Norway (68). Note that the record does not have robust age model and is instead presented on depth, but has been aligned to mid-MIS 11c as the original study states that there is a hiatus in the early MIS 11c section, and absolute dating places the record mid-late MIS 11c; (D): $d^{18}O$ (‰) varve and grass pollen (%) records from Marks Tey, Eastern England (31). Note that the varved (~12 – 16m) section of this record has a robust tephra-based chronology (36); however, the rest of the record does not. As a result, this has been presented on depth and broadly aligned using the tephra age at 15.47m (36) on the AICC23 timescale. (E) Alkenone (U^{k}_{37}) derived Sea Surface Temperature record of MIS 11c from Site MD01-2443 (37.9°N, 10.2°W; water depth 2925m), Iberian margin, on the age model of this study (51). (F) Faunal (Modern Analogue Technique) near surface ocean temperature record of MIS 11c from ODP Site 958 (24.0°N, 20.0°W; water depth 3728m), North-west African margin, on the age model of this study (69). (G) Global climate forcings including Carbon dioxide (red), methane (green) (30) and boreal summer insolation in kilowatt-hours per square metre ($Kwh\ m^{-2}$) (dashed black) (62). CDJ = Carbon Dioxide Jump; Light grey shading indicates broad AMOC weakening; dark grey shading indicates abrupt AMOC weakening. All records, unless otherwise stated, are presented on the AICC23 timescale (70).

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Main text figure captions:

Fig. 1. Core locations and data synthesis compared to key climate forcings: (Top) IODP Site U1302 (50.2°N, 45.6°W; 3250m depth) (this study); IODP Site U1305 (57.5°N, 48.5°W; 3463m depth) (this study, (38); ODP Site 983 (60.4°N, 23.6°W; 1983m depth) (this study, (35); ODP Site 984 (61.0°N, 24.0°W; 1650m depth) (this study); IODP Site U1304 (53.1°N, 33.5°W; 3024m depth) (this study); ODP Site 980 (55.5°N, 14.7°W; 2170m depth) (this study), (62); IODP Site 1055 (32.47°N, 76.2°W; 1800m depth) (this study); projected on a map of North Atlantic Ocean

Temperatures ($^{\circ}\text{C}$, 0 – 200m depth) using data from WOA23 plotted in Ocean Data View. Study core sites are depicted as white circles with black outlines. Key comparison sites in this study are depicted by black circles. Key surface (solid) and intermediate/deep (dashed) ocean currents are depicted. **(A, C)** Key climate forcings (duplicated above each dataset) including greenhouse gas data (CO_2 and CH_4) from EPICA Dome C (30) and summer insolation in kilowatt-hours per square metre (kWh m^{-2}) at July 65°N (63). **(B)** % *Neogloboquadrina pachyderma* (*Nps*) records compared with compiled normalised IRD from study sites. **(D)** Mean sortable silt in microns (μm) data compared with a compiled normalised Ice Rafted Debris record from study sites. Coloured bars: pale brown at onset and end of MIS 11c = ice rafting intervals; light grey = broad mid-interglacial cooling; dark grey = abrupt mid-interglacial cooling. DSOW = Denmark Strait Overflow Water; ISOW = Iceland Scotland Overflow Water; LSW = Labrador Sea Water; NADW = North Atlantic Deep Water; NAC = North Atlantic Current; EGC = East Greenland Current; WGC = West Greenland Current; LC = Labrador Current; SPG = Sub-Polar Gyre; UNADW = Upper North Atlantic Deep Water; IRD = Ice Rafted Debris.

Fig.2. Physical and Index-based reconstructions of AMOC compared to potential forcing mechanisms during the main interglacial period of MIS 11c: **(A):** Lower North Atlantic Deep Water (blue solid line) reconstructed from grain size records of Iceland Scotland Overflow Water (ISOW – purple dashed line) (ODP Sites 983 and 984) and Denmark Strait Overflow Water (DSOW – green dotted line) (IODP Sites U1302 and U1305). **(B):** Upper North Atlantic Deep Water (maroon solid line) reconstructed from the grain size record of ODP Site 1055. **(C):** Compiled Physical (grain size) based record of AMOC across the main warm interval of MIS 11c (AMOC_{SS} – black solid line) assuming equal contribution of each AMOC component; **(D):** Compiled Index (ocean temperature) based record of AMOC across the main warm interval of MIS 11c ($\text{AMOC}_{\text{SPG-T}}$ – red solid line); Records **A-D** presented with reference to modern (0) with modern indicated by a black dashed line; **(E):** Proxy records (Benthic $\delta^{13}\text{C}$ and % CaCO_3) of changes in the Deep Western Boundary Current (DWBC) from ODP Site 1063 (46,64). % CaCO_3 was calculated using equations from (65); **(F)** Greenland ice sheet (GrIS) dynamics derived from CaCO_3 -free silt flux (grey dotted polygon) (27) and pollen concentrations (green polygon) (28). Modelled GrIS meltwater flux was derived from (66) (grey solid line) and has been temporally adjusted within age model uncertainties so that the highest rates of meltwater flux coincide with the smallest sGrIS extent. Note this has been inverted. Coloured vertical bars: light grey = broad mid-interglacial cooling; dark grey = abrupt mid-interglacial cooling. Shading around compiled records indicate 1σ (darker) and 2σ error (lighter).

Fig. 3. Records of Nordic Seas overflows compared to key forcings across MIS 11c: (67). **(A-D)** Compiled Iceland Scotland Overflow Water (ISOW) record compared to forcing mechanisms. **(A and E):** Boreal summer insolation was obtained from (63) in kilowatt-hours per square metre (kWh m^{-2}) at July 65°N . **(B):** Compiled ISOW record (purple) derived from ODP Sites 983 and 984 **(C):** Oxygen isotopes derived from planktic foraminifera *Neogloboquadrina pachyderma* (black) and *Turborotalita quinqueloba* (grey) (43). Highlighted in red are consistent (multi-datapoint) offsets of $\geq 0.4\text{‰}$, indicating a thermal regime favouring increased convection and stronger ISOW. **(D):** Nitrate utilisation ($\delta^{15}\text{N}$) indicates reduced utilisation during the mid-late interglacial, indicating increased mixed-layer depth, favouring convection (51); **(E-H)** Compiled Denmark Strait Overflow Water (DSOW) record compared to forcing mechanisms; **(F):** Compiled DSOW record from IODP Sites U1302 and U1305 (green); **(G-H):** Lead isotope data indicates the earliest **(G)** and latest **(H)** timing of the Laurentide Ice Sheet retreating to the late Holocene extent

(Canadian Archipelago) on the age model of this study (late Holocene levels = dashed line). The larger Laurentide Ice Sheet coincides with strongest DSOW. Panels B and F are presented with reference to modern (0), indicated by a dashed black line for ISOW DSOW. Coloured bars: light blue = broad mid-interglacial AMOC weakening; dark blue = abrupt mid-interglacial AMOC weakening. Shading around compiled records indicate 1σ (darker) and 2σ error (lighter).

Fig. 4. Wider comparison of key MIS 11c records that show evidence for mid-interglacial instability associated with AMOC weakening: (A): The Synthetic Greenland $\delta^{18}\text{O}$ (‰) record used to generate the age models in this study (68) shown for context; (B): our best-estimate of MIS 11c AMOC strength ($\text{AMOC}_{\text{Total}}$ – brown solid line) compiled by averaging AMOC_{SS} and $\text{AMOC}_{\text{SPG-T}}$; (C): $\delta^{18}\text{O}$ (‰) Speleothem record from Søylegrotta Cave, Norway (69). Note that the record does not have robust age model and is instead presented on depth, but has been aligned to mid-MIS 11c as the original study states that there is a hiatus in the early MIS 11c section, and absolute dating places the record mid-late MIS 11c; (D): $\delta^{18}\text{O}$ (‰) varve and grass pollen (%) records from Marks Tey, Eastern England (31). Note that the varved (~12 – 16m) section of this record has a robust tephra-based chronology (36); however, the rest of the record does not. As a result, this has been presented on depth and broadly aligned using the tephra age at 15.47m (36) on the AICC23 timescale. (E) Alkenone (U^{k}_{37}) derived Sea Surface Temperature record of MIS 11c from Site MD01-2443 (37.9°N, 10.2°W; water depth 2925m), Iberian margin, on the age model of this study (52). (F) Faunal (Modern Analogue Technique) near surface ocean temperature record of MIS 11c from ODP Site 958 (24.0°N, 20.0°W; water depth 3728m), North-west African margin, on the age model of this study (70). (G) Global climate forcings including Carbon dioxide (red), methane (green) (30) and boreal summer insolation in kilowatt-hours per square metre (Kwh m^{-2}) (dashed black) (63). CDJ = Carbon Dioxide Jump; Light grey shading indicates broad AMOC weakening; dark grey shading indicates abrupt AMOC weakening. All records, unless otherwise stated, are presented on the AICC23 timescale (71).