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Freshwater influx, hydrographic reorganisation and the dispersal of Ice Rafted Detritus in the sub-polar North Atlantic Ocean during the last deglaciation.

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ABSTRACT

A sediment core from the northeast North Atlantic contains high-resolution co-registered foraminiferal $\delta^{18}O$ and IRD record for the last deglaciation. These reveal a distinct ice-rafting event that occurred at the time of Greenland Interstade 1d (GI-1d), a feature also seen in other high-resolution cores from the North Atlantic. The occurrence of a geographically widespread peak in ice rafted detritus (IRD) at ice distal sites at a time when increased freshwater flux to the surface ocean is inferred to have caused rapid cooling suggests a mechanistic link between the processes, analogous to the Younger Dryas (GS-1) cooling episode. The general absence of IRD at southern locations at other times during GI-1 when the flux of icebergs from surviving ice sheets to northern locations continued, suggests that the GI-1d IRD peak represents a time of hyrdrographic reorganisation which changed IRD dispersal. While numerous studies have suggested freshwater flux as a major driver of rapid climate oscillations observed around the North Atlantic during the last deglaciation, the evidence presented here both supports that mechanism and highlights the potential for rapid and major

reorganisation of the of the North Atlantic's surface hydrography to explain changes in IRD flux independently of ice sheet calving dynamics.

Key Words

Ice rafted detritus, North Atlantic, deglaciation, sea surface temperature, hydrography.

1. Introduction

High-resolution records spanning the Last Glacial-Interglacial Transition (LGIT) have consistently revealed a climate system punctuated by numerous abrupt climate transitions (Severinghaus and Brook, 1999; Alley *et al.*, 2003; Steffensen *et al.*, 2008). Several of these events have been linked to changes in Atlantic Meridional Overturning Circulation (AMOC) and its associated heat flux because of its sensitivity to increased freshwater input (Clark *et al.*, 2001; Clark *et al.*, 2002a; McManus *et al.*, 2004). To firmly establish freshwater forcing as an underlying causal mechanism of these abrupt climate transitions requires firstly, well constrained chronostratigraphies such that events can be correlated between records with confidence and, secondly, widespread geological evidence that links the North Atlantic's surface conditions to changes in the deep and intermediate ocean.

The occurrence of Ice Rafted Detritus (IRD) within marine sediments has long been used to investigate the links between climate, oceans and ice sheets (Heinrich, 1988; Bond *et al.*, 1993; Elliot *et al.*, 2001; Knutz *et al.,* 2001; Hemming, 2004; Peck *et al.*, 2007; Hendy and Cosma, 2008; Scourse *et al.*, 2009). In order to fully understand these links, it is not only important to understand how variations in IRD relate to ice sheet advance and/or retreat (McCabe and Clark, 1998; Marshall and Koutnik, 2006) but also to develop an understanding of the way hydrographic controls can influence the dispersal of IRD within an oceanic basin. There has been relatively little work relating hydrographic factors to IRD records, although some authors have argued for minimal hydrographic control in parts of the sub-polar North Atlantic (Elliot *et al.*, 2001) and in relation to the first occurences of Heinrich like events in the earlier Pleistocene (Naafs *et al.,* 2011). One situation where hydrographic controls have been cited as influencing the pattern of IRD deposition is in regards to the location of the IRD Belt (Figure 1) (Ruddiman*,* 1977; Ruddiman and McIntyre, 1981; Scourse *et al.*, 2009).

Potential hydrographic controls include the pattern of oceanic surface currents which affect the dispersal of icebergs and thus IRD and, potentially more importantly, sea surface temperature (SST). Palaeoclimatic records indicate that large and abrupt changes to AMOC during the last glacial period were associated with changes in climate (Vidal *et al.*, 1997; Austin and Kroon, 2001; Clark *et al.*, 2002a; Rahmstorf, 2002; McManus *et al.*, 2004; Thornalley *et al.,* 2010). Freshwater input to the North Atlantic is suggested to be one of the major drivers of changes to AMOC and attendant climatic impacts (Broecker, 1994). This link is supported by proxy data (Elliot *et al.*, 2002; McManus *et al.*, 2004) and modeling studies (LeGrande *et al.*, 2006; Clarke *et al.*, 2009; Liu *et al.*, 2009; He *et al.,* 2013), which demonstrate that a weakening of AMOC is associated with cooling and lower SSTs. SSTs influence the melt rates of icebergs and hence their longevity in the open ocean; as such, icebergs are more likely to travel large distances across ocean basins during times of lower SSTs. IRD peaks in ice distal sites at these times may reflect the increased persistence and dispersal of icebergs, relating to hydrographic conditions, rather than an increased total flux of icebergs related to ice sheet dynamics. Comparing IRD records from distal sites with records relating to temporal variations in iceberg flux (both proximal and distal) can allow these links to be investigated.

High resolution records spanning the LGIT are punctuated by numerous IRD events which some authors have linked directly to changes in ice sheet dynamics (eg: Knutz *et* al, 2001), given the potential influence of SST variations highlighted above it is imperative to establish if making such inferences is valid. Such an understanding will aid attempts to integrate records of oceanic changes with ice sheet behaviour and explore the two-way forcing relationship that exists (Clark *et al*, 2001). Here we present a new, high resolution IRD record from the northeast North Atlantic and use it to identify a distinct and widespread LGIT ice rafting event. It's timing, as indicated by a major change in the co-registered foraminiferal

$\delta^{18}O$, coincides with the cold interval of GI-1d (Lowe *et al.*, 2008). Episodic freshwater input to the North Atlantic has been proposed as the cause of such cold intervals that punctuate the LGIT (Thornalley *et al.*, 2010). Establishing the effects of such input has important implications for our understanding of rapid climate change during the last deglaciation as it provides a linking mechanism between ice sheets and changes in oceanic circulation.

2. Study site and methods.

2.1. Study site

Giant Piston Core MD95-2007 was collected in 1995 from the *RV Marion Dufresne* in the St Kilda basin on the Hebridean shelf, Northwest Scotland (57° 31.057' N, 08° 23.171' W, 158 m water depth, 19.35m recovery; Figure 1). The St Kilda basin is a glacially over-deepened basin that is within the limits of the last British-Irish Ice Sheet (BIIS) (Davies *et al.*, 1984; Peacock *et al.*, 1992; Stoker *et al.*, 1993). This conclusion is supported by a ^{14}C date of 22,480±300 ^{14}C a BP (27.1 cal ka BP [OxCal v4.1 (Bronk-Ramsey, 2009), MARINE09 (Reimer *et al.,* 2009)]) on marine shell material to the west of morainal banks marking the aforementioned BIIS limit and of 'basal' ages <16 ^{14}C ka BP (<19 cal ka BP) within those same limits (Peacock *et al.*, 1992; Austin and Kroon, 1996). These ages demonstrate that this sector of the BIIS was at its maximum extent prior to 20 ka. As MD95-2007 is located within these ice limits it is thought to record nearly the entire deglacial sequence following initial deglaciation of the shelf edge.

The potential for cores recovered from the St Kilda basin to record high-resolution records of the LGIT was initially demonstrated from two vibrocores VE57/-09/89 and VE57/-09/46 (Austin*,* 1991*;* Peacock *et al.*, 1992; Austin and Kroon, 1996). These cores showed an expanded LGIT sedimentary sequence but a poorly resolved Holocene sequence. From the variety of sedimentary, micropalaeontological and isotopic evidence recorded in these cores it was proposed that the St Kilda basin deglaciated at 15.2 ^{14}C ka BP (17.6 cal ka BP) after which its waters remained cold with low salinity until 13.5 ^{14}C ka BP (15.6 cal ka BP). Following this time, mostly warm interstadial conditions prevailed until a major cooling associated with the onset of GS-1 was observed at 11.6 ^{14}C ka BP (13.0 cal ka BP). The return to warm temperatures at the beginning of the Holocene occurred prior to 10 ^{14}C ka BP (11 cal ka BP) (Austin and Kroon, 1996).

2.2. A revised chronostratigraphy for MD95-2007

The original age model for MD95-2007 (*Wilson*, 2004) was based on 16 AMS ^{14}C ages calibrated using Calib4_2 (Stuiver and Reimer, 1993; Stuiver *et al.*, 1998), following a reservoir correction ($R_{(t)}$) reflecting the modern values of seawater (i.e. ΔR=0). The availability of a $\delta^{18}O$ record ($\delta^{18}O_{foram}$), measured in the epi-benthic foraminifera *Cibicides lobatulus* (originally reported in Austin *et al*., 2011), provides an additional means of improving the age-depth relationship. Given the rapid and abrupt nature of the GI-1 climate oscillations, the high resolution of the MD95-2007 $\delta^{18}O_{foram}$ record and inherent uncertainty about the variable marine reservoir effect during this period (Austin *et al.*, 1995; 2011), it is important to determine the timing of particular climatic episodes during GI-1 *vis à vis* the candidate cold episodes GI-1d or GI-1b (Figure 3). This is done by constraining the $\delta^{18}O_{foram}$ record using recalibrated AMC ^{14}C ages (OxCal v4.1, MARINE09 (Bronk-Ramsey, 2009; Reimer *et al.,* 2009) with three different values for $R_{(t)}$: the modern value 400 years (ΔR = 0), the commonly cited GS-1 value 700 years (ΔR = 300) and a maximum value of 1,100 years (ΔR = 700) (Waelbroeck *et al.*, 2001). This approach provides a good first order constraint to the age of the interstadial $\delta^{18}O_{foram}$ excursion (Austin *et al.*, 2011). In order for this excursion to correlate with GI-1b (13.3-13.1 ka b2k) the reservoir age correction would need to exceed 1000 years (Figure 3). A reconstruction of $R_{(t)}$ at this time from MD95-2007 indicates that it was lower than the GS-1 value of ~1000 years (Austin *et al.*, 2011) thus the $\delta^{18}O_{foram}$ excursion is correlated with GI-1d (Figure 3).

Based upon this interpretation of the climate event-stratigraphy, the $\delta^{18}O_{foram}$ stratigraphy can be tuned to the NGRIP $\delta^{18}O_{ice}$ record using the GICC05 chronology (Rasmussen *et al.*, 2006). For the rapid $\delta^{18}O_{foram}$ changes at the onset and end of GS-1 and GI-1d the tie-point was assigned to the mid-point of the transition. The Vedde Ash has been indentified within MD95-2007 and more widely across the St Kilda basin (Austin *et al.*, 1995; Peters *et al.*, 2010; Austin *et al.*, 2011). This tephra occurs at a core depth of 281 cm and has been assigned an age of 12,171 a b2k (Rasmussen *et al.*, 2006). Table 2 summarizes the tie-points, their core depths and the ages assigned to them. During this interval the age uncertainty within the GICC05 timescale is 100-200 years (Rasmussen *et al.*, 2006) however as we are comparing tuned records this has no influence on our conclusions. This age model follows the INTIMATE protocols (Lowe *et al.*, 2008; Austin and Hibbert, 2012). It must be noted that this approach assumes synchroneity and therefore any information about time leads/lags is lost and conclusions based on the results must respect this limitation.

One complicating factor in the use of the tuning method to construct an age model for MD95-2007 is that there is no obvious structure to the local climate event-stratigraphy beyond the cold excursion GI-1d. As a result age control in the lower section of the core is difficult, further dating may improve this but would be hampered by a general scarcity of suitable material in the lower core (Austin pers. Comm). To anchor the base of the record it is therefore necessary to use the basal radiocarbon determination (13,950±130 ^{14}C a BP) at a depth of 1821 cm. This subsequently introduces an uncertainty related to the choice of $R_{(t)}$ used in the correction and then calibration of this age. The differences in the basal age calculated using the various corrections are significant and would affect any interpretations based on the timing of events in the lower 0.6 m of our record. For this reason we avoid making interpretations based on data from the un-tuned section of the age model which must be considered tentative. The effects of a variable correction are, however, restricted to this basal section of the core and are not a key factor when considering the tuned section to which this study relates.

2.3. A new IRD record from MD95-2007

IRD_{flux} is calculated from the IRD concentration and bulk mass accumulation rate (BMAR), such that:

$$IRD_{flux} = IRD_{conc} * BMAR$$

The BMAR in turn is calculated using a linear sedimentation rate (LSR) derived from the age model and the dry bulk density (ρDB) of the sediment such that:

$$BMAR = LSR * \rho DB$$

Calculation of the LSR involves linear interpolation between the tie points used in construction of the core's age model. ρDB is calculated using the wet and dry mass of known volumes of sediment assuming sediment particle and pore water densities of 2650 and 1025 kg m^{-3} and pore water salinity of 35 g kg^{-1}.

Lithic counts were carried out on the coarse (>250 μm) fraction. Traditionally, grains coarser than 150 μm are considered to be ice rafted (Hemming, 2004), but we use a coarser fraction because of the possibility that the shelf was a higher energy environment compared to the deep ocean, especially at times, such as the LGIT, when sea level was lower.

3. Results

The IRD_{flux} record from core MD95-2007, plotted against the benthic $\delta^{18}O_{foram}$ record (Figure 4), shows 3 periods of increased IRD flux to the core site. The initial, and highest, period of increased flux occurs near the base of the core. IRD_{flux} during this time consistently exceeds 150,000 grains cm^{-2} a^{-1}. This period corresponds to the missing part of the $\delta^{18}O_{foram}$ stratigraphy such that age-control is poor and resultant IRD flux uncertainty relatively high. Following this there is a period of near zero IRD_{flux} which lasted until ~14.1 ka. The subsequent peak in IRD corresponds to a significant $\delta^{18}O_{foram}$ excursion correlated to GI-1d (see section 2.2). After this brief (121±4 a; NGRIP/GICC05 timescale) period IRD_{flux} returns to the very low background levels observed prior to GI-1d. This low rate continues until a slow increase in IRD_{flux} prior to the onset of GS-1 that is marked by a distinct increase in the IRD_{flux} to the core site. Following the end of GS-1, as defined in the $\delta^{18}O_{foram}$ stratigraphy, IRD_{flux} rates are zero.

4. Discussion

The period of increased IRD_{flux} centered on 14.1 ka coincides with a $\delta^{18}O_{foram}$ excursion that has been correlated with the cold oscillation GI-1d observed within the NGRIP $\delta^{18}O_{ice}$ record (Rasmussen *et al.*, 2006). Recent provenance work using U-Pb dating of detrital minerals identifies a distinct distal component to the IRD found within MD95-2007, inferred to be sourced from north eastern Canada, Baffin Island or East Greenland (Small *et al.*, 2013). The provenance data presented by Small *et al.* (2013) do not preclude a contribution from local sources (i.e. the BIIS) and the abundance of coarse (≥250 μm) is suggestive of a local source given the relationship between IRD grain size and transport distance (Andrews, 2000). However, it is doubtful that the BIIS had marine margins capable of supplying IRD to the offshore environment at this time (Bradwell *et al*., 2008; Ballantyne and Stone, 2012). It is possible that grain-specific provenance studies may be biased particularly if the analysed grains come from a particular size fraction. Despite this potential limitation the distinct distal signal identified within the IRD provenance data provides clear evidence that IRD was transported some distance across the North Atlantic during the short cold oscillation GI-1d.

The IRD_{flux} and benthic $\delta^{18}O_{foram}$ records from MD95-2007 are shown in comparison to proxy records from several other North Atlantic cores (DAPC-2, RAPid-15-4P, TTR-451; Figure 5). The highlighted peak in IRD_{flux} observed at 14.1 ka, which is co-registered with a $\delta^{18}O_{foram}$ excursion within MD95-2007, also coincides with distinct increases in the IRD records in both DAPC-2 and RAPid-15-4P indicating some common mechanism is responsible. However, it should be pointed out that the IRD record from RAPid-15-4P is concentration rather than flux however the consistent pattern between this and the flux records (MD-95-2007, DAPC-2) gives us confidence that it records the same event.

The IRD peak in RAPid-15-4P corresponds to the occurrence of an Icelandic tephra interpreted as being the Katla Ash. This tephra is also observed within the NGRIP record and can thus be assigned a precise age of 14.02 ka b2k (Thornalley *et al.*, 2010; Rasmussen *et al*, 2006). The age model of DAPC-2 was constructed by visual tuning of the Nps% record with the GISP2 $\delta^{18}O$ record (Knutz *et al.,* 2007) which results in an age offset compared to the cores tuned to the NGRIP record (Figure 10 in Rasmussen *et al.,* 2006). The availability of AMS radiocarbon dates within this core is not sufficient to independently verify the tuning in this part of the stratigraphy. However, the stratigraphic position of the IRD peak suggests that it is correlative to GI-1d as it occurs at the same time as the earliest recorded peak in Nps% during GI-1. This strongly indicates that the IRD peak represents the same event that is seen in MD95-2007 and RAPid-15-4P.

Considering the evidence for a distal contribution of IRD during GI-1d (Small *et al.,* 2013) it can be hypothesised that, *if* the overall flux of icebergs to the North Atlantic from these distal sources is the fundamental control on the occurrence of IRD within MD95-2007, then the MD95-2007 IRD_{flux} record should reflect variations in this flux. A similar relationship should exist during GI-1 for DAPC-2, which is inferred to have been predominantly supplied with BIIS sourced IRD for most of its history (Knutz *et al.,* 2007). Obtaining comparable provenance data from DAPC-2 would allow this assumption to be tested.

The IRD peaks during GI-1 seen within DAPC-2 and RAPid-15-4P occur during times of low SST indicated by relative high abundance of the planktonic foraminifera *N. pachyderma* sinistral (Nps%) (Knutz *et al.*, 2007; Thornalley *et al.*, 2010). The correlation between IRD peaks and peaks in Nps% indicates that they occurred during times when the sites were located north of the Polar Front (Scourse *et al.,* 2009). If the observed GI-1d IRD peaks simply reflected an increased flux of IRD to the North Atlantic, then it would be expected that they would record IRD at other times when it is recorded in ice proximal core sites. Core TTR-451 from the Eirik Drift (Stanford *et al.*, 2011) shows increased IRD flux throughout GI-1. It can be inferred from this that significant amounts of icebergs were calved from the Greenland Ice Sheet (GIS) at times when no IRD was reaching the distal core sites. Furthermore, models suggest that the GIS would have maintained calving margins for a large part of GI-1, providing a persistent possible source for icebergs (Simpson *et al.*, 2009). This implies that some other control was acting to prevent deposition at the ice distal sites at times of continuing iceberg flux to higher latitudes.

A series of rapid variations in SST are observed in the Nps% records from cores DAPC2 and RAPid-15-4P (Knutz *et al.*, 2007; Thornalley *et al.*, 2010). In each core one of these variations is correlated with GI-1d and is coincident with the GI-1 IRD peak. Nps% peaks indicating cooler periods of SST during GI-1 are also seen in core MD95-2006, taken from the Barra fan, <100 km south west of MD95-2007 (Wilson *et al.*, 2002; Peters *et al.*, 2010; Hibbert *et al.,* 2010). It is reasonable to assume that one of the peaks corresponds to GI-1d given the widespread and simultaneous nature of climate change in the North Atlantic at this time (Bjorck *et al.*, 1996; Broecker, 2000; Rohling *et al.*, 2003). An additional record from the Barra Fan based on the planktonic foraminiferal assemblages in core VE56/-10/36 also shows a variation in SST around the time of GI-1d (Kroon *et al.*, 1997). The occurrence of these brief periods of lower SST would favour the southerly penetration of icebergs calved from the surviving North Atlantic ice sheets because of the fundamental control SST has on the survival of icebergs in the ocean (Dowdeswell and Murray, 1990). Given the absence of IRD at times of higher SST's when there was a continuing flux of icebergs from the same potential sources, it is likely that it is SST which was the fundamental control on IRD dispersal to the sub-polar North Atlantic during GI-1. The occurrence of a peak in IRD flux associated with GI-1d is inferred to be the result of the attendant reduction in SST's associated with this climatic oscillation.

SST's in the North Atlantic depend strongly on AMOC, with a weaker AMOC associated with lower SST's at higher latitudes (Schmittner *et al.*, 2005; Barker *et al.*, 2009). Rerouting events have been identified that correspond to the abrupt climate reversals of the last deglaciation (Clark *et al.*, 2001) and one such release of freshwater is proposed as the cause of GI-1d and its concomitant cooling that is seen across the North Atlantic (Rasmussen *et al.*, 2006; Thornalley *et al.*, 2010). This freshwater input, likely caused an AMOC slowdown with an associated decrease in SST's, visible in the palaeorecords and sufficient to allow icebergs (and IRD) to reach MD95-2007 and other ice distal sites.

The LGIT was punctuated by periods of increased meltwater input from the decaying ice sheets (Fairbanks, 1989; Hanebuth *et al.*, 2000; Bard *et al.*, 2010). These meltwater pulses had various sources but their effects, both in terms of sea level rise and climate, were widespread (Stanford *et al.*, 2006; Stanford *et al.*, 2010). The largest of the identified meltwater pulses is MWP-1a, whose initial contributor is thought to be the Antarctic Ice Sheet (AIS), where partial collapse released freshwater into the Southern Ocean (Clark *et al.*, 2002a). Resumption in NADW formation forced by a Southern injection of meltwater and resulting in warming in the Northern Hemisphere (Weaver *et al.*, 2003) would explain the Bølling warming that marks the end of the LGM and the start of GI-1 (Lowe *et al.*, 2008). This warming would have favoured melting of the Northern Hemisphere ice sheets which would have made a subsequent contribution to MWP-1a (Carlson *et al.,* 2012), in turn diminishing the vigour of AMOC. This interaction produces a feedback between ice sheets and climate (Clark *et al.,* 2001; McManus *et al.*, 2004; Meissner and Clark, 2006; Clarke *et al.*, 2009; Thornalley *et al.*, 2010; He *et al.,* 2013). The evidence presented in this study demonstrates a rapid alteration to the North Atlantics surface hydrography during the last deglaciation, in agreement with this proposed feedback mechanism.

5. Conclusions and Implications

The co-registered, high-resolution IRD_{flux} and $\delta^{18}O_{foram}$ records from MD95-2007 provide evidence of an IRD peak during GI-1 that is coincident with a period of lower $\delta^{18}O_{foram}$ values. The timing of this event is constrained using ^{14}C dates and tuning of the record to NGRIP and is correlated with the short-lived cooling episode GI-1d (14,075-13,954 a b2k). Given our knowledge of the distribution of the pan-North Atlantic ice sheets and IRD provenance fingerprinting using U-Pb dating of detrital minerals (Small *et al*, 2013) it has been established that the IRD peak in MD95-2007 reflects input of distally sourced material. The absence of IRD in southerly (ice distal) cores at times when its flux at a northerly (ice proximal) core was continuing suggests that IRD flux to the wider North Atlantic from the surviving ice sheets is not the primary control. To explain this pattern of IRD occurrence it is necessary to invoke a hydrographic control that prevents IRD deposition at ice distal sites except during a defined cold interval, namely lowered SST.

The release of freshwater into the North Atlantic is proposed as the driving mechanism of the short-lived and abrupt climate variations such as GI-1d. The effects of freshwater input to the North Atlantic are primarily manifested through a slowdown of AMOC that reduces SST's. A reduction in SST's favours the survival of icebergs and their wider dispersal. As such we suggest that it is by this mechanism that IRD was deposited within these sub-polar cores during GI-1d. The widespread effects of meltwater input to the North Atlantic during the LGIT are clearly documented, but our evidence is some of the first that does not rely on planktonic foraminiferal $\delta^{18}O$ alone.

Our conclusion that the GI-1 IRD peak within MD95-2007 contains distal material (Small *et al.,* 2013) and that hydrographic conditions are important controls on sub-polar IRD dispersal has far reaching implications. IRD has regularly been used to infer fine scale behaviour of individual ice sheets. For example, IRD deposited during the last glacial cycle in the sub-polar North Atlantic records sub-Milankovitch (millennial) scale climatic changes that have been linked to the abrupt calving dynamics of marine ice-sheet margins (eg: Knutz *et al.*, 2001; Scourse *et al.*, 2009; Hibbert *et al.*, 2010). By combining geographically distinct IRD records, information regarding IRD flux to the wider ocean, and IRD provenance studies it is possible to demonstrate that hydrography may be an important additional control on IRD occurrence. Our results highlight, particularly at the millennial and sub-millennial scale, that

IRD flux records may reflect a complex interplay between changes to oceanic conditions and ice sheet calving dynamics.

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Figure Caption

Figure 1. Map of the North Atlantic showing the location of MD95-2007 (red star) and other cores mentioned in the text. Also shown is the IRD belt of Ruddiman, (1977) and the approximate location of the polar front (dashed red line) at various times in the past after McManus *et al.,* (1994).

Figure 2. Final age model for MD95-2007. Tuning of the MD95-2007 benthic $\delta^{18}O_{foram}$ record (Austin *et al*., 2011) to the NGRIP $\delta^{18}O$ record on the GICC05 timescale (Rasmussen *et al.,* 2006). a) MD95-2007 benthic $\delta^{18}O$ against core depth; b) NGRIP $\delta^{18}O$; c) MD95-2007 benthic $\delta^{18}O$ tuned using basal AMS ^{14}C date corrected for $\Delta R = 300$. The solid lines show the tie points based on major climate transitions visble in both records, the dashed line indicates the basal radiocarbon age. Black dots on top axis are available ^{14}C dates. The Vedde Ash isochron is labelled.

Figure 3. MD95-2007 $\delta^{18}O_{foram}$ plotted using 3 'preliminary' age models constrained by the available AMS ^{14}C dates (Table 1) calibrated using OxCal v4.1 and MARINE09 (Bronk Ramsey, 2009; Reimer *et al.,* 2009) and three different values for ΔR; 0, 300 and 700. The tuned age model is shown for comparison.

Figure 4. MD95-2007 IRD_{flux} record plotted against $\delta^{18}O_{foram}$ record using the tuned age model (Figure 2). The stratigraphic divisions are as recommended by INTIMATE (Lowe *et al.,* 2008).

Figure 5. Composite stratigraphic plot of the IRD records discussed in the text plotted on their original timescales. The tuned proxies from MD95-2007 (this study; Austin *et al*., 2011), DAPC-2 (Knutz *et al*., 2007) and RAPid-15-4P (Thornalley *et al*., 2010) are shown alongside the IRD records. The NGRIP $\delta^{18}O$ record (Rasmussen *et al*., 2006) is shown alongside the

IRD record from TTR-451 (Stanford *et al*., 2011). The age offset between DAPC-2 and the other records is the result of this record being tied to GISP2 $\delta^{18}O$ (Groots and Stuiver, 1997); the other records were tied to NGRIP $\delta^{18}O$ on the GICC05 timescale (Rasmussen *et al*., 2006).

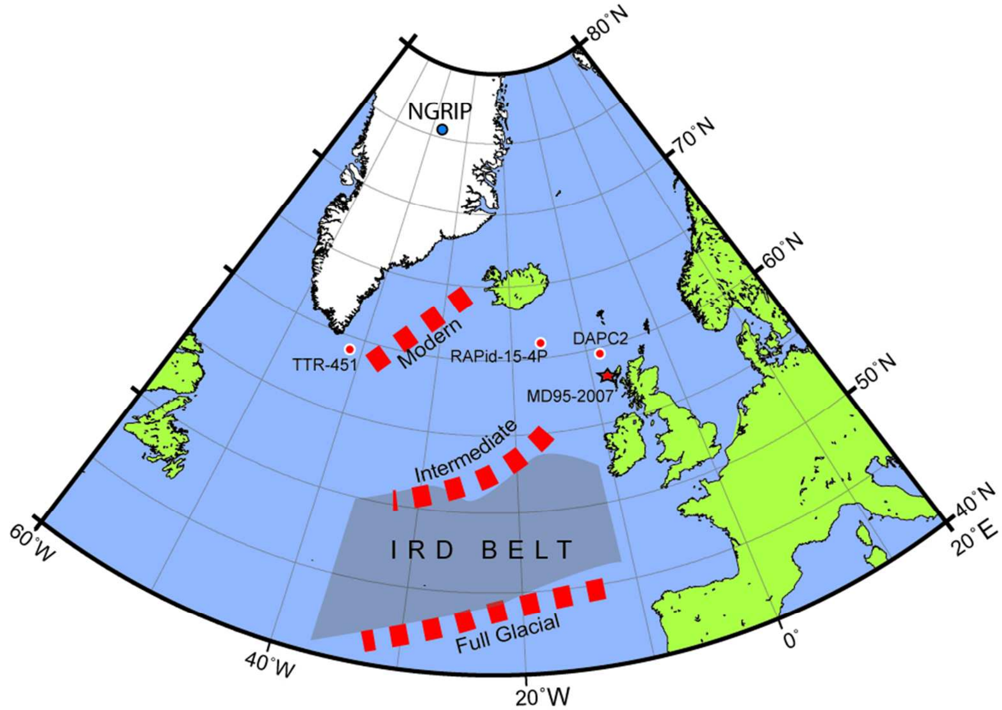
Table 1. ^{14}C Ages from MD95-2007

Sample	Core depth cm)	Radiocarbon age (^{14}C a BP ±1sigma)	Calibrated Age (cal a BP) ΔR=0	ΔR=300	ΔR=700	Original age*
AA-41753	21	2279±36	2285±62	1537±56	1140±51	1879
AA-41754	121	10664±65	12601±61	11394±143	10926±121	11825**
AA-41762	375.5	11353±62	13231±65	12583±81	11987±137	12907
AA-41755	396.5	11299±66	13195±72	12512±88	11873±158	12890
AAR-2602	425	11500±90	13355±97	12689±96	12249±164	13000
AA-41763	442.5	11296±77	13192±85	12494±98	11861±183	12889
AA-41756	556.5	11471±62	13331±67	12657±67	12202±128	12990
AA-41757	741.5	12353±74	14396±247	13502±101	13152±98	13832
AAR-2603	826	12630±100	14880±275	13777±127	13379±111	14109
AAR-2604	974.5	12790±120	15235±352	13949±194	13545±129	14289
AA-41758	1008.5	12789±88	15211±271	13926±121	13542±108	14289
AA-41759	1345.5	12953±74	15526±318	14133±229	13695±105	14721**
AAR-2605	1663	13810±170	16925±198	15890±416	15028±457	15995
AA-41760	1674	13020±110	15721±366	14356±280	13765±137	14739**
AAR-2606	1815.5	14250±150	17346±222	16664±279	15958±402	16502
AA-41761	1821	13950±130	17029±170	16152±385	15384±374	16157

Ages calibrated using OxCal v4.1, MARINE09. ΔR *=0, 300 & 700 (Bronk Ramsey, 2009; Reimer et al., 2009)*
**Ages from Wilson* (2004) *calibrated using Calib4_2* ΔR *=0, uncertainties not reported.*
*** Average of 3 ages from shell fragments*

Table 2. Tie points used in construction of tuned age model.

Tie Point	Core depth cm)	Age assigned (a b2k)	Reference
End of GS-1	101	11703	*Lowe et al.* 2008
Vedde Ash	281	12171	*Rasmussen et al.* 2006
Start of GS-1	521	12896	*Lowe et al.* 2008
End of GI-1d	941	13954	*Lowe et al.* 2008
Start of GI-1d	1016	14075	*Lowe et al.* 2008



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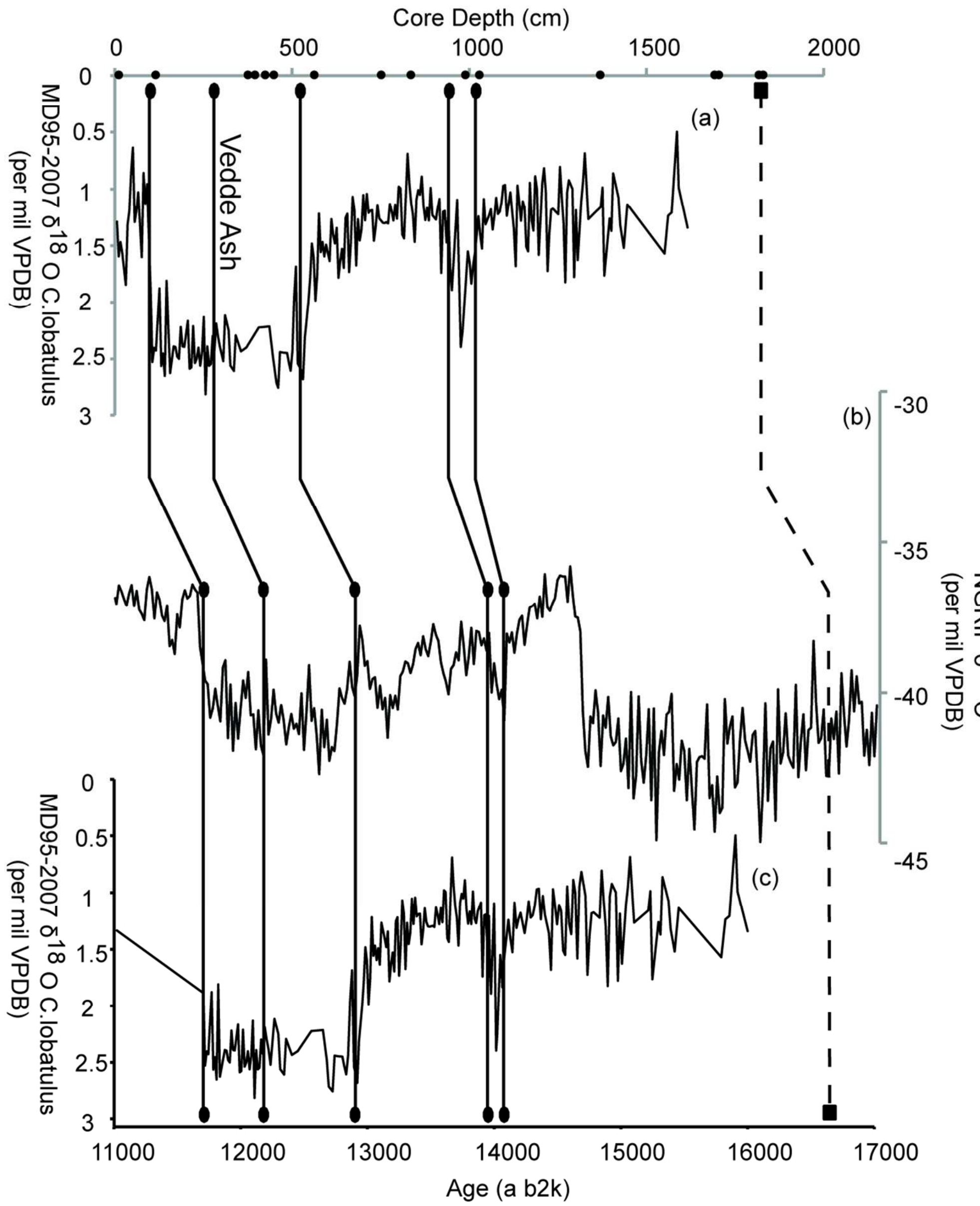

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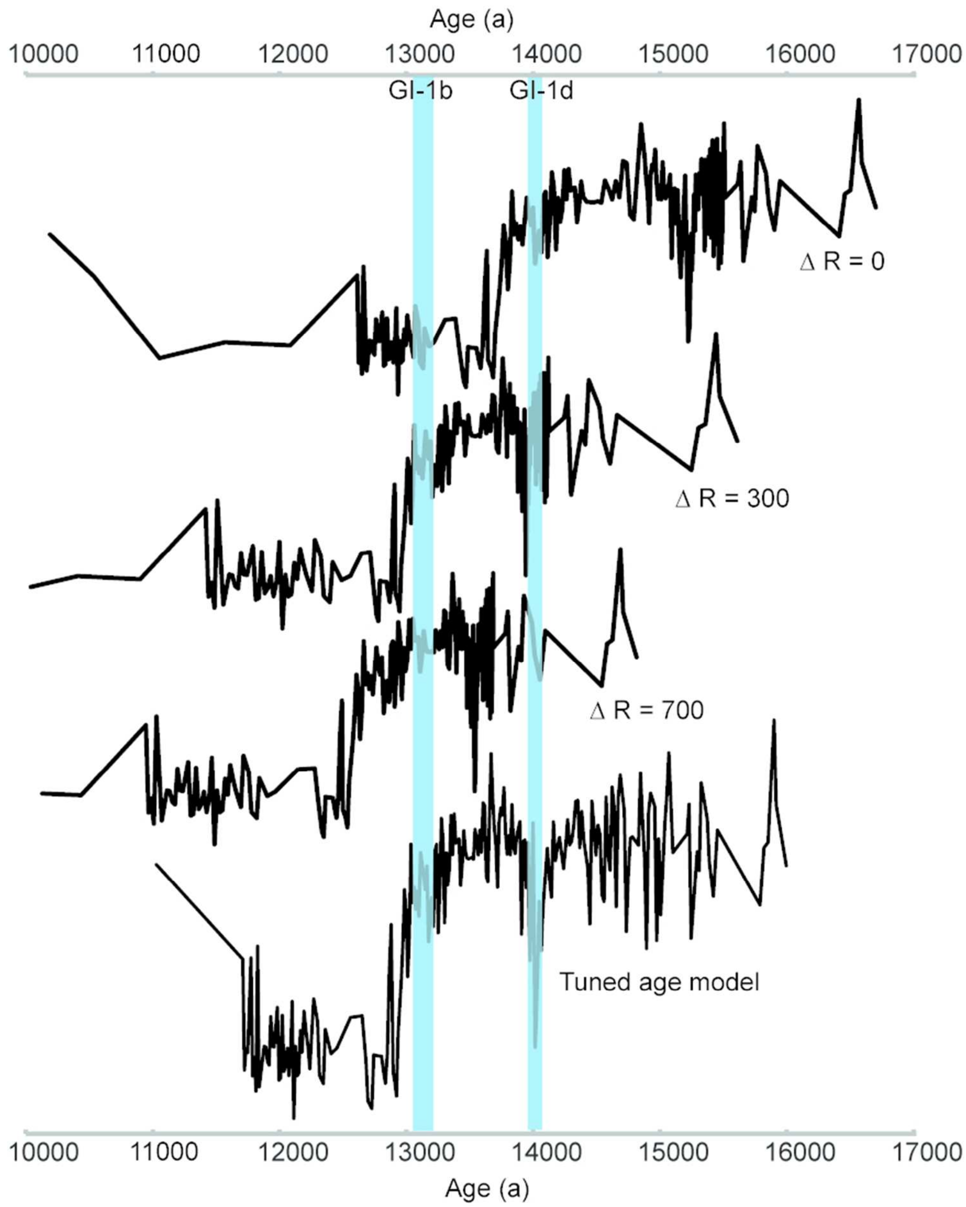

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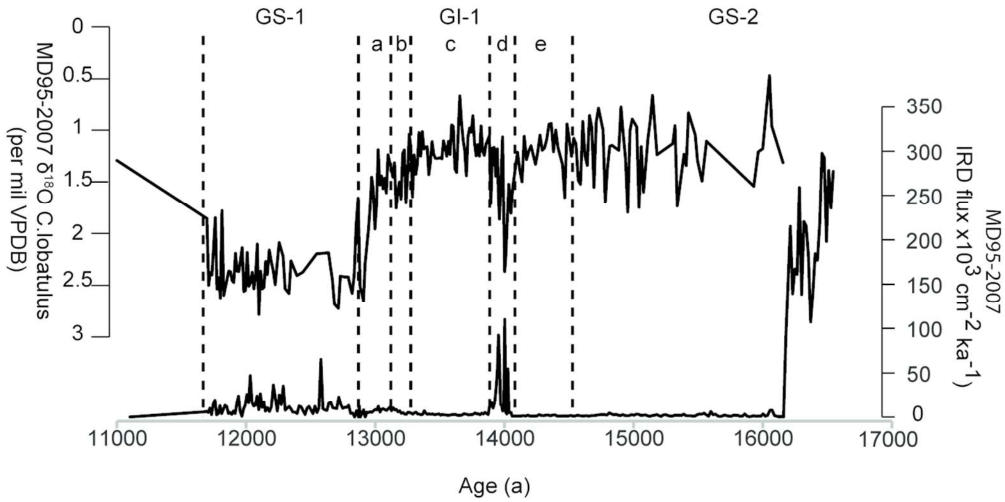

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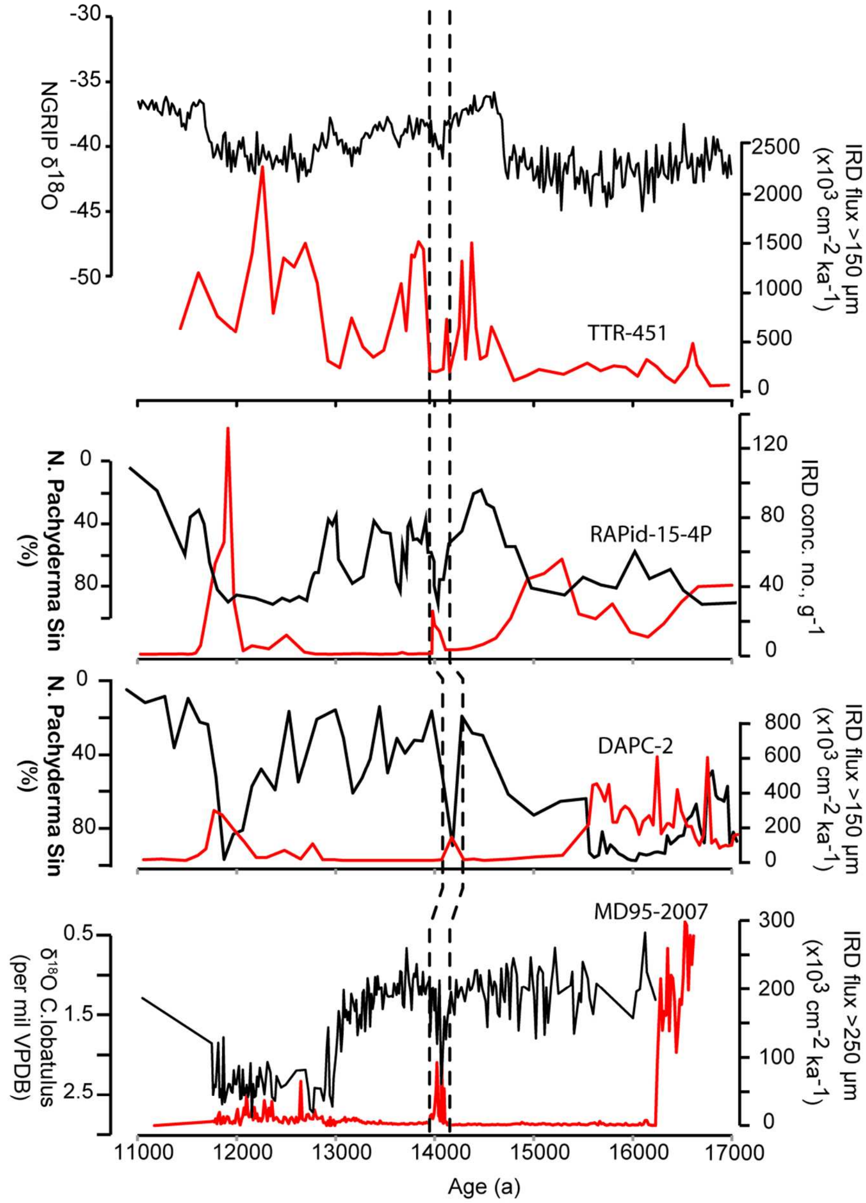

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