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

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1 Freshwater influx, hydrographic reorganisation and the dispersal of Ice
2 Rafted Detritus in the sub-polar North Atlantic Ocean during the last
3 deglaciation.
4
5 David Small\textsuperscript{1*}, William Austin\textsuperscript{1}, Vincent Rinterknecht\textsuperscript{1}
6 School of Geography and Geosciences, University of St Andrews, St Andrews, Scotland.
7 *Now at School of Geographical and Earth Sciences, University of Glasgow, Glasgow, Scotland.
8
9 Corresponding author D. Small, School of Geographical and Earth Sciences University of
10 Glasgow, Glasgow, Scotland G12 8QQ. Email: David.Small@glasgow.ac.uk
11
12 ABSTRACT
13 A sediment core from the northeast North Atlantic contains high-resolution co-registered
14 foraminiferal $\delta^{18}O$ and IRD record for the last deglaciation. These reveal a distinct ice-rafting
15 event that occurred at the time of Greenland Interstade 1d (GI-1d), a feature also seen in other
16 high-resolution cores from the North Atlantic. The occurrence of a geographically widespread
17 peak in ice rafted detritus (IRD) at ice distal sites at a time when increased freshwater flux to
18 the surface ocean is inferred to have caused rapid cooling suggests a mechanistic link between
19 the processes, analogous to the Younger Dryas (GS-1) cooling episode. The general absence
20 of IRD at southern locations at other times during GI-1 when the flux of icebergs from
21 surviving ice sheets to northern locations continued, suggests that the GI-1d IRD peak
22 represents a time of hydrographic reorganisation which changed IRD dispersal. While
23 numerous studies have suggested freshwater flux as a major driver of rapid climate
24 oscillations observed around the North Atlantic during the last deglaciation, the evidence
25 presented here both supports that mechanism and highlights the potential for rapid and major

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reorganisation 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 occurrences 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
δ¹⁸O, 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 ¹⁴C date of 22,480±300 ¹⁴C 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 ¹⁴C 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\(_0\)) reflecting the modern values of seawater (i.e. \(\Delta R=0\)). The
availability of a \(\delta^{18}\)O record (\(\delta^{18}\)O_{foram}), measured in the epi-benthic foraminifera Cibicidales
lobatus (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\(_0\); the modern value 400 years (\(\Delta R = 0\)),
the commonly cited GS-1 value 700 years (\(\Delta R = 300\)) and a maximum value of 1,100 years
(\(\Delta R = 700\)) (Waelbroeck et al., 2001). This approach provides a good first order constraint to
the age of the interstadial δ¹⁸O<sub>foram</sub> 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<sub>0</sub> 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 δ¹⁸O<sub>foram</sub>
excursion is correlated with GI-1d (Figure 3).

Based upon this interpretation of the climate event-stratigraphy, the δ¹⁸O<sub>foram</sub>
stratigraphy can be tuned to the NGRIP δ¹⁸O<sub>ice</sub> record using the GICC05 chronology
(Rasmussen et al., 2006). For the rapid δ¹⁸O<sub>foram</sub> 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 synchronicity 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 ¹⁴C a BP) at a

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

$$IRD_{\text{flux}} = IRD_{\text{conc}} \times BMAR$$

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

$$BMAR = LSR \times \rho_{DB}$$

Calculation of the LSR involves linear interpolation between the tie points used in construction of the core’s age model. $\rho_{DB}$ 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$_{\text{flux}}$ record from core MD95-2007, plotted against the benthic $\delta^{18}\text{O}_{\text{foram}}$ 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$_{\text{flux}}$ during this time consistently
exceeds 150,000 grains cm$^{-2}$ a$^{-1}$. This period corresponds to the missing part of the $\delta^{18}\text{O}_{\text{foram}}$
stratigraphy such that age-control is poor and resultant IRD flux uncertainty relatively high.

Following this there is a period of near zero IRD$_{\text{flux}}$ which lasted until ~14.1 ka. The
subsequent peak in IRD corresponds to a significant $\delta^{18}\text{O}_{\text{foram}}$ excursion correlated to GI-1d
(see section 2.2). After this brief (121±4 a; NGRIP/GICC05 timescale) period IRD$_{\text{flux}}$ returns
to the very low background levels observed prior to GI-1d. This low rate continues until a
slow increase in IRD$_{\text{flux}}$ prior to the onset of GS-1 that is marked by a distinct increase in the
IRD$_{\text{flux}}$ to the core site. Following the end of GS-1, as defined in the $\delta^{18}\text{O}_{\text{foram}}$ stratigraphy,
IRD$_{\text{flux}}$ rates are zero.

4. Discussion

The period of increased IRD$_{\text{flux}}$ centered on 14.1 ka coincides with a $\delta^{18}\text{O}_{\text{foram}}$
excursion that has been correlated with the cold oscillation GI-1d observed within the NGRIP
$\delta^{18}\text{O}_{\text{ice}}$ 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

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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\textsubscript{flux} and benthic δ\textsuperscript{18}O\textsubscript{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\textsubscript{flux} observed at 14.1 ka, which is co-registered with a δ\textsuperscript{18}O\textsubscript{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 δ\textsuperscript{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\textsubscript{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 \textit{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; Thoronalley 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; Thoronalley 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$_{\text{flux}}$ and $\delta^{18}$O$_{\text{foram}}$ records from MD95-2007
provide evidence of an IRD peak during GI-1 that is coincident with a period of lower
$\delta^{18}$O$_{\text{foram}}$ 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

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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 δ¹⁸O 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 visible 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 $\Delta R$; 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. ¹⁴C Ages from MD95-2007

<table>
<thead>
<tr>
<th>Sample</th>
<th>Core depth cm</th>
<th>Radiocarbon age (¹⁴C a BP ±1sigma)</th>
<th>ΔR=0</th>
<th>ΔR=300</th>
<th>ΔR=700</th>
<th>Original age*</th>
</tr>
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<tbody>
<tr>
<td>AA-41753</td>
<td>21</td>
<td>2279±36</td>
<td>2285±62</td>
<td>1537±56</td>
<td>1140±51</td>
<td>1879</td>
</tr>
<tr>
<td>AA-41754</td>
<td>121</td>
<td>10664±65</td>
<td>12601±61</td>
<td>11394±143</td>
<td>10926±121</td>
<td>11825**</td>
</tr>
<tr>
<td>AA-41762</td>
<td>375.5</td>
<td>11353±62</td>
<td>13231±65</td>
<td>12583±81</td>
<td>11987±137</td>
<td>12907</td>
</tr>
<tr>
<td>AA-41755</td>
<td>396.5</td>
<td>11299±66</td>
<td>13195±72</td>
<td>12512±88</td>
<td>11873±158</td>
<td>12890</td>
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<tr>
<td>AAR-2602</td>
<td>425</td>
<td>11500±90</td>
<td>13355±97</td>
<td>12689±96</td>
<td>12249±164</td>
<td>13000</td>
</tr>
<tr>
<td>AAR-2603</td>
<td>826</td>
<td>12630±100</td>
<td>14880±275</td>
<td>13777±127</td>
<td>13379±111</td>
<td>14109</td>
</tr>
<tr>
<td>AAR-2604</td>
<td>974.5</td>
<td>12790±120</td>
<td>15235±352</td>
<td>13949±194</td>
<td>13545±129</td>
<td>14289</td>
</tr>
<tr>
<td>AAR-2605</td>
<td>1008.5</td>
<td>12789±88</td>
<td>15211±271</td>
<td>13926±121</td>
<td>13542±108</td>
<td>14289</td>
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<tr>
<td>AAR-2606</td>
<td>1345.5</td>
<td>12953±74</td>
<td>15526±318</td>
<td>14133±229</td>
<td>13695±105</td>
<td>14721**</td>
</tr>
<tr>
<td>AAR-2607</td>
<td>1663</td>
<td>13810±170</td>
<td>16925±198</td>
<td>15890±416</td>
<td>15028±457</td>
<td>15995</td>
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<tr>
<td>AAR-2608</td>
<td>1674</td>
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<tr>
<td>AAR-2609</td>
<td>1815.5</td>
<td>14250±150</td>
<td>17346±222</td>
<td>16664±279</td>
<td>15958±402</td>
<td>16502</td>
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<tr>
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<td>1821</td>
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<td>17029±170</td>
<td>16152±385</td>
<td>15384±374</td>
<td>16157</td>
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</tbody>
</table>

*Ages calibrated using OxCal v4.1, MARINE09. ΔR = 0, 300 & 700 (Bronk Ramsey, 2009; Reimer et al., 2009)*

**Average of 3 ages from shell fragments**

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

<table>
<thead>
<tr>
<th>Tie Point</th>
<th>Core depth cm</th>
<th>Age assigned (a b2k)</th>
<th>Reference</th>
</tr>
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<tbody>
<tr>
<td>End of GS-1</td>
<td>101</td>
<td>11703</td>
<td>Lowe et al. 2008</td>
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<tr>
<td>Vedde Ash</td>
<td>281</td>
<td>12171</td>
<td>Rasmussen et al. 2006</td>
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<tr>
<td>Start of GS-1</td>
<td>521</td>
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<td>Lowe et al. 2008</td>
</tr>
<tr>
<td>End of GI-1d</td>
<td>941</td>
<td>13954</td>
<td>Lowe et al. 2008</td>
</tr>
<tr>
<td>Start of GI-1d</td>
<td>1016</td>
<td>14075</td>
<td>Lowe et al. 2008</td>
</tr>
</tbody>
</table>
73x92mm (300 x 300 DPI)