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RESEARCH ARTICLE

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Key Points:

- An H0 layer is studied from Southern Newfoundland
- Biomarkers and mineralogy indicate a Hudson Strait source
- The H0 event is linked to rapid ice retreat at the onset of the Holocene

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Heinrich 0 on the east Canadian margin: Source, distribution, and timing

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Abstract The last deglaciation was marked by intervals of rapid climatic fluctuations accompanied by glacial advances and retreats along the eastern margin of the Laurentide Ice Sheet. One of these climatic events, the Younger Dryas cold reversal, was accompanied by a detrital carbonate event referred to as “Heinrich event 0” (H0) that deposited ice-rafted debris and especially detrital carbonate-rich (DC) sediment layers in the western and southern Labrador Sea. The precise age, duration, source, and geographical distribution of the H0 DC event, however, are not entirely clear. A high-sedimentation rate sequence cored off southern Newfoundland yielded an age of ~11.5–11.3 calibrated kyr B.P. for this layer, thus pointing to its deposition directly following the Younger Dryas termination, likely associated with rapid ice retreat and warming at the onset of the Holocene. At the study site, the H0 layer contains increased concentrations of detrital carbonates (calcite/dolomite ratio ~2.5:1) and a mature biomarker composition, similar to that found in Heinrich Layers 1, 2, 4, and 5, which together suggest an origin from Hudson Strait. Grain size analysis indicates that the H0-sediments were transported to the study site mostly through shallow plumes along the Labrador Shelf, with some hyperpycnal dispersal, deeper, along the slope and rise. Our data thus point to a large meltwater release caused by retreat of the ice margin in the Hudson Strait area. The relatively short duration of the H0 layer, and its timing right after a major climatic transition, combined with its ubiquitous regional occurrence makes it valuable for correlating different sediment records from the western Labrador Sea.

1. Introduction

The North Atlantic marine sediment record of the last glacial period is characterized by the occurrence of sediment layers of (1) increased ice rafted debris (IRD), (2) enhanced detrital carbonate (DC) content, and (3) low abundances of planktonic foraminifera, dominated by polar species (*Neogloboquadrina pachyderma* s.) and have been designated “Heinrich layers” [Heinrich, 1988; Andrews and Tedesco, 1992; Broecker et al., 1992; Hemming, 2004; Naafs et al., 2013]. Although the precise mechanisms driving the Heinrich events are still a subject of research, they are believed to have mainly originated from instability of the Laurentide Ice Sheet (LIS), resulting in large-scale iceberg and glacial meltwater discharge [Hesse, 1992; Hesse and Khodabakhsh, 1998; Hesse et al., 2004] with a near-global impact on the climate system [Broecker, 1994; Hemming, 2004; Naafs et al., 2013]. Due to their large geographical extent, Heinrich layers in the North Atlantic also have an important stratigraphic value, as they can be used for regional correlation between different records during the (late) Quaternary [Rasmussen et al., 2003; Hodell et al., 2008; Channell et al., 2012; Naafs et al., 2013].

Deposition of Heinrich layers during the last glacial period occurred in discrete periods, originally labeled from the youngest Heinrich 1 (H1) to the oldest H6 [Bond et al., 1992; Broecker et al., 1992]. Other Heinrich events were defined later, e.g., H0 [Andrews et al., 1994, 1995; Hillaire-Marcel et al., 1994] and H5a [Rashid et al., 2003], originally named Detrital Carbonate layer 6 [Stoner et al., 1996]. Sediment layers of increased detrital carbonate content in the northwestern Labrador Sea named DC layers were found by Andrews and Tedesco [1992] to be coeval with the iceberg rafting Heinrich events 1 and 2, suggesting the eastern Laurentide Ice Sheet as source for the icebergs. Although the origin of these events has often been identified

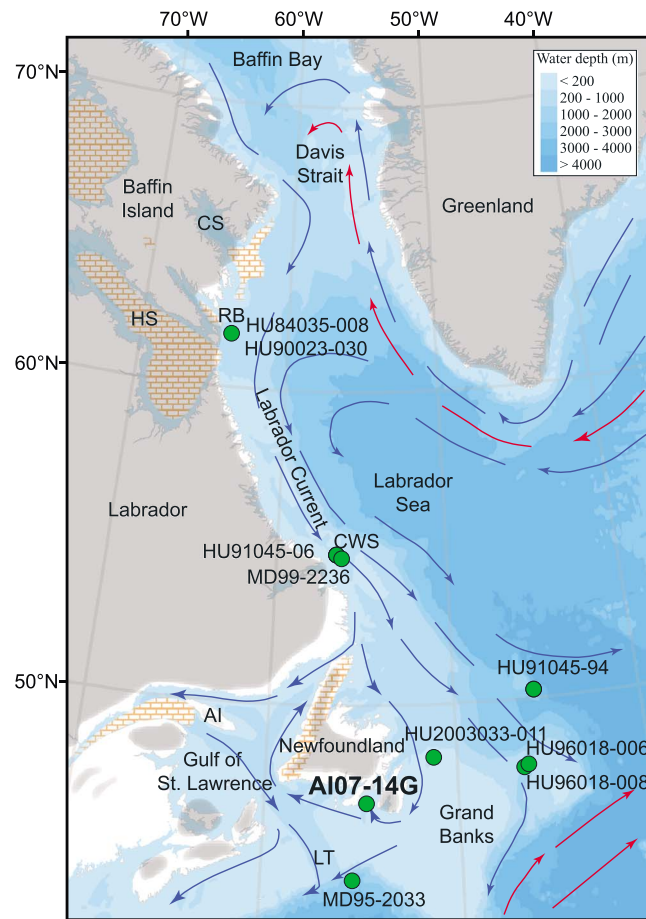


Figure 1. Map of Labrador Sea and surroundings with bathymetry; surface ocean currents (cold currents in blue, warm in red), redrawn from Pearce *et al.* [2014]; and distribution of carbonate bedrock (brick pattern) redrawn from Bond *et al.* [1992], Kirby [1998], and MacLean [2001]. All cores mentioned in the text are shown as green dots on the map. The gray-shaded areas show the extent of the Laurentide Ice Sheet at Younger Dryas age, the 10.5 ¹⁴C ka margins from Dyke *et al.* [2003]: AI, Anticosti Island; LT, Laurentian Trough; CWS, Cartwright Saddle; RB, Resolution Basin; HS, Hudson Strait; CS, Cumberland Sound. References for sediment cores are as follows: HU94035-008, HU90023-030 [Andrews *et al.*, 1995], HU91045-06, HU2003033-011 (as P19) [Hillaire-Marcel *et al.*, 2007], HU91045-94 [Hillaire-Marcel *et al.*, 1994], MD99-2236 [Jennings *et al.*, 2015], HU96018-006, HU96018-008 [Li and Piper, 2015], and MD95-2033 [Stevenson *et al.*, 2008].

et al., 1995] was based on sediment cores from Resolution Basin and uses the terms Heinrich 0 or H0 and DC-0 equivalently. In that study, sediments in core HU90023-030 (Figure 1) showed a dramatic rise in DC between (uncalibrated) radiocarbon dates of 13160 ± 115 ¹⁴C years B.P. (736 cm depth) and $11,065 \pm 105$ ¹⁴C years B.P. (441 cm) with a calcite/dolomite ratio during H0 of 2.5:1 compared to <1:1 for the ratio of coeval sediments sourced from Baffin Bay (Figure 1) [Andrews *et al.*, 1995, 2012].

Subsequently, sediment units associated with H0 have been identified in several marine sediment cores, which indicate that its sediment signature is generally restricted to the western and southern Labrador Sea [Hillaire-Marcel *et al.*, 2007; Stevenson *et al.*, 2008; Rashid *et al.*, 2011; Jennings *et al.*, 2015; Li and Piper, 2015]. However, in contrast to Andrews *et al.* [1995], where the H0 layer was found to be approximately time equivalent to the entire Younger Dryas cold period, the sediment records from farther south in the Labrador Sea seem to correspond to only the very end of the event [Hillaire-Marcel *et al.*, 2007; Stevenson *et al.*, 2008; Pearce *et al.*, 2013; Jennings *et al.*, 2015]. Based on reassessment of the marine reservoir offset (ΔR)

as the Hudson Strait region [Bond *et al.*, 1992; Hemming, 2004], there are other potential sources of carbonate along the eastern/northeastern flank of the Laurentide Ice Sheet, including extensive Paleozoic outcrops in the Gulf of St. Lawrence, on Newfoundland, and on the islands and channels at the head of Baffin Bay (Figure 1). Baffin Bay detrital carbonate layers have, however, been shown to be generally not correlative with North Atlantic Heinrich events [Andrews *et al.*, 2012]. The Baffin Bay detrital carbonates are characterized by a much higher dolomite content, i.e., a calcite-dolomite ratio of ~0.25:1 compared to $\geq 2.5:1$ for Hudson Strait sourced carbonates [Andrews *et al.*, 2012, 2013; Simon *et al.*, 2012].

The most recent of the Heinrich events, Heinrich 0 (H0), is characterized by a calcite-rich DC event in Resolution Basin, and the eastern limit of Hudson Strait [Andrews *et al.*, 1995], which is more or less synchronous to a dolomite-rich DC event recorded in sediments within Baffin Bay [Andrews *et al.*, 1996, 2013]. Heinrich 0 is generally associated with a northward ice readvance in Hudson Strait, which occurred during the Younger Dryas cold period [Andrews *et al.*, 1994, 1995; Hillaire-Marcel *et al.*, 1994; Kirby, 1998; Rashid *et al.*, 2011], although both Cumberland Sound [Kirby, 1998] and Baffin Bay [Andrews *et al.*, 2012] have been proposed as an alternative primary source for H0 sediments. The original definition of H0 [Andrews

and calibration of a radiocarbon date at the top of the carbonate layer in HU84035-008 (Figure 1) from the original H0 study [Andrews *et al.*, 1995], Jennings *et al.* [2015] determined the end of the H0 interval to be at $11,708 \pm 219$ calibrated (cal) years B.P., using the marine calibration curve Marine09 [Reimer *et al.*, 2009]. This range encompasses the age of the Younger Dryas (YD)-Holocene transition at $11,653 \pm 99$ cal years B.P. as defined from the NGRIP Greenland ice core based on the GICC05 time scale [Rasmussen *et al.*, 2006] and may indicate a link between the retreat of the H0 ice advance and the period of rapid climatic warming.

Recent studies of marine sediment core AI07-14G from Placentia Bay off southern Newfoundland (Figure 1) revealed a 40 cm thick layer of increased detrital carbonate content, characterized by an abrupt onset of deposition at ~ 11.5 cal kyr B.P., hence linked to H0 [Pearce *et al.*, 2013, 2014]. The hydrography of Placentia Bay is dominated by the coastal branch of the Labrador Current, which flows southeastward along the Labrador Shelf (Figure 1) and is a mixture of cold, low salinity waters from the Baffin Current and warmer, more saline waters of the West Greenland Current, as well as outflow from Hudson Strait [Petrie and Anderson, 1983; Lazier and Wright, 1993]. There is, however, also some input of Gulf Stream-derived waters coming into the bay from the south, the influence of which has been variable throughout the Holocene and the last deglaciation [Shaw *et al.*, 2013; Sheldon *et al.*, 2015]. Sea ice in Placentia Bay may be present from late winter-early spring, while icebergs carried by the Labrador Current from Baffin Island or West Greenland are rare [Catto *et al.*, 1999; Vale Inco, 2008].

To investigate the provenance of the detrital carbonate and further characterize the H0 layer from AI07-14G, sediments were analyzed using quantitative X-ray diffraction (qXRD) and organic geochemical methods, complementing previously published data of the sediment core. The aim of the present study is to investigate the H0 event in more detail, to identify its source and the mechanisms responsible for its deposition, and study its regional distribution on the Labrador Shelf and Grand Banks of Newfoundland. As the event is relatively well constrained in age from several sediment cores, we also explore its potential as a stratigraphical marker in the southwestern Labrador Sea.

2. Materials and Methods

2.1. Coring and Chronology

Gravity core AI07-14G (46.98878°N 54.70077°W ; Figure 1) was retrieved from 239 m water depth near the southward opening of Placentia Bay, Newfoundland, during a research cruise on R/V *Akademik Ioffe* in 2007. The core was cut into 1 m sections, split in half lengthwise and stored at 2°C , after which it was subsampled for further analyses. The chronology of AI07-14G is based on six radiocarbon dates, indicating an age span from ~ 10 to 13 calibrated kyr B.P.; a detailed description of the age-depth model is given in Pearce *et al.* [2014]. Radiocarbon measurements were calibrated using the Marine13 calibration curve [Reimer *et al.*, 2013] and a marine reservoir offset ΔR of 139 ± 61 years [Reimer and Reimer, 2001; Solignac *et al.*, 2011; Pearce *et al.*, 2014], using the age modeling software Oxcal v.4.2 [Ramsey, 2009]. The marine reservoir offset (ΔR) is defined as the contemporaneous ^{14}C offset from the marine calibration curve.

2.2. Sedimentological and Inorganic Geochemical Analyses

The core sections were photographed using an ITRAX core scanner at the Department of Geoscience of Aarhus University. Gray scale values were calculated at 5 mm resolution as the average of the RGB values retrieved from the core photograph. The elemental composition of the sediment core was analyzed through X-ray fluorescence (XRF) on the ITRAX scanner equipped with a Molybdenum anode tube [Croudace *et al.*, 2006]. Using this non-destructive method, semiquantitative multi-element measurements were made at high resolution, with the data obtained for calcium (Ca) and strontium (Sr) abundances presented here at 5 mm resolution. The grain size distribution was determined at ~ 5 cm resolution at the Department of Geoscience, Aarhus University, using the methods described in Nørnberg and Dalsgaard [2005]. The calcium carbonate (CaCO_3) content was measured at ~ 5 cm resolution using a Metrohm 855 Robotic Titrosampler by sodium hydroxide titration until a neutral pH was reached using 1 M HCl and 1 M NaOH, at the Department of Geography and Geology, University of Copenhagen. The above sedimentological data have previously been reported by Pearce *et al.* [2013, 2014]. Total organic carbon (TOC) measurements were also presented in Pearce *et al.* [2014], but for this study additional samples were measured at different sampling resolution (varying from 5 to 50 cm step size) focusing on the DC

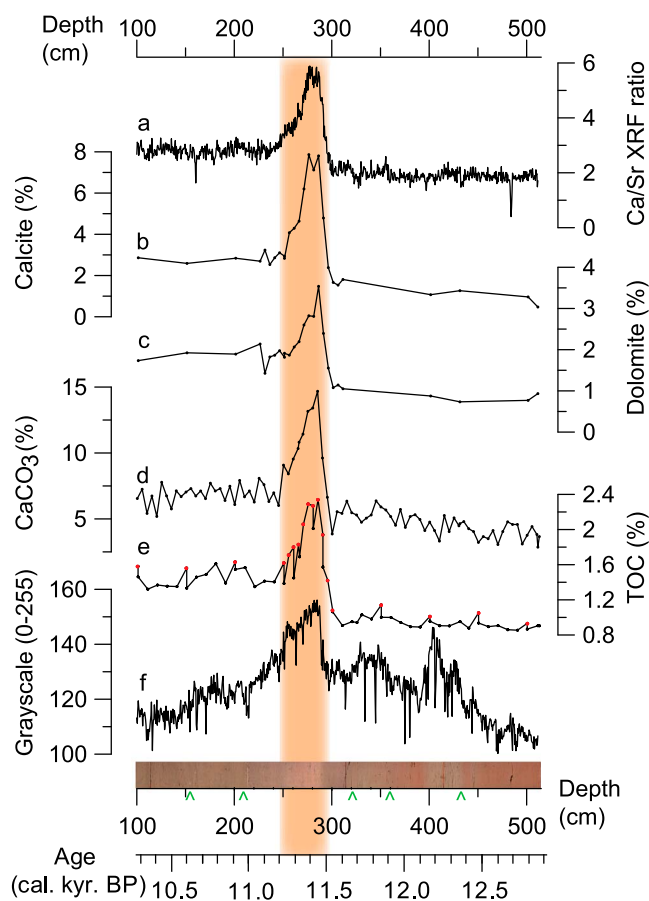


Figure 2. Proxy data of core AI07-14G: (a) XRF Ca/Sr ratios, (b) qXRD Calcite, (c) qXRD Dolomite, (d) CaCO₃, (e) TOC, (f) gray scale, and a core photograph. The variables are shown on both the depth and age scale; Figures 2a, 2d, 2e (black dots), and 2f were previously published in Pearce *et al.* [2014]. New TOC measurements from this study are shown in red (Figure 2e). Green markers indicate position of radiocarbon dates [Pearce *et al.*, 2014]. Shaded area indicates location of the detrital carbonate layer.

based on previously published analysis of biomarkers in Heinrich layers [Nicholl *et al.*, 2012; Naafs *et al.*, 2013]. Due to the extremely low concentrations of the studied biomarkers, large sediment samples were necessary for analysis, and therefore, the biomarker composition was only measured for a total of six samples (sample depths: 123, 203, 283, 288, 323, and 403 cm; 2 samples from within the DC layer and 2 on either side). Eleven to fifteen grams of freeze-dried sediments were extracted using dichloromethane/methanol (2:1) and ultrasonication to yield a total lipid extract. Aliphatic and aromatic (≤ 3 rings) hydrocarbons were isolated from the total extract on open column chromatography (SiO₂) using hexane and analyzed on an Agilent Technologies 7890 gas chromatograph coupled to an Agilent 262 Technologies 5975C inert XL mass selective detector equipped with a 30 m HP5MS column (0.25 mm i.d., 0.25 μ m film thickness). Helium was used as carrier gas. The oven temperature was programmed to rise from 40 to 300°C at 10°C/min and held at the final temperature for 10 min. Operating conditions for the mass spectrometer were 250°C for the ion source temperature and 70 eV for the ionization energy. Individual compound identification was based on interpretation of their mass spectra and comparison with published data [Peters and Moldowan, 1993]. Diagnostic ion fragments were used to monitor specific compounds: *m/z* 133 + 134 (isorenieratane and palaerenieratane), 231 (triaromatic steroids), 253 (C-ring monoaromatic steroids), 365 (D-ring monoaromatic 8,14-secohopanooids), 191 (17 α (H),21 β (H)hopanes, and 365 (D-ring monoaromatic 8,14-secohopanooids) [Rashid and Grosjean, 2006]. Results are shown in Figure 3. Due to the very low concentrations, absolute concentrations were not precisely determined, and values are semiquantitatively

interval. TOC content was determined using a continuous flow EuroVector elemental analyser coupled to an IsoPrime stable isotope ratio mass spectrometer at the Department of Physics and Astronomy of Aarhus University. Samples were pretreated with 1 M HCl at 60°C for 1 h and subsequently rinsed to neutral pH, freeze dried, and packed into tin capsules.

2.3. Mineralogical Analysis

The quantitative X-ray diffraction (qXRD) method used for this study to determine the calcite and dolomite concentrations is fully documented in Eberl [2003] and its application to Quaternary marine geology outlined in Andrews *et al.* [2010, 2012]. The method derives wt % estimates for the <2 mm sediment fraction. Replicate measurements of several known mixtures of nonclay and clay minerals indicate that the method error is in the range of $\pm 1\%$. Comparisons of calcite wt % made by qXRD and other methods resulted in highly significant correlations [Eberl, 2004; Andrews and Eberl, 2007].

2.4. Organic Geochemical Analysis

Hydrocarbon biomarkers were determined in selected samples across the DC layer. The specific choice of compounds that were analyzed was

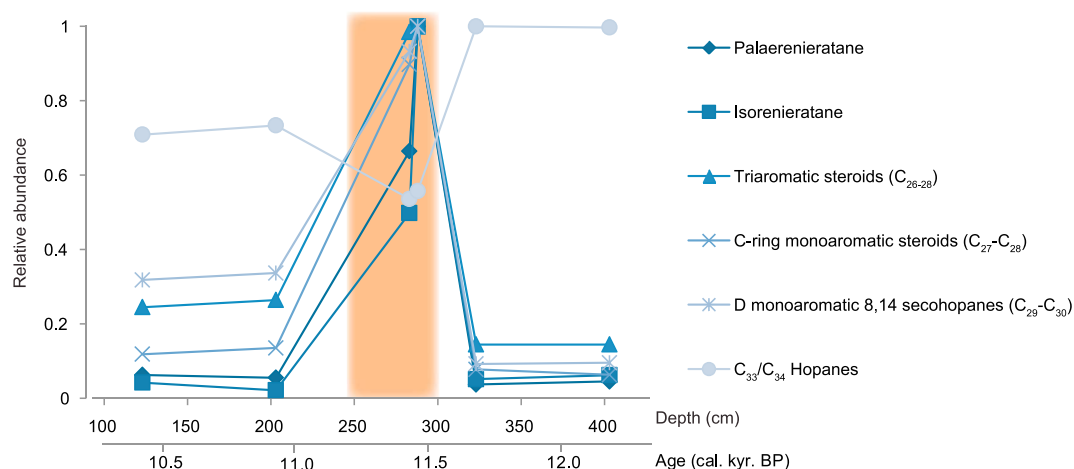


Figure 3. Biomarker composition of selected samples in core AI07-14G. The middle two samples fall within the detrital carbonate layer. All concentrations are normalized; i.e., they are shown as relative values based on their maximum occurrence.

presented as relative abundances with respect to their maximum concentration, which is sufficient for the purpose of this study, as it allows comparison of biomarker composition between different samples [Rashid and Grosjean, 2006; Naafs et al., 2011; Nicholl et al., 2012].

3. Results

The results from the XRF core scanning show a clear peak in the ratio of calcium to strontium counts (Ca/Sr) just above 300 cm depth in the core (Figure 2a), indicating an increased content of strontium-poor, detrital carbonate as opposed to strontium-rich biogenic carbonate [Hodell et al., 2008; Channell et al., 2012]. This ratio shows a very abrupt rise starting around 296 cm and a more gradual return to normal values around 245 cm depth. Although visual inspection of the core did not directly reveal the position of the detrital carbonate layer, the high-resolution photographic image and gray scale values clearly identify the layer as a lighter interval between approximately 240 and 290 cm depth (Figure 2f). Weight percentages of total organic carbon (TOC) range from 0.9 to 2.3% (Figure 2e). The lowest values, fluctuating around 1%, are found in the oldest part of the core, from the core bottom up to 300 cm depth. Beyond this point values increase rapidly to a peak value just over 2% at 280 cm, after which percentages drop back just as fast, but stay higher than before at around 1.6%. The added TOC data presented here are fully consistent with the data presented earlier in Pearce et al. [2014], although the concentrations are slightly higher than previously measured by about ~0.1% on average (Figure 2e). The calcium carbonate (CaCO₃) contents in the sediments (Figure 2d) show the same pattern as the TOC variability, although concentrations were higher and fluctuate between 2.5 and 14.5%. It should be noted here, however, that the applied method for CaCO₃ analysis might have also partly dissolved some dolomite, which may thus have contributed to the CaCO₃ percentages shown in Figure 2e. Again, here there are two long intervals with minimal variability, with the lowest values in the oldest section, separated by a short-lived maximum around 280 cm highlighting the detrital layer. Quantitative X-ray diffraction mineralogy analysis shows clear peaks in dry-weight percentages of both calcite (Figure 2b) and dolomite (Figure 2c) in samples from within the carbonate layer. Calcite concentrations lie between 1 and 3% and rise up to 7.9% at 276 cm depth, while dolomite concentrations show the same trend with a 1–2% background and 3.5% peak in the carbonate layer.

Biomarker analysis of six samples also indicates a clear difference in composition of the carbonate layer when compared to the background sediments (Figure 3). There is a predominance of C₃₄ over C₃₃ hopanes in the DC layer, and the concentrations of both aromatic hydrocarbons palaerenieratane and isorenieratane were measured to be approximately 16 times higher in the DC layer than in the embedding sediment. Other mature hydrocarbons showing the same trend include aromatic steroids and secohopanoids (Figure 3).

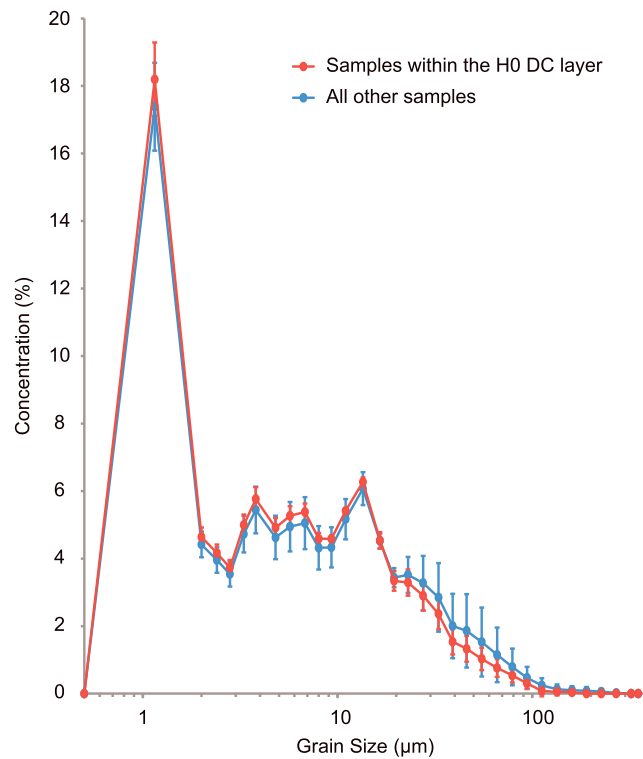


Figure 4. Grain size distribution (log-scale). Comparison between samples from in and outside of the detrital carbonate layer in core AI07-14G. The plots are averages of all the samples (errors bars = 1 standard deviation).

Grain size measurements indicate that the sediments consist mostly of silt (~80 %) and clay (~20%) and only very small amounts of sand [Pearce *et al.*, 2014]. The grain size analyses further show no significant difference in grain size distribution between samples within and out of the DC layer (Figure 4).

4. Discussion

4.1. The Heinrich 0 Detrital Carbonate Layer in AI07-14G

The elevated Ca/Sr ratios and the presence of both calcite and dolomite in the carbonate peak of AI07-14G point to an increased input of detrital material [Andrews *et al.*, 1995, 2012; Channell *et al.*, 2012], which is further confirmed by the results of the biomarker analysis. Samples from within the detrital layer contain elevated levels of diagenetic petrogenic compounds, with a biomarker composition being practically identical to the one reported in Hudson Strait sourced Heinrich layers from sediment cores

across the North Atlantic [Rosell-Melé *et al.*, 1997; Naafs *et al.*, 2013]. Of all measured parameters, the ITRAX measurements (XRF and gray scale) were carried out at the highest resolution (5 mm) and therefore allow for the most accurate identification of the boundaries of the detrital layer. At 296 cm depth, Ca/Sr ratios rise sharply and remain high until they gradually return to lower values around 245 cm. The abrupt onset of the event at 296 cm is dated at 11.51 ± 123 cal kyr B.P. [Pearce *et al.*, 2014], thus matching the timing of the end of H0 at the onset of the Holocene.

4.2. Source Area and Transport Mechanisms

The results of the qXRD and biomarker analyses are used to investigate the source of the carbonate in AI07-14G (Figures 2 and 3). The presence of detrital carbonate containing dolomite limits the potential source of the H0 layer to the Paleozoic carbonate outcrops of eastern Canada [Bond *et al.*, 1992; Kirby, 1998; MacLean, 2001; Naafs *et al.*, 2011]. The main sources include the Hudson Strait region and parts of Baffin Island, although for core AI07-14G, potential inputs of detrital carbonates from proximal sources such as western Newfoundland and areas along north shore of the Gulf of St. Lawrence and Anticosti Island should also be considered (Figure 1). However, as the detrital carbonate dispersal is assumed to be driven by glacial action (meltwater and iceberg discharges), these more local sources are less likely since during this time interval (YD) the ice sheet was situated largely north of the Paleozoic outcrops [Richard and Occhietti, 2005; Occhietti, 2007] and the residual Newfoundland ice cap was limited in size (Figure 1) [Dyke, 2004].

Based on the prevailing current system with the dominant influence of the southward flowing Inner Labrador Current (Figure 1) at our study site [Petrie and Anderson, 1983; Catto *et al.*, 2003], a northern origin for the carbonates seems likely, despite the greater distance of the corresponding source areas. Based on the qXRD analyses, calcite to dolomite ratios within the DC layer in our core are between 2 and 3, which is compatible with a Hudson Strait region source [Parnell *et al.*, 2007] as opposed to a Baffin Bay source, where this ratio is approximately reversed [Andrews *et al.*, 2012; Simon *et al.*, 2012]. The organic geochemical signature of the sediments in the DC layer (Figure 3) is nearly identical to that of Heinrich Layers 1, 2, 4, and 5 originating from Hudson Strait [Rashid and Grosjean, 2006; Naafs *et al.*, 2013], and thus also independently suggests a carbonate source in the Hudson Strait area.

To investigate the transport mechanism for the DC sediments to our study site, a detailed grain size analysis was carried out. Results show silt-clay-sized sediments in samples throughout core AI07-14G without any coarser IRD (Figure 4), ruling out the possibility of iceberg-rafting as the main sedimentation process this far south. Instead, the dominantly fine-grained sediments would suggest transport of the carbonates as shallow meltwater plumes [Hesse *et al.*, 1997] carried southward by the Labrador Current along the shelf. Surface plume transport has also been found to be responsible for depositing Hudson Strait sourced carbonates over long distances along the Labrador Shelf during earlier Pleistocene ice advances [Hesse *et al.*, 1997] and later during the Holocene drainage of Lake Agassiz [Lewis *et al.*, 2012]. Our results thus suggest long-distance transport of the detrital carbonate mostly as shallow-plumes with the Labrador Current along the shelf [Hesse *et al.*, 1997], although indication for some dispersal of the H0 detrital carbonates, deeper, along the slope and upper rise also exists [Stoner *et al.*, 1996; Veiga-Pires and Hillaire-Marcel, 1999]. Additionally, farther north on the Labrador Shelf (e.g., MD99-2236; Figure 1), there is distinct presence of IRD associated with the event, indicating ice rafting as an additional transport mechanism for sites closer to the source [Jennings *et al.*, 2015]. As outlined in the following section, the DC layer in AI07-14G can be correlated to sediment layers in several other cores, where these horizons have been associated with H0 originating from the Hudson Strait source area.

4.3. Heinrich 0 Layers on the Labrador Shelf

With a Hudson Strait source, the H0 DC layer should also be present in other marine sediment records from the Labrador Shelf, i.e., within the Inner Labrador Current trajectory, upstream of core AI07-14G. This is confirmed in a recent study of marine sediment core MD99-2236 from Cartwright Saddle on the Labrador Shelf (Figure 1), which shows a high-resolution sequence of detrital carbonate layers occurring in the very early Holocene [Jennings *et al.*, 2015]. The oldest major carbonate peak in core MD99-2236 is dated at ~ 11.5 cal kyr B.P., thus coinciding with the AI07-14G H0 event, and is also attributed to the termination of the H0 glacial event [Jennings *et al.*, 2015]. Cores HU96018-006 and HU96018-008 from Flemish Pass (Figure 1), just east of the Grand Banks of Newfoundland, also contain carbonate layers of similar age associated with H0 [Li and Piper, 2015]. Here the increased carbonate content and input of IRD were preceded by a strengthening of the Labrador Current indicated by significant changes in the grain size distribution [Li and Piper, 2015]. Heinrich 0 has also been reported in lower sedimentation rate cores from the western Labrador Sea: HU91045-06, HU2003033-11 (as P19), and HU91045-94 (Figure 1) [Hillaire-Marcel *et al.*, 1994, 2007; Rashid *et al.*, 2011], although here the chronology lacks the precision for a very robust correlation. Besides a low number of radiocarbon dates, this is also due to the impact of bioturbation and of variable fluxes of ^{14}C -carrier foraminifer shells as illustrated in Hillaire-Marcel and de Vernal [2008]. It is noted here that core HU2003033-011 was erroneously named core P19 in Hillaire-Marcel *et al.* [2007].

Stevenson *et al.* [2008] suggested a more local source for the H0 event, based on Sm-Nd and Nd isotopic data analyzed on sediment record MD95-2033 (Figure 1). Here a small isotopic excursion during the H0 carbonate event was assumed to be caused by the final melting of the remaining ice sheet covering eastern Newfoundland and the Grand Banks region [Stevenson *et al.*, 2008]. However, since detrital carbonates from this source and those from Hudson Strait are of similar age (lower Paleozoic), radiogenic isotope signatures cannot allow deciphering between the two sources. It is therefore likely that the observed H0 carbonate event in MD95-2033 is also linked to the Hudson Strait source.

4.4. Timing and Duration of Heinrich 0

The deposition of the H0 detrital carbonate layer has been linked to the Younger Dryas interval [Andrews *et al.*, 1995; Veiga-Pires and Hillaire-Marcel, 1999; Rashid *et al.*, 2011], with some suggestion for a temporal offset [Hillaire-Marcel and de Vernal, 2008]. In most of these earlier records, however, the time resolution achieved was unable to decipher the precise temporal linkage of H0 with the Younger Dryas cooling event, adding to the fact that the Younger Dryas is not clearly recorded in the northwest Labrador Sea record [Gibb *et al.*, 2014], neither at the outlet of the Gulf of St. Lawrence [de Vernal *et al.*, 1996]. In our studied record no detrital input was recorded during the Younger Dryas cold period itself until the deposition of H0 right after, very similar to what is observed on the Labrador Shelf in Jennings *et al.* [2015]. The age of onset of the event in AI07-14G was dated to approximately 11.500 ± 130 cal years B.P. and carbonate concentrations stayed elevated for about 100–150 years [Pearce *et al.*, 2014]. The increase of the sedimentation rate due to the enhanced flux of detrital carbonate during the interval was not more than $\sim 15\%$, based on the abundance

percent of calcite and dolomite above background (Figure 2), thus having a negligible impact on the estimated duration. This would indicate input of the detrital carbonate over a longer time period rather than a short-lived meltwater event. The estimated duration of the event is, however, limited by the uncertainties of the age model, possible changes in reservoir ages throughout the interval and sediment mixing at the upper limit of the DC layer (Figure 2) by bioturbation [Smith and Schafer, 1984; Anderson, 2001]. Therefore, the possibility of an abrupt, more intense meltwater pulse of shorter duration cannot be excluded.

The timing of our DC event, directly following the end of the Younger Dryas cold episode, argues for a link with the rapid warming at the onset of the Holocene and the associated retreat of the Younger Dryas ice margin in Hudson Strait, on the shelf and the surrounding landmasses [Stevenson *et al.*, 2008; Jennings *et al.*, 2015]. This implies that the H0 DC event as found in several records in the western Labrador Sea is in no way time equivalent to the entire Younger Dryas period, but corresponds to the rapid retreat phase of this ice advance at the onset of the Holocene. This is supported by Jennings *et al.* [2015], who assume that the main discharges of ice and meltwater out of Hudson Strait occur during either the ice advance phase or during the subsequent retreat (see also Pfeffer [2001]).

An important limitation, however, to the precise age determination of the DC event and correlation with other cores is the unknown marine radiocarbon reservoir offset, which is time and location dependent. The question of an appropriate reservoir offset ΔR for the region is fraught with arguments as different authors have used different values, ranging from $\Delta R = 0$ to an offset of several hundreds of years [Hillaire-Marcel *et al.*, 2007; Stevenson *et al.*, 2008; Rashid *et al.*, 2011; Hoffman *et al.*, 2012; Jennings *et al.*, 2015; Li and Piper, 2015]. Although some of these large ranges may be explained by the spatial extent of the different study sites, the variability of reservoir ages within the Labrador Current, from Hudson Strait down to the Grand Banks of Newfoundland, is expected to be very minimal [Sherwood *et al.*, 2008]. In recent years, several investigations have aimed to compensate for the potentially increased reservoir age during the early Holocene, caused by prolonged sea ice cover, by subtracting an additional 200 years from the conventional radiocarbon ages [Levac *et al.*, 2011; Lewis *et al.*, 2012; Rashid *et al.*, 2014]. However, although the ΔR value has likely varied through deglacial and glacial periods [Bard *et al.*, 1994; Olsen *et al.*, 2014], differences to modern reservoir age values such as presented by McNeely *et al.* [2006] and Reimer and Reimer [2001] are very difficult to ascertain without the presence of absolute age markers. Volcanic ash markers in marine sediment cores in the Labrador Sea are extremely rare, although Vedde Ash shards have been recognized in YD sediments from deeper cores from the Labrador Rise (e.g., Hillaire-Marcel and de Vernal [2008]). In this context, it should be noted that core AI07-14G has been thoroughly investigated for the presence of microtephra following the methods described in Turney [1998], but no volcanic ash horizons were found (see also Pearce *et al.* [2014]). Nevertheless, the applied ΔR of 139 ± 61 years for core AI07-14G [Solignac *et al.*, 2011] is consistent with other findings of the modern value for ΔR in this region: 132 ± 23 ^{14}C years for the northwest Labrador Sea [Sherwood *et al.*, 2008], 150 ± 60 ^{14}C years for SE Baffin Island [Coulthard *et al.*, 2010], and 144 ± 38 ^{14}C years for the Labrador Shelf [McNeely *et al.*, 2006]. However, as a result of the earlier mentioned challenges with radiocarbon dates and reservoir ages, the attained ages for H0 are associated with uncertainties of probably up to a few centuries. Despite this lack of precision, from analysis of other proxies in Pearce *et al.* [2013, 2014], where the carbonate layers follow a clear warming of the surface waters and loss of sea ice, it is clear that the H0 DC layer occurs right at the onset of the Holocene. This has the important implication that the H0 event itself did not seem to have any significant influence on large-scale circulation patterns, as it occurred during strengthening of the Atlantic Meridional Overturning Circulation (AMOC) [McManus *et al.*, 2004; Gherardi *et al.*, 2009].

4.5. Possible Link to Later Glacial Events

More recently, during the early Holocene, smaller and lesser known ice advances on the LIS in Hudson Strait (e.g., Noble Inlet and Gold Cove events) and the final Lake Agassiz drainage event are also known to have deposited detrital carbonate layers in the western Labrador Sea [Hillaire-Marcel *et al.*, 2007; Levac *et al.*, 2011; Rashid *et al.*, 2014; Jennings *et al.*, 2015; Pearce, 2015]. Therefore, the studied DC layer may, theoretically, have originated from the earliest of these events, the Gold Cove readvance. The Gold Cove glacial event was a very rapid expansion and subsequent retreat of an ice stream over the Hudson Strait in the early Holocene [Miller and Kaufman, 1990; Kaufman *et al.*, 1993; Pfeffer *et al.*, 1997; Manley and Miller, 2001; Andrews and MacLean, 2003; Dyke, 2004; Pearce, 2015]. The main reason for discarding the Gold Cove event as source

for the studied DC event, however, is its timing, since the onset of this re-advance occurred after the H0 event. It is relatively well dated to a maximum calibrated age of ~11.3 cal kyr B.P. (with absolute minimal reservoir age offset $\Delta R=0$) [Kaufman *et al.*, 1993; MacLean *et al.*, 2001; Pearce, 2015], which is too young when compared to the AI07-14G DC layer. The unlikelihood of the link with the Gold Cove advance is also confirmed by the carbonate stratigraphy of Cartwright Saddle [Jennings *et al.*, 2015], in which the Gold Cove and H0 are clearly different events represented by separate DC peaks.

5. Conclusions

Marine sediment core AI07-14G from Placentia Bay, Newfoundland, contains a DC layer, which is dated at ~11.5–11.3 cal kyr B.P. and attributed to the H0 glacial event. Sediments in the H0 layer are slightly lighter in color; have high Ca/Sr ratios; contain elevated concentrations of calcium carbonate, TOC, calcite and dolomite; and have a biomarker composition that is characteristic for Heinrich layers. The biomarkers are identical to those found in older Heinrich events and suggest a Hudson Strait region source, although input from other, more local, source areas cannot be totally ruled out based on the geochemical signature. However, the phasing of the H0 event and the sedimentological properties of the corresponding layer in cores from the Hudson Strait, Labrador, and Newfoundland shelves, as well as from deeper sites, suggest a single Hudson Strait source. The timing of the H0 DC event at ~11.5–11.3 cal kyr B.P. suggests a link with the rapid warming at the onset of the Holocene and the associated collapse and retreat of the Younger Dryas glacial ice masses. Grain size analysis showed no increased presence of coarser IRD during the event. This excludes the possibility of iceberg transport at our study site, but rather indicates surface water plume transport along the Labrador Shelf. The layer, observed from Hudson Strait to the SW slope of the Grand Banks, has thus great potential as a correlation tool between different sediment records in the western and southern Labrador Sea, aside its potential linkage to a short-duration climate event [Rasmussen *et al.*, 2007]. Another important finding is that the age of the meltwater event coincides with the timing of the enhancement of the Atlantic Meridional Overturning Circulation (AMOC) at the very end of the Younger Dryas [Piotrowski *et al.*, 2004; Gherardi *et al.*, 2009]. This is in agreement with earlier findings by Pearce *et al.* [2013] from the same sediment core, where the occurrence of H0 followed immediately after the surface water warming and rapid loss of sea ice associated with a strengthening of the AMOC. The H0 meltwater event itself thus had no direct measurable impact on the large-scale ocean circulation, as also suggested for similar scenarios in model experiments [Condron and Winsor, 2012].

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