1 **The deglaciation over Laurentian Fan: history of diatoms, IRD, ice and fresh water**

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11 **Abstract**

12 A high-resolution diatom census coupled with other proxy data from Laurentian Fan (LF)
13 provides a detailed description of the last deglaciation, bringing new insight to that period by
14 revealing directly the timing of sea-ice formation and melting. Cold events Heinrich Event 1
15 (H1) and the Younger Dryas (YD) were multiphase events. H1 (~16.8-15.7 cal kyr BP) was
16 defined by a two-pulse release of icebergs promoting sea-ice formation. Melting of sea-ice
17 after H1 corresponds to a cold and fresh anomaly that may have kept the Bølling colder than
18 the Allerød. At ~13.6 cal kyr BP, a cooling trend culminated with sea-ice formation, marking
19 the YD onset (~12.8 cal kyr BP). The decrease in sea-ice (~12.2 cal kyr BP) led to a YD
20 second phase characterized by very cold winters. However, the contribution of warm water
21 diatoms tends to increase at the same time and the YD gradual end (~11.6 cal kyr BP)
22 contrasts with its abrupt end in Greenland ice cores. The YD cannot be regarded as an event
23 triggered by a freshwater input through the Laurentian Channel since only one weak brief
24 input nearly 1000 yrs after its onset is recorded. Very cold and cool conditions without ice
25 mark the following Preboreal. A northward heat flux between 10.8 and 10.2 cal kyr BP was
26 interrupted by the increased influence of coastal waters likely fed by inland melting. There
27 was no further development of sea-ice or ice-drift then.

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30 **1. Introduction**

31 The North Atlantic exerts a strong influence on present day climate, with its
32 oceanographic changes affecting directly the climate of both sides of the basin (Dickson et al.,
33 1996; Visbeck, 2002). Fresh water input and the presence of ice in the North Atlantic are
34 assumed to have played a major role during abrupt climate changes through their impact on
35 the thermohaline circulation - THC (Broecker et al., 1989; Rahmstorf, 2002), in particular

36 during the deglaciation. However, there is no direct tracer for fresh water. Past fresh water
37 input is inferred from oxygen isotope measurements on planktonic and benthic foraminifera,
38 and the presence of melting icebergs is inferred from Ice Rafted Debris (IRD) in the
39 sediments (Bond et al., 1992; Heinrich, 1988; Hemming, 2004). Furthermore, the absence of
40 IRD does not necessarily mean the absence of ice, especially sea-ice. Regarding modeling
41 experiments, the massive discharge of icebergs during Heinrich events (HE) is often
42 simulated by a fresh water hosing (Flückiger et al., 2006; Liu et al., 2009; Prange et al., 2004),
43 but those simulations do not accurately reflect fresh water pathways in the ocean (Condon
44 and Winsor, 2011). Few studies have attempted to simulate an iceberg discharge with
45 progressive melting and transport (Jongma et al., 2012; Levine and Bigg, 2008; Roche et al.,
46 2014).

47 Diatoms are siliceous microscopic algae with the ability to colonize diverse
48 environments (including ice and fresh water) with diverse species. The diatoms preserved in
49 marine sediments therefore reflect directly the properties of the surface water (photic zone),
50 e.g. salinity, temperature and nutrients. They are also a major component of the phytoplankton
51 community in polar environments associated with ice (Medlin and Priddle, 1990). Diatom
52 assemblages can therefore offer an original description of the main climatic events of the last
53 deglaciation over our study area, the Laurentian Fan (LF).

54 The few diatom records from the North Atlantic for the deglaciation are mainly from
55 its equatorial and subtropical eastern margin (Romero et al., 2008; Stabell, 1986) or from the
56 subpolar North Atlantic and Nordic Seas (Andersen et al., 2004; Knudsen et al., 2004; Koç
57 Karpuz and Jansen, 1992; Koç et al., 1993). The only detailed deglacial marine record from
58 the western side of the North Atlantic, on Bermuda Rise, is unfortunately limited in time
59 because the diatoms are not preserved in the sediments after H1 (Gil et al., 2009). For this part
60 of the North Atlantic, only one study from the Gulf of St Lawrence (Lapointe, 2000a) is

61 available. Here, we provide a detailed description of the deglaciation in the north western
62 Atlantic with the aim of tracing directly the major forcings of the climate system that are
63 assumed to have played an important role: ice cover and the fresh water input.

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66 **2. Regional setting**

67 Laurentian Fan underlies part of the Slope Water System, where waters from subpolar
68 and subtropical origin meet. Gatién (1976) described two distinct water zones of opposite
69 characteristics: the Warm Slope Water Current, a bifurcation of the Gulf Stream and a well-
70 mixed water found between 0 and 400 m depth, flowing primarily eastward; and the cold
71 poorly mixed Labrador Slope Water flowing deeper and westward (Fig. 1). These water
72 masses are juxtaposed and their borders and flows vary in response to changes in intensity of
73 the Gulf Stream and of the Labrador Current (Pickart et al., 1999). A strong temperature and
74 salinity front results therefore from the competing influence of these water masses along the
75 North wall of the Gulf Stream. In addition, brackish and fresh water from the Gulf of St
76 Lawrence potentially overflows LF (Ohashi and Sheng, 2013). Finally, the site was located
77 near the Laurentide Ice Sheet (LIS) margin during the Last Glacial Maximum (LGM)
78 according to Dyke's reconstruction (2004). Diatoms (when preserved) appear therefore as an
79 appropriate proxy to describe the contrasted environments of the Slope Water System.

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82 **3. Material and methods**

83 **3.1 Material and age model**

84 Gravity core OCE326GGC14 (GGC14 – 43°03.959'N; 55°49.992'W; 3525 m depth)
85 was retrieved from LF (Fig. 1) in July 1998 on board the R/V Oceanus from the western
86 valley of LF(Keigwin et al., 2005).

87 GGC14 age model is based on seven AMS¹⁴C dates on planktonic
88 foraminifera(Keigwin et al., 2005). One additional ¹⁴C date at 189 cm (¹⁴C 10800 ±70 years)
89 was obtained from the National Ocean Sciences Accelerator Mass Spectrometer (NOSAMS)
90 facility at Woods Hole Oceanographic Institutionto better define the Younger Dryas in this
91 study. ¹⁴C ages were calibrated using the INTCAL13 curve (Reimer et al., 2013) with a
92 standard marine correction of 400 years. Age-depth modeling was elaborated using Clam 2.2
93 (Blaauw, 2010) R package. Ages were linearly interpolated at 95% confidence ranges using
94 Monte Carlo sampling (Fig. 2). The ages below the last dated level were extrapolated.

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96 **3.2Diatom analyses and ecology**

97 The core was sampled every 4 cm for diatom analyses. The samples were treated and
98 counted following the method described in Gil et al. (2009). The total diatom abundances are
99 reported as the median, minimum and maximum values (Fig. 3G).Diatom abundances were
100 calculated in 89 samples and 63 had diatoms abundant enough to determine the
101 assemblages. Three hundred diatom specimens (on a *Chaetoceros* free basis, as they are
102 common) were identified to calculate the contribution of each diatom species. The minimum
103 amount of 300 is required to assume that a contribution of 2% is significant with a confidence
104 level of 95.4% (Galehouse, 1971). The diatom flora consists of 260 species(cf. supplementary
105 dataincluding references for taxonomy) and a main environmental preference other than
106 marine (e. g fresh water, brackish, ice, cold and warm) was attributed to severalspecies.
107 The first step of diatom analysis was to comparethe GGC14 record with the
108 modernsurface distribution of diatoms in the Gulf of St Lawrence and estuary (Lapointe,

2000b). It was not conclusive: only one core in that study represented the open sea conditions overlying GGC14 and most of her species listed inhabit brackish waters. There is no reference to warm water diatoms reflecting the influence of the Gulf Stream as found in GGC14.

Another option would be to apply a transfer function. Although diatom transfer functions for the northern North Atlantic (including the Labrador Sea) are available (De Sève, 1999; Jiang et al., 2001; Justwan and Koç, 2008; Koç Karpuz and Schrader, 1990), they do not encompass our study area. No reliable transfer function for reconstruction is possible without a calibration.

We settled on a principal component analysis (PCA) including a varimax rotation using R (R core team, 2013). Species contributions of less than 2% were considered too rare for the analysis (Imbrie and Kipp, 1971) and removed. However, for comparison purposes, a low salinity diatom group was constituted with species considered brackish or freshwater even though they scored less than 2% (cf. supplementary data).

3.3 Ice Rafted Debris (IRD) and planktonic foraminifera > 150 µm

Bulk samples were dried, weighed, disaggregated in tap water and washed over a 63 µm mesh screen. The abundance of lithic grains (recognized to be IRD), the percentage of *Neogloboquadrina pachyderma* (sinistral), and the number of planktonic foraminifera was determined in the >150 µm fraction (in 99 samples for the IRD and 74 for foraminifera counts). When either the IRD or the foraminifera were very abundant, the samples were split before counting. Relative abundances of IRD components (quartz, hematite stained quartz and detrital carbonate grains) were estimated on a semi-quantitative basis.

4. Results

4.1 The diatom record: Principal Component Analysis

We present the loading scores for each factor (Fig. 3) and the individual contributions of the species that contributed to their construction (Fig. 4). The remaining species with a contribution >2% in more than one sample are plotted in Figure 5.

For the PCA, a total of 42 diatom entities (Table 1) were used and the eigenvalues indicate that 92.23% of the variance is explained by five factors. Factors 3 and 4 (made of species that were particularly abundant at a specific time and not all along the core) explain less than 7% of the variance.

Factor 1 (Fig. 3A) is primarily made by *Fragilariopsis oceanica* (Cleve) Hasle and secondarily by *Delphineis surirella* (Ehrenberg) Andrews. The former is known to thrive in ice and drifting ice and/or very cold environments (Hasle et al., 1996; Jiang et al., 2001; Justwan and Koç, 2008; Koç Karpuz and Schrader, 1990; Medlin and Priddle, 1990), whereas the second is found in “cool to temperate seas” (Andrews, 1981). This factor reflects cold conditions clearly dominated by ice.

Odontella aurita (Lyngbye) Agardh and *Thalassiosira gravida* / *antarctica* var. *borealis* Cleve resting spores contributed negatively to factor 2 (Fig. 3B). We refer to *Thalassiosira gravida* / *antarctica* var. *borealis* because there is some confusion regarding the nomenclature of *Thalassiosira gravida* Cleve. Although reference studies mention *Thalassiosira gravida* resting spores (De Sève, 1999; Justwan and Koç, 2008; Koç Karpuz and Schrader, 1990), this species is not supposed to form any (Syvertsen (1977) and cf. discussion in Hasle et al. (1996)). We choose to indicate both names as previously done (Jiang et al., 2001). *O. aurita* is regarded as a cosmopolitan species (Hasle et al., 1996), but this species is present in first year ice and is known to appear in monospecific blooms after particularly severe winters (Melnikov et al., 2002). It flourishes between -1.5°C and 2°C (Baars, 1985) and is abundant in fast-ice (McMinn et al., 2008). The formation of resting spores of *T. gravida* / *antarctica* var. *borealis* implies the degradation of its survival

conditions(McQuoid and Hobson, 1996), which in this case, might be the rapid establishment of very cold conditionslikely expressing a strong seasonal contrast. The score of factor 2 is reversed to make clearer that its interpretation is based on negative scores (Fig. 3B).

Factor 3(Fig. 3C) is constructed by the eurythermal and euryhaline diatom species*Paralia sulcata*(Ehrenberg) and *Thalassionema nitzschioides*(Grunow) Mereschowsky(Smayda, 1958). In the Nordic Seas, these species belong to the Norwegian-Atlantic Current assemblage influenced by Atlanticwaters, with salinitiesup to 34.9 andsea surface temperature above 3°C(Koç Karpuz and Schrader, 1990).*P. sulcata* is also a meroplanktonic species common in coastal plankton (Hasle et al., 1996) and isa factor on its own in Lapointe’s study (2000b) where it corresponds to coastal nutrient rich waters (summer temperature of 10-12°C and salinities of 26-29). This factor represents therefore coastal – temperate waters, an environment probably affected by rapid changes in salinity and temperature.

Thalassiosira oestrupii (Ostenfeld) Hasle constitutes factor 4. This species is found in warm to temperate waters (Hasle et al., 1996) and is clearly associated with warm Atlantic waters (Jiang et al., 2001; Justwan and Koç, 2008; Koç Karpuz and Schrader, 1990; Kohly, 1998; Maynard, 1976).

Finally, the opposite contribution of *D. surirella*(positive) and *Detonula confervacea* (Cleve) (negative) made factor 5.*D. surirella*is a benthic speciesassociated withcool to temperate waters(Andrews, 1981) and is common in marine and brackish coastal waters (Hendey, 1964).*D. confervacea* is an Arctic species that also appears in temperate waters during the cold season (Hasle, 1973). It is also reported from the water column associated with the sea-ice and in the ice itself (Ikeya et al., 2001; Robineau et al., 1997). Extreme low temperatures favor this speciesthat tolerates lower salinities, with a faster growth at a salinity of 25 (McQuoid, 2005)but not surviving salinities below 8 or temperatures above 15°C

(Guillard and Ryther, 1962). Cultures of this species show that a warming (from 2 to 7°C) increased its mean growth rate (Holt and Smayda, 1974). Factor 5 represents cool conditions, with negative extremes possibly indicating decreased surface salinity.

The absence in Lapointe's study (2000b) of some diatom species constructing the GGC14 PCA factors (as *F. oceanica*, *D. surirella*, *T. oestrupii*, *D. confervacea*) highlights the environmental and spatial variability of this region.

4.2 Diatom abundance, species contribution and factors loading

The record begins at ~19 cal kyr BP (Fig. 3G) and extremely low diatom abundances are found in the red-clay sediments (dotted box in Fig. 3 and 4) until about 17.6 cal kyr BP. Such red and brown sediments appear to be characteristic of the LGM over LFI and along the Scotian margin and preceded H1 (Keigwin and Jones, 1995; Piper and Skene, 1998).

From 16.1 cal kyr BP, the diatom abundances started to increase and diatoms were still abundant until 14.1 cal kyr BP. Then, and until 12.8 cal kyr BP, abundances were relatively low before rising and reaching their maximum abundance close to 12 cal kyr BP. After that, the abundances diminished constantly and reached low values towards the end of the record (~9.4 cal kyr BP).

The ice record (factor 1 in Fig. 3A explaining 65.47% of the variance) is characterized by two general decreases: from the beginning of the record until 13.9 cal kyr BP and from 12.7 to 9 cal kyr BP (accentuated minima at ~12.4 and 11.8 cal kyr BP). The minima at ~16.3, 15.7, 14.4 and 13.7 cal kyr BP separate the highest levels. Factor 2 (11.76% of the variance) was most notably negative between 15.6 and 13.9 cal kyr BP (less ~15 cal kyr BP) and increasingly negative from 13.7 to 10.8 cal kyr BP, when the trend was reversed. *O. aurita* (Fig. 4D) explains the factor variability between 15.6 and 15 cal kyr BP and at ~12.4 cal kyr BP. The variability of factor 3 (6.44% of the variance; Fig. 3C) is limited and the most

relevant feature is its score increase from 11.7 to 9 cal kyr BP. Factor 4 indicates (5% of the variance) warmer conditions starting around 14.6 cal kyr BP (Fig. 3D) that lingered until ~12.8 cal kyr BP, except notably ~14 cal kyr BP. Then, the conditions started to warm slowly again from 12.2 to 10.5 cal kyr BP before decreasing. Factor 5 explains little of the variance (3.56%). The most distinctive features are related to *D. surirella* with higher contributions at 16.3, ~13.9 and 14.8 cal kyr BP. *D. confervacea* contributed clearly to the variability recorded between 12.2 and 11.7 and at ~13.9 cal kyr BP. Finally, the contribution of low salinity diatoms was mostly significant (>2%) from 17.2 to 14 cal kyr BP (Fig. 3F), except principally around 15.4 cal kyr BP.

4.3 The IRD and the *N. pachyderma* (s) records

Grains of quartz and detrital carbonate >150 µm, considered to be IRD started to appear in the sediment at 18.2 cal kyr BP, with low values (~10/g) (Fig. 3H, Fig. 6). The abundances rose at 16.8 cal kyr BP (5449/g), initiating the period of highest abundances that lasted 500 years. This interval, referred to here as pulse A, has the only significant detrital carbonate representation, which we estimate to be about 20% of the grains >150 µm, between about 376 and 382 cm (Fig. 6). Hematite stained quartz never dominates the IRD, but reaches 40-45% at the beginning of H1A, and 40-50% at the end of H1B. Otherwise, the hematite-stained grains range from nearly absent to 30% of the lithic fraction. Total IRD registered low abundances between 16.3 and 15.9 cal kyr BP (~500/g), before a second period of high abundance from 15.9 to 15.7 cal kyr BP. IRD abundances were extremely low again, until the mid-Allerød to the mid-YD (13.4 to 12.2 cal kyr BP).

The *N. pachyderma* (s) record (Fig. 3J) covers the period between ~11.3 and 18.1 cal kyr BP. The highest values (between 60 and 100%) are recorded during the intervals 18.1-15.5 (except at 16.5 and 15.9) and 12.1-12.7 cal kyr BP. From 14.8 to 13.8 cal kyr BP, the

contribution increased from 15 to 53%, between periods of minimum values (15.2-14.8 and 13.3-12.8 cal kyr BP).

5. Discussion

5.1 Anatomy of Heinrich Event 1 on Laurentian Fan

IRD in marine sediments helps to identify the HE and define them *sensus stricto*. However, our record reveals a succession of oceanographic changes far beyond the IRD deposits, but directly associated with the iceberg discharges and melting. H1 had therefore a longer temporal expression than the simple record of IRD and this period fits within the range of the “Mystery Interval” (Denton, 2006). In GGC14, H1 is indicated by two distinct releases of IRD at LF: the first, pulse A (H1A) between 16.8 and 16.3 cal kyr BP, and a second IRD pulse B (H1B) from 15.9 to 15.7 cal kyr BP (Fig. 3H). GGC14 IRD deposits associated with H1 are about 32 cm thick (12 cm for H1A alone). This is thicker than in nearby cores (deGelleke et al., 2013), but is the same order of magnitude if only H1A is considered. GGC14 can also present higher sedimentation rates. Because of its distinctive detrital carbonate component, H1A is likely related to a detrital carbonate 1 (DC-1) interval previously interpreted to indicate a dominant Labrador Current supply of icebergs that later mixed with icebergs from the Gulf of St Lawrence over LF (Piper and Skene, 1998). Hematite stained quartz grains indicating a supply from the Gulf of St Lawrence are only notably recorded at the beginning and at the end of H1. The two main IRD pulses in our LF record are consistent with our previous work on Bermuda Rise indicating that H1 is made of more than a single pulse of icebergs (Gil et al., 2009).

H1A was longer (~500 yrs) and has about twice as many grains per gram as H1B. The maximum in IRD seems to match a small but distinct warming of the NGRIP record which

we use here to provide a Northern Hemisphere context to our results (Fig. 3L). H1A was preceded by a relative higher contribution in low salinity water diatoms (Fig. 3F), but their contribution rapidly decreased, while cool and coastal waters prevailed (Fig. 3E). The total diatom abundance also increased progressively. H1A was followed by a ~200 yrs period of reduced iceberg discharge. During this pause, ice and cool coastal low salinity diatoms increased, marking the development of sea-ice. The NGRIP record also indicates a cooling. At the same time, there was no low salinity water advected by cold core rings over the Bermuda Rise or resulting from *in situ* melting icebergs (Gil et al., 2010; 2009), supporting the pause in the iceberg discharge.

During H1B (~200 yrs), IRD abundances were lower and ice diatoms clearly dominated. *N. pachyderma* (s) (Fig. 3J) recorded its highest abundance. Perhaps at that time pervasive sea-ice and icebergs coexisted. Abundances of low salinity diatoms were highly variable during the event, but are relatively more abundant shortly before H1B.

Similarly to LF, two peaks in IRD are also recorded at the Northwest European margin (Walden et al., 2006) and at the Portuguese margin precisely at 17.5 and 16 cal kyr BP (Bard et al., 2000). The variability of the Iberian continental vegetation also points to two distinct phases during H1: the first was characterized by cold and wet conditions while the second corresponded to less cold conditions and increased dryness (Naughton et al., 2009). The west-east variability of these records also suggests a differential spatial melting of icebergs and IRD release. The IRD pulses over LF are accompanied by low $\delta^{18}\text{O}$ and are interpreted as a freshening (Keigwin et al., 2005), but the contribution of the fresh water diatoms does not clearly mark the pulses. The formation of sea-ice might rather explain the low $\delta^{18}\text{O}$ values, as proposed by Hillaire-Marcel and de Vernal (2008). Indeed, the diatom record indicates the development of sea-ice in between the pulses and during H1B. During H1A, the diatoms were too few (due to numerous IRD and a minor fraction of fine

sediment possibly containing diatoms) to obtain a statistically reliable result in some samples. A large volume of sediment would have been necessary to check if those samples would also contain ice diatoms, as suggested by Sancetta (1992).

The ice diatom assemblage culminated between ~16.2 and 15.8 cal kyr BP, roughly between the two IRD pulses, and its decline by the end of H1 corresponded to the clear rise of the diatom *O. aurita* (Fig. 4D). It seems to be the response to the end of the iceberg discharge and evidences the first step of sea-ice waning. The thinning of the sea-ice and the extremely low abundance of IRD suggest that the strong cooling sustaining the sea-ice and sea-ice formation was likely imposed by icebergs. The sudden decrease of *O. aurita* (~15.6 cal kyr BP) is compensated by the rise in low salinity diatoms from ~15.3 to 14.6 cal kyr BP (Fig. 3F), likely representing the major melting of this sea-ice previously formed, even though some sea-ice is still likely present. Thus from the diatom perspective, the freshening signal resulting from H1 appears after the event. HE are often only considered as a low salinity anomaly in the North Atlantic when it is principally a strong cooling promoting sea-ice formation according to the diatom record (Fig. 3A), at least at this site.

Finally, Mg/Ca inferred temperatures (calcification temperature - CT) in this core (Fig. 3I) indicate a gradual warming (Obbink et al., 2010) that culminates at 16.6 cal kyr BP during H1A. That result contrasts with the high variability of the diatom record, the low $\delta^{18}\text{O}$ of *N. pachyderma* (s), and the high percentage of that species (>90%) during H1. These results are consistent with previous studies that show very low temperatures associated with drifting ice and sea-ice can possibly bias *N. pachyderma* (s) calcification temperatures that are based on Mg/Ca (Kozdon et al., 2009; Meland et al., 2006).

5.2 The Bølling, the Older Dryas and the Allerød

The Bølling-Allerød is particularly well distinguished by factor 4 and the warm water diatom *T. oestrupii* (Fig. 3D and 4G). Their variability fit the evolution of the NGRIP $\delta^{18}\text{O}$ record (Rasmussen et al., 2006) within dating errors and calcification temperatures (CT) (Obbink et al., 2010) (Fig. 3I). Based on diatoms, the Allerød period appears relatively warmer than the Bølling, while the ice core and CT records indicate the opposite. The lack of significant presence of low salinity water/meltwater during the Allerød might explain such a difference by favoring a strong mode of the THC. Indeed, the onset of the Bølling followed the important melting of sea-ice and a strong seasonal contrast prevailed during the Bølling, with the near freezing cold water foraminifera *N. pachyderma* (s) still abundant (Fig. 3J). The role of meltwater in the transition to the Bølling has already been studied through modeling experiments (Liu et al., 2009) and the results revealed that the abrupt warming would be due to the cessation of the meltwater flux (a sudden shut-off at 14.67 cal kyr BP or a linear decrease to 0 at 14.2 cal kyr BP) and not to the resumption of the Atlantic Meridional Overturning Circulation (AMOC). Our results support the idea that the presence of meltwater modulates the intensity of the warming. The presence of low salinity water during the Bølling at our site is not necessarily contradictory with the modeling results (Liu et al., 2009), depending on the sensitivity of the model they used to fresh water forcing and where the fresh water was prescribed.

The Bølling and the Allerød are separated by a decrease in the warm water factor, which corresponds also to an increase in ice, coastal and low salinity diatoms centered ~14 cal kyr BP (Fig. 3ACEF). Mg/Ca inferred salinity in GGC14 also indicates a fresh water discharge (Obbink et al., 2010). Oddly, *N. pachyderma* (s) was not outstandingly more represented (Fig. 3J), but its contributions was still significant (~20%). By the same time, a major calving ice retreat event occurred (Shaw et al., 2006). This freshening and interruption of the warming is known in terrestrial records as the Older Dryas (12 000 – 11 800 ^{14}C BP in

Mangerud et al. (1974)). Previous high-resolution studies advocated that the abrupt cooling of the Older Dryas would correspond to the meltwater pulse 1a (MWP 1a) (Stanford et al., 2006), while coral measurements from Tahiti indicate that the MWP1a was simultaneous to the Bølling onset (Deschamps et al., 2012). If the hypothesis that meltwater from the LIS would have had contributed to the MWP 1a is valid (Carlson et al., 2007), the larger input would have preceded the Bølling although low salinity is still significantly present over the site during the second half of the Bølling and the Older Dryas. Without a better characterization of the contribution of the LIS to MWP 1a, it is not possible to indicate the timing of its imprints over LF sediments.

From the Older Dryas to the Allerød, the contribution of low salinity diatoms dropped and clearly marks the return to warmer conditions, with a maximum warming at ~ 13.8 cal kyr BP.

5.3 The Younger Dryas

The YD was preceded by mid-Allerød (~13.8 cal kyr BP) changes including increased abundance of ice related diatoms and IRD, and a decreased contribution of warm water diatoms. Broecker et al. (2010) already pointed out that the slow onset of the YD showed in numerous records argues for a non-catastrophic event. The elevated contribution of ice-related diatoms between 12.7 and 12.2 cal kyr BP clearly marked the onset of the YD, while from ~12.2 ka a slow increase in warm water diatoms (Fig. 3D) and *D. confervacea* (Fig. 4H) marked a second phase (12.4 – 11.6 cal kyr BP).

The initial peak in ice related diatoms is associated with few IRD and very cold waters testifying the presence of some icebergs. These icebergs could have been brought through the Laurentian Channel and / or a stronger Labrador Current and promoted sea-ice formation, as during H1. A major contribution of *O. aurita* (12.6 – 12.4 cal kyr BP) followed, similarly to the

diatom succession observed during and after H1. Contrary to H1, there was no evidence for a
 low salinity anomaly afterwards. Instead, the concomitant increase in warm and cool water
 diatoms (including *D. confervacea*) points to milder conditions and may reflect also a
 seasonal contrast. A similar warming trend is recorded on the west side of the Laurentian
 Channel and on the Scotian Shelf (Levac et al., 2015). Relatively warmer waters would also
 favor an increased growth rate of *D. confervacea* (Holt and Smayda, 1974). The tolerance of
D. confervacea to lower salinities, its affinities with sea-ice and its prevalence during the YD
 phase 2 is relevant since the YD has been hypothesized to result from a fresh water outburst
 from Lake Agassiz into the North Atlantic preventing the formation of North Atlantic Deep
 Water – NADW (Broecker et al., 1989). With only one brief significant signal of low salinity
 water input nearly 1000 yrs after the YD onset (Fig. 3F), the low salinity anomaly cannot be
 considered a striking feature of the YD in the slope water system. Our record is consistent with
 the interpretation of previous dinocyst records (de Vernal et al., 1996; Levac et al., 2015): low
 temperatures, thin sea-ice cover and reduced meltwater from the outlet of the Gulf of St
 Lawrence into the Atlantic. This interpretation, however, contradicts the results obtained from
 Mg/Ca on foraminifera (Carlson et al., 2007) that might have been strongly influenced by the
 development of sea-ice and the presence of icebergs (notably during phase 1). The potential
 seasonal sea-ice formed during the second phase of the YD differs from the one previously
 formed (factor 1) by its thickness and persistence. The mix of warm and cold conditions
 during the second phase of the YD suggests indeed that the thin sea-ice was related to the
 episodic presence of very cold waters. The high abundance of *D. confervacea* associated with
 the decline of thin sea-ice and enhanced primary productivity by the end of the event is also
 detected in a Newfoundland shallow coastal site (Pearce et al., 2013; 2014).

The dual feature of the YD is present elsewhere in several high-resolution records,
 principally from lakes (Bakke et al., 2009; Brauer et al., 2008; Neugebauer et al., 2012), but

also in oceanic sites such as Cariaco basin (Lea et al., 2003) or off Northwest Africa (Cole et al., 2009). These studies point to an early YD (first phase) that was windier off Northwest Africa and colder over the Cariaco basin. Over LF, the conditions were also colder, with more abundant ice related diatoms attesting the development of seasonal sea-ice. Such conditions would have shifted the Inter Tropical Convergence Zone southward –ITCZ (Chiang and Bitz, 2005) and it would explain the Northwest Africa and Cariaco records. The decrease of *D. confervacea* clearly marks the end of the YD. An outstanding contribution of the warm water diatoms does not mark the end of the event as it would be expected. Instead, this assemblage presents a steady evolution contrasting with the warming indicated by the NGRIP record.

5.4 The early Holocene

The Holocene record of GGC14 confirms the continuous decline of ice. Very cold and cool conditions prevailed until ~10.8 cal kyr BP. These conditions could have been established rapidly (cf. section 4.1) and represent in part the Preboreal Oscillation cold event (Hald and Hagen, 1998). After ~10.8 cal kyr BP, the waters warmed and the influence of coastal and temperate waters conditions continued to progress, even more from ~10.2 cal kyr BP when the waters cooled again. The warming between ~10.8 and 10.2 cal kyr BP supports a stronger AMOC with a northward penetration of the heat flux initiating the retreat of the ice-margin at Disko Bugt on the Western Greenland Ice sheet (Kelley et al., 2013) and getting Frobisher Bay (Labrador Sea) ice-free from at least 10.5 to 10.2 cal kyr BP (Miller and Kaufman, 1990). This warming was equivalent to the Bølling warming (Fig. 3D). The halt of the heat flux and northern melting is likely due to the routing of some LIS meltwater to the Gulf of Mexico (Aharon, 2003; Flower and Kennett, 1990; Poore et al., 2003; Rasmussen and Thomsen, 2012), however the evidences for such routing are missing in some cores of the region

(Marchitto and Wei (1995) and references herein). The progression of the coastal water over LF suggests also that the cooling was related to the melting of ice on land that started after ~13 cal kyr BP (Shaw et al., 2006) and especially after 10 cal kyr BP (Shaw et al., 2002). These increased coastal waters did not bring ice or lead to very cold conditions and, even if the timing is similar, there is no evidence for a catastrophic event as the Gold Cove event (Kaufman et al., 1993). Finally, the increased volume of coastal-temperate waters was deduced from the high contribution of *P.sulcata* that is also found in a shallower (1381 m) nearby core (Lapointe, 2000a).

6. Conclusion

The diatom record shows that the major well-known climatic periods of the last 19 ka were in fact part of a sequence. H1 (and consecutive oceanographic changes) and the YD were multiphase cold events initially described by common diatom assemblages: ice related diatoms followed by the outstanding contribution of *O. aurita*. The two steps release of icebergs during H1 had a strong cooling effect that promoted the development of sea-ice. Between H1 and the YD, the formation and melting of sea-ice played an important role, with melting sea-ice preceding the warm Bølling and attenuating the initial warming. This warm period, as the Allerød, was clearly indicated by the highest rates of warm water diatoms of the deglacial record. The interruption of the warming between the Bølling and the Allerød represents the Older Dryas (~14 cal kyr BP). The YD was announced by a long cooling trend initiated during the mid-Allerød. The YD started at ~12.8 cal kyr BP and had two main phases: until ~12.42 cal kyr BP, when some IRD are still found in the sediments and the conditions dominated by sea-ice; and a second phase marked by a strong seasonal contrast since the presence of a thin seasonal sea-ice is accompanied by the increased contribution of warm

water diatoms. Although the dominant diatom species of this period (*D. confervacea*) can tolerate low salinity water, there is no clear evidence for a massive surface freshening.

Our records highlight the importance of sea-ice during the last 19 ka and the role of surficial low-salinity water in modulating the intensity of the warming. The spatial and temporal extent of sea-ice in the North Atlantic, and its melting, remain important in understanding how climate system recovers from such abrupt coolings or salinity anomalies.

Keywords: deglaciation; sea-ice; North Atlantic; diatom; Heinrich Event 1; Younger Dryas; Bølling-Allerød; Older Dryas; Preboreal

Figure captions

Table 1: Factors score matrix and corresponding variance from GGC14 diatom assemblages

Figure 1: Location of core GGC14 and modern surface water distribution in the region of the Grand Banks of Newfoundland after Pickart et al. (1999) and Petrie and Anderson (1983). Cold and warm currents are represented by dashed and plain arrows respectively.

Figure 2: GGC14 age-depth modeling obtained with CLAM 2.2 R package (Blaauw, 2010). Points on the curve and error bars represent AMS dates on planktonic foraminifera calibrated to calendar years assuming the standard 400 year reservoir effect and calibration probability. Grey area illustrates the 95% probability range.

Figure 3: General paleoceanographic data from GGC14 versus time: A-Factor 1 / ice; B-Factor 2 / very cold; C-Factor 3 / coastal-temperate; D-Factor 4 / warm; E-Factor 5 / cool; F-Low salinity diatoms (in %); G- Total diatom abundance (number of valves per gram):

median (line) and maximum and minimum values (dots); H- IRD (>150 μm) abundances in number of grains per gram; I-*N. pachyderma* (s) calcification temperature – CT in $^{\circ}\text{C}$ (Obbink et al., 2010); J- Abundance of *N. pachyderma* (s) (in %); K- $\delta^{18}\text{O}$ (in ‰) of *N. pachyderma* (s) (Keigwin et al., 2005); L- $\delta^{18}\text{O}$ of NGRIP ice core (in ‰) (Rasmussen et al., 2006). The dotted box indicates the red-clay sediments. Arrows on the age scale indicate AMS ^{14}C dates and the grey numbers their depth (cm).

Figure 4: GGC14 diatom species contribution (in %) versus time: A- *Fragilariopsis oceanica*; B- *Delphineis surirella*; C- *Thalassiosira antarctica* var. *borealis/gravida* resting spores; D- *Odontella aurita*; E-*Paralia sulcata*; F-*Thalassionema nitzschioides*; G- *Thalassiosira oestrupii*; H- *Detonula confervacea*. The dotted box indicates the red-clay sediments. Arrows on the age scale indicate AMS ^{14}C dates and the grey numbers their depth (cm).

Figure 5: Remaining GGC14 diatom species with a contribution > 2% in more than one sample versus time. Only Heinrich event 1, the Bølling-Allerød, the Younger Dryas and the Preboreal (PB) are indicated.

Figure 6: GGC14 semi-quantitative estimate of IRD composition (in %): A- quartz grains; B- detrital carbonate grains with the asterisks indicating rare occurrence (<5%); C- hematite stained quartz grains; D- IRD (>150 μm) abundance (IRD/g).

Acknowledgements

This research is a contribution to the Fundação para a Ciência e Tecnologia - FCT project MONA (PTDC/AAC-AMB/108449/2008) managed by CIIMAR (Interdisciplinary Centre of Marine and Environmental Research, Porto, Portugal) and partially hosted by LNEG (The

National Laboratory for Energy and Geology, Alfragide, Portugal). Financial support was provided by FCT to IMG through postdoctoral fellowships (BPD/27214/2006 and BPD/877185/2012) and the POPH/FSE program. This research was also partially supported by the European Regional Development Fund (ERDF) through the COMPETE - Operational Competitiveness Programme and national funds through FCT, under the project “PEst-C/MAR/LA0015/2011. This work was also supported by NSF grant OCE0822854 to LDK. Thanks to A. Inês for support in the sample preparation for diatom analysis at LNEG. We finally also thank the reviewers for their constructive comments and suggestions.

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Table 1

Diatom species	Factor 1	Factor 2	Factor 3	Factor 4	Factor 5
<i>Actinocyclus octanarius</i>	0,010	-0,002	0,026	0,004	-0,002
<i>Actinopterychus splendens</i>	-0,004	-0,008	0,0158	0,082	0,029
<i>Azpeitia africana</i>	-0,002	0,003	0,02	-0,003	-0,002
<i>Azpeitianeocrenulata</i>	-0,006	-0,015	0,013	0,005	-0,038
<i>Azpeitianodulifera</i>	-0,007	-0,003	0,031	0,006	-0,006
<i>Bacteriosira fragilis (spore)</i>	-0,008	-0,013	0,058	0,020	-0,108
<i>Bacteriosira fragilis</i>	0,001	-0,004	0,035	0,013	-0,076
<i>Coscinodiscus divisus</i>	0,011	-0,004	0	-0,003	-0,0008
<i>Coscinodiscus kutzingii</i>	0,005	-0,021	0,007	-0,006	0,006
<i>Coscinodiscus marginatus</i>	0,024	0,005	0,049	0,019	0,035
<i>Coscinodiscus radiatus</i>	-0,001	-0,006	0,012	0,003	-0,008
<i>Delphineis surirella</i>	0,416	-0,026	0,028	0,237	0,672
<i>Detonula confervacea</i>	0,017	-0,018	0,067	-0,001	-0,561
<i>Fragilariopsis cylindrus</i>	0,02	0,003	0,0006	0,001	-0,005
<i>Fragilariopsis doliolus</i>	-0,021	-0,026	0,056	0,031	-0,093
<i>Fragilariopsis linearis</i>	0,004	0,002	0,003	0,001	0,0007
<i>Fragilariopsis oceanica</i>	0,885	0,029	0,082	-0,006	-0,351
<i>Fragilariopsis sublinearis</i>	0,055	0,002	0,009	-0,0004	0,032
<i>Odontella aurita</i>	-0,048	-0,614	0,007	0,045	-0,069
<i>Paraliasulcata</i>	0,042	0,042	0,918	-0,008	0,033
<i>Porosira glacialis</i>	0,01	0,005	0,013	0,005	-0,017
<i>Rhizosolenia bergonii</i>	-0,006	0,004	0,039	0,015	-0,016
<i>Rhizosolenia setigera</i>	-0,002	0,0008	0,018	-0,001	-0,005
<i>Rhizosolenia simplex</i>	-0,002	0,002	0,018	-0,002	-0,003
<i>Rhizosolenia styliformis</i>	-0,003	-0,007	0,035	-0,008	-0,027
<i>Rhizosolenia spp</i>	0,0009	0,007	0,125	-0,016	0,004
<i>Roperia tessellata</i>	-0,011	-0,015	0,045	0,037	-0,030
<i>Thalassionema nitzschioides</i>	0,026	0,014	0,31	0,155	-0,111
<i>Thalassiosira aff. constricta</i>	0,07	-0,01	0,003	-0,001	0,077
<i>Thalassiosira angulata</i>	0,08	0,012	0,044	0,079	0,104
<i>Thalassiosira anguste-lineata</i>	-0,018	-0,061	0,008	0,038	-0,082
<i>Thalassiosira antarctica</i> var. <i>borealis</i>	0,002	0,007	0,005	0,004	-0,012
<i>Thalassiosira eccentrica</i>	0,009	-0,023	0,049	0,043	-0,002
<i>Thalassiosira ferrelinaea</i>	-0,003	0,0005	0,0003	0,036	-0,028
<i>Thalassiosira</i> <i>gracilis/antarctica</i> var. <i>borealis (spore)</i>	0,051	-0,775	0,034	0,040	0,031
<i>Thalassiosira gracilis</i>	0,044	0,0004	0,036	0,028	-0,038
<i>Thalassiosira hyalina</i>	0,04	0,003	0,004	0,013	-0,011
<i>Thalassiosira lineata</i>	-0,007	-0,026	0,022	0,034	-0,076
<i>Thalassiosira nordenskiöldii</i>	0,019	-0,043	0,025	0,0457	0,007
<i>Thalassiosira ostrupii</i>	-0,11	0,076	0,063	0,942	-0,146
<i>Thalassiosira spp</i>	0,014	-0,045	0,0003	0,004	0,046
<i>Thalassiotrix longissima</i>	0,019	-0,006	0,025	0,014	0,019
variance	65,47	11,76	6,44	5	3,56
cumulative variance	-	77,23	83,67	88,67	92,23

Figure 1

Figure 2

Figure 4

Figure 6

Graphical Abstract