A new concept for the paleoceanographic evolution of Heinrich event 1 in the North Atlantic

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New records of planktonic foraminiferal δ18O and lithic and foraminiferal counts from Eirik Drift are combined with published data from the Nordic Seas and the “Ice Rafted Debris (IRD) belt”. To portray a sequence of events through Heinrich event 1 (H1). These events progressed from an onset of meltwater release at ~19 ka BP, through the ‘conventional’ H1 IRD deposition phase in the IRD belt starting from ~17.5 ka BP, to a final phase between 16.5 and ~ 15 ka BP that was characterised by a pooling of freshwater in the Nordic Seas, which we suggest was hyperpycnally injected into that basin. After ~ 15 ka BP, this freshwater was purged from the Nordic Seas into the North Atlantic, which preconditioned the Nordic Seas for convective deep-water formation. This allowed an abrupt re-start of North Atlantic Deep Water (NADW) formation in the Nordic Seas at the Bølling warming (14.6 ka BP). In contrast to previous estimates for the duration of H1 (i.e., 1000 years to only a century or two), the total, combined composite H1 signal presented here had a duration of over 4000 yrs (~19–14.6 ka BP), which spanned the entire period of NADW collapse. It appears that deep-water formation and climate are not simply controlled by the magnitude or rate of meltwater addition. Instead the location of meltwater injections may be more important, with NADW formation being particularly sensitive to surface freshening in the Arctic/Nordic Seas.

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1. Introduction

Heinrich (H) events are characterised in North Atlantic sediments by horizons with increased Ice Rafted Debris (IRD) concentrations, low foraminiferal abundances, and light planktonic foraminiferal calcite δ18O (meltwater dilution). They occurred quasi-periodically with a spacing of 5000–14,000 yrs (e.g., Heinrich, 1988; Broecker, 1994; Broecker et al., 1992; Bond et al., 1992, 1999; Bard et al., 2000; Rohling et al., 2003; Hemming, 2004), and most likely ~7200 yrs (van Kreveld et al., 2000). These layers are distinct in marine sediment cores removed from the so-called “IRD belt” (40°N–55°N), but they can also be recognised throughout most of the North Atlantic (e.g., Heinrich, 1988; Broecker, 1994; Broecker et al., 1992; Bond et al., 1992, 1999; Bond and Lotti, 1995; Cortijo et al., 1997; Hemming et al., 2000, 2002; Hemming and Hajdas, 2003; Grousset et al., 1993, 2000, 2001; for a review, see Hemming, 2004). IRD can even be found as far south as the subtropical gyre; for example, the north Sargasso Sea (Keigwin and Boyle, 1999; Benetti, 2006; Gil et al., 2009) and the Iberian margin (Bard et al., 2000).

It is thought that H events mainly represent periodic collapses of the Laurentide ice sheet (MacAyeal, 1993), although there are strong indications that the Greenland, Icelandic, Fennoscandian, and British ice sheets were also involved (e.g., Bond et al., 1997, 1999; Scourse et al., 2000; Knutz et al., 2001, 2007; Grousset et al., 2001; Hemming et al., 2000; Hemming, 2004; Jullien et al., 2006; Nygård et al., 2007; Peck et al., 2006, 2007b). Furthermore, records from around the North Atlantic, and even throughout the Northern Hemisphere, indicate dramatic marine and terrestrial temperature reductions and increased aridity during H events (e.g., Atkinson et al., 1987; Alm, 1993; Bond et al., 1992; Mayewski et al., 1994, 1997; Vidal et al., 1999; Broecker, 2000; Bard et al., 2000; Gasse, 2000; Rohling et al., 2003; Hemming, 2004). The most widely accepted hypothesis holds that the low temperatures associated with these events resulted from reduced oceanic poleward heat transport due to surface freshwater dilution in the North Atlantic and a consequent shutdown of the Atlantic meridional overturning circulation (AMOC) (e.g., Broecker, 1991; Rahmstorf,
Sediment cores from the IRD belt suggest a common age for H1 (the last major H event occurring at the onset of the deglaciation) between 16 and 17.5 thousand years before present (ka BP, where present refers to AD 1950) (e.g., Bond et al., 1992, 1997, 1999; Bard et al., 2000; Grouset et al., 2001; Rohling et al., 2003; Hemming, 2004; McManus et al., 2004), in which case it had terminated well before the abrupt Bølling warming (14.64 ka BP) (Lea et al., 2003; Rasmussen et al., 2003). Coincident with this timing of H1, 231Pa/230Th ratios recorded in a Bermuda Rise sediment core GGC5 suggest virtually complete AMOC shutdown (McManus et al., 2004). However, this proxy for AMOC intensity suggests a gradual slowdown from around 19 ka BP into the AMOC collapse, and a sharp AMOC resumption coincident with the Bølling warming (McManus et al., 2004). The 231Pa/230Th method underlying these reconstructions has been challenged (e.g., Keigwin and Boyle, 2008; Gil et al., 2009) but, the GGC5 231Pa/230Th record has been independently corroborated by magnetic grain size data from core TTR13-AT451G (1927 m water depth; 30°30.886′N, 44°54.333′W) (hereafter referred to as TTR-451), recovered from Eirik Drift, offshore of S. Greenland (Stanford et al., 2006). The timing presented by McManus et al. (2004) of the gradual slowdown from around 19 ka BP to full AMOC collapse at around 17.5 ka BP has also been corroborated by a 231Pa/230Th record from a drift deposit in the Rockall Trough (Hall et al., 2006). Reduced deep-water ventilation and, hence, decreased AMOC intensity during H1 have also been inferred from δ13C records, and from deep-water 14C ages and Cd/Ca ratios (e.g., Boyle and Keigwin, 1987; Boyle, 1992; Sarnthein et al., 1994; Curry et al., 1999; van Kreveld et al., 2000; Robinson et al., 2005; Marchitto et al., 2007; Keigwin and Boyle, 2008).

Marine sediment core TTR-451 (Fig. 1) was recovered from Eirik Drift, beneath the modern pathway of the East Greenland Current (EGC), which today constitutes the main mechanism for the export of cold and relatively fresh surface waters out of the Arctic (Aagaard and Carmack, 1989; Bacon et al., 2002; Holliday et al., 2009). Eirik Drift is a contourite formed from deposition of suspended sediment in Proto North Atlantic Deep Water (Proto-NADW) as it rounds the southern tip of Greenland (Chough and Hesse, 1983; Hunter et al., 2007a; b). Sediments from core TTR-451 therefore record changes in the intensity of Proto-NADW flow.

Here, we investigate co-registered records of stable oxygen isotope values from the calcite tests of the planktonic foraminiferal species Neogloboquadrina pachyderma (left-coiling), IRD/lithic counts of grains (>150 μm) per gram of dried sediment, numbers of planktonic foraminifera (>150 μm) per gram of dried sediment, and the ratio of susceptibility of anhysteretic remanent magnetization (kARM) to low-field magnetic susceptibility (κ) for core TTR-451. This combination of analyses enables investigation into changes in both surface and deep-water currents that exited the Nordic Seas during H1. We combine our results from Eirik Drift with proxy records from around the North Atlantic and Nordic Seas for the time interval that spans H1, from marine sediment cores SU90-09 (43°05′N, 31°05′W, 3375 m water depth), MD95-2010 (66°41′N, 10°W, 3085 m water depth), and TTR-451 (1927 m water depth).
04° 34′E, 1226 m water depth), and GGC5 (33° 42′N, 57° 35′W, 4550 m water depth). Records from these cores were published previously by Grousset et al. (2001), Dokken and Jansen (1999), and McManus et al. (2004), respectively, and their locations are shown in Fig. 1. We also present a synthesis of the timings of global ice sheet and glacier advances and retreats, and we compare them with sea-level and circum-North Atlantic terrestrial temperature proxy records through the Last Glacial Maximum (LGM), H1 and the Bølling warming. We consider H1 in the wider context rather than just the changes associated with the IRD event, as the transient, process point of view is important in understanding the drivers behind the climate anomalies. These events are then placed in the context of North Atlantic surface and deep-water hydrographic changes. We provide a review on various aspects of the North Atlantic hydrography during H1, and we aim to elucidate the evolution of H1 in the northern North Atlantic, with specific attention to the Nordic Seas, and to reconcile the sequence of events associated with H1 in the North Atlantic to the entire period of AMOC collapse.

2. Methods

κARM/κ ratios for core TTR-451 were determined from discrete samples and were cross-validated by comparing additional u-channel κARM measurements (not shown) with higher-resolution whole core magnetic susceptibility data (Fig. 2b). The whole core κ was measured with a Bartington Instruments MS2E1 point sensor that was placed in contact with the surface of a split core (labelled as ‘whole core’ in Fig. 2b). The discrete sample κ measurements were made using a Kappabridge KLY-4 magnetic susceptibility meter, and κARM was imparted using a 50 μT bias field and an alternating field of 100 mT, with measurements made using a 2-G Enterprises cryogenic magnetometer in a magnetically shielded laboratory.

Core TTR-451 was sub-sampled continuously at 0.5-cm intervals from a previously collected u-channel sample (2 cm cross-section). The sediment was prepared by freeze drying, weighing, and wet sieving at 63 μm. It was then oven dried, weighed and dry sieved, and the total weight of the > 150 μm sediment fraction was recorded. The sediment was then split into aliquots: one for stable
oxygen isotope analyses and the other for lithic counts. Stable oxygen isotope analyses for core TTR-451 were carried out on 21–32 individuals of *N. pachyderma* (left-coiling) ($\delta^{18}O_{op}$) with sizes ranging between 225 and 275 $\mu$m, using a Europa Geo-2020 mass spectrometer with an individual acid dosing preparation system. Stable isotope ratios are expressed as $\delta$ (delta) value, and are reported as per mil ($\permil$) relative to the Vienna Pee Dee Belemnite standard (VPDB). External reproducibility as determined using more than 100 blind standard analyses is better than 0.027$\permil$ for $\delta^{13}C$ and 0.053$\permil$ for $\delta^{18}O$. Lithic counts were performed on the 150 $\mu$m fraction of the allocated aliquot using a binocular microscope. On average, ~500 grains were counted per sample, and the grain types were petrologically identified.

3. Results

3.1. Eirik Drift core TTR-451

The chronology of core TTR-451 is primarily constrained by seven accelerator mass spectrometer (AMS) $^{14}$C datings of monospecific left-coiling *N. pachyderma* samples that have been calibrated using Calib 6.0.1 (Reimer et al., 2009) with a reservoir age correction $\Delta R = 0$ (Table 1). Within these constraints, the age model is fine-tuned by correlation between the TTR-451 magnetic susceptibility ($k$) record and the GRIP ice core stable oxygen isotope ($\delta^{18}O$) record (see correlation lines in Fig. 2). A similar approach has been used in several other studies (e.g., Kissel et al., 1999; Elliot et al., 2002). For further details and discussion, see Stanford et al. (2006). We extend the correlation of TTR-451 and Greenland ice core $\delta^{18}O$ records on the GICC05 timescale (Rasmussen et al., 2006, 2008; Andersen et al., 2006; Svensson et al., 2006) down to the LGM. Low abundances of planktonic foraminifera prior to H1 and a consequent lack of carbonate for AMS $^{14}$C dating mean that the age model for this portion of TTR-451 relies entirely upon the correlation shown in Fig. 2. Ice core $\delta^{18}O$ records reflect temperature and air mass variations over the ice sheet (e.g., Dahl-Jensen et al., 1998; Severinghaus et al., 1998; Severinghaus and Brook, 1999; Stuiver and Grootes, 2000). More negative $\delta^{18}O$ generally represent colder conditions, and less negative values represent warmer conditions. The GRIP and GISP$2$ ice core $\delta^{18}O$ records have significantly different values for the LGM/H1 transition (Fig. 2c, d) despite their close proximity (Fig. 1). Therefore, we consider a mean composite Greenland ice core $\delta^{18}O$ (temperature) record in units standard deviation (Fig. 2e) for the purpose of our correlation.

Stanford et al. (2006) estimated radiocarbon reservoir age correction ($\Delta R$) values using data from core TTR-451. The inferred $\Delta R$ for the Younger Dryas ($\Delta R = 351$ yrs, Table 1) compares well (within 1$\sigma$) with independently derived $\Delta R$ estimates of 371 yrs from cores from the Norwegian margin (Bonnevik et al., 2006) and of around 300 yrs by Bard et al. (1994). However, we note that both the AMOC configuration as well as CO$_2$ transfer may have been very different between H1 and the Younger Dryas. The GICC05 chronology for the GRIP and GISP$2$ $\delta^{18}O$ records has since been extended beyond 14.7 ka BP (Rasmussen et al., 2006, 2008; Andersen et al., 2006; Svensson et al., 2006). Thus, we provide revised $\Delta R$ estimates for H1 (Table 1). A $\Delta R$ of 1458 (1912–1081) yrs is inferred for TTR-451 (sample KIA-27859). These $\Delta R$ estimates are slightly lower than the ~2000 yrs suggested by Waelbroeck et al. (2001), suggesting that there likely was considerable spatial variability in $\Delta R$ during H1.

The sediment accumulation rate in core TTR-451 is found to have been considerably reduced during H1 (Fig. 3a). Relative changes in the abundances of IRD and planktonic foraminifera might have been caused by changes in sediment dilution, so we convert these records for TTR-451 from grains per gram of dried sediment into fluxes (number of grains per cm$^2$ per year), based on mass accumulation rate (Fig. 3b, c), to permit comparison with other cores (notably SU90-09).

A $\text{KAMS/kry}$ record for core TTR-451 (Stanford et al., 2006) is plotted alongside the $2^{13}Pb/2^{36}Th$ record for core GGC5 from the Bermuda Rise (McManus et al., 2004) in Fig. 4g. The record of $\text{KAMS/kry}$ (black line in Fig. 4g) gradually increases from ~19 ka BP (note the inverted axis), indicating a slowdown of AMOC intensity, and culminating in a collapse at 17.5 ka BP. A sharp decrease in values at ~14.6 ka BP suggests an abrupt AMOC resumption at the Bølling warming (Fig. 4a, g).

From ~5.5 to 17 ka BP, strong shifts to lighter $\delta^{18}O$ at magnitudes of ~15$\permil$ develop in core TTR-451 (Fig. 4e), which indicates increased freshening that culminates in a broad peak at ~15.1 ka BP. The IRD flux for TTR-451 (Fig. 4d) does not notably increase during this time interval. Planktonic foraminiferal accumulation flux in TTR-451 (Fig. 4c) is generally low during the H1 interval, with a notable minimum between 17.2 and 16.5 ka BP. After the Bølling warm transition at ~14.6 ka BP, $\delta^{18}O_{op}$ increased to a maximum at ~13.2 ka, followed by decreasing $\delta^{18}O$ into the YD cold period (Fig. 4e). Lithic and planktonic foraminiferal fluxes dramatically increase at ~14.6 ka BP (Fig. 4c, d) and remain variable until the top of core TTR-451 (at about the termination of the Younger Dryas).

3.2. Re-evaluation of terrestrial records of ice sheet and glacier extent and temperature for the LGM, H1 and Bølling warming

We used Calib6.01 (Reimer et al., 2009) to re-calibrate AMS $^{14}$C radiocarbon convention ages for terrestrial temperature proxy records (Atkinson et al., 1987; Alm, 1993), terminal moraines, and for marine sediment horizon (Graudt and Frezzotti, 1997; Benson et al., 1998; McCabe and Clark, 1998; Bowen et al., 2002; Dyke et al., 2002; Marks, 2002; Ivy-Ochs et al., 2006). These re-calibrated datings are used along with previously published calendar ages (e.g., McCabe et al., 2005, 2007) and 3Cl boulder exposure ages (e.g., Bowen et al., 2002; Dyke et al., 2002; Licciardi et al., 2004; Rinterknecht et al., 2006). Results are presented in Fig. 5f and are compared with global sea-level history (Fig. 5d, e), climate

<table>
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<tr>
<th>KIA sample number</th>
<th>Depth (cm)</th>
<th>Species</th>
<th>Radiocarbon conventional age</th>
<th>Calibrated age (Calib 6.0.1)</th>
<th>$2\sigma$ age range (Yrs BP)</th>
<th>Inferred $\Delta R$ (Yrs)</th>
<th>$2\sigma$ $\Delta R$ range (Yrs)</th>
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<tr>
<td>KIA-265988</td>
<td>12.00</td>
<td><em>N. pachy. (lc)</em></td>
<td>10905 ± 60 BP</td>
<td>12450 BP</td>
<td>12597–12208 (1)</td>
<td>351</td>
<td>537–85</td>
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<tr>
<td>KIA-27856</td>
<td>40.00</td>
<td><em>N. pachy. (lc)</em></td>
<td>12220 ± 55 BP</td>
<td>13680 BP</td>
<td>13810–13463 (1)</td>
<td>1164</td>
<td>1806–551</td>
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<tr>
<td>KIA-25853</td>
<td>57.50</td>
<td><em>N. pachy. (lc)</em></td>
<td>12690 ± 55 BP</td>
<td>14130 BP</td>
<td>14593–13938 (0.95)</td>
<td>1806</td>
<td>1338–457</td>
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<tr>
<td>KIA-27857</td>
<td>76.00</td>
<td><em>N. pachy. (lc)</em></td>
<td>12825 ± 55 BP</td>
<td>14422 BP</td>
<td>14974–14126 (1)</td>
<td>14866</td>
<td>14712–505</td>
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<tr>
<td>KIA-27858</td>
<td>83.50</td>
<td><em>N. pachy. (lc)</em></td>
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<td>KIA-25854</td>
<td>102.25</td>
<td><em>N. pachy. (lc)</em></td>
<td>14890 ± 60 BP</td>
<td>17672 BP</td>
<td>17907–17239 (1)</td>
<td>972</td>
<td>1338–457</td>
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</table>

Inferred $\Delta R$ values are for AMS $^{14}$C datings for the Younger Dryas and Heinrich event 1. KIA = Kiel Institut für Altersbestimmungen. For the calibrated $2\sigma$ age ranges, the probabilities for those distributions of ages are given in the brackets.
variability in Greenland and Antarctica (Fig. 5a–c), and changes in Northern Hemisphere summer insolation (Fig. 5a). Our recalibration of earlier AMS$^{14}$C dates for terrestrial records reveals surprising results and our findings are summarised in Table 2. Note, however, that although it is tempting to reconstruct phase relationships from Fig. 5f, this is not warranted because dating uncertainties on boulder exposure ages can exceed 1000 yrs, and because of large differences in the spatial distribution of regional datasets.

4. Discussion

4.1. TTR-451 $k_{\text{ARM/k}}$

Our $k_{\text{ARM/k}}$ record for core TTR-451 (Stanford et al., 2006), and the $^{231}$Pa/$^{230}$Th record for core GCCS from the Bermuda Rise (McManus et al., 2004) (Fig. 4g) offer independent proxies for NADW flow intensity. The chronologies of both records are independent of each other (McManus et al., 2004; Stanford et al., 2006), and both records indicate an AMOC collapse initiating at $\sim$18.8 ka BP, also in agreement with Hall et al. (2006), and an abrupt AMOC resumption at the Bølling warming at $\sim$14.6 ka BP (Fig. 4a, g). The AMOC resumption in both records does not coincide with the ‘conventional’ termination of H1 (as recorded in sediments from the IRD belt) at $\sim$16 ka BP (e.g., Grousset et al., 2001; Hemming, 2004; Fig. 4e). Instead, it is coincident with the Bølling warming (McManus et al., 2004; Stanford et al., 2006; Fig. 4a, g). This timing for the AMOC flow intensity increase at $\sim$14.6 ka BP underlines suggestions that the AMOC “switch on” and the Bølling warming were intrinsically linked (McManus et al., 2004; Stanford et al., 2006).

4.2. Comparison of H1 records from Eirik Drift with the IRD belt and Nordic Seas

Numbers of planktonic foraminifera and lithic grains $>150\ \mu$m, and $\delta^{18}O_{\text{H2O}}$ are shown in Fig. 4c–e, respectively, for core SU90-09 from the central IRD belt (Grousset et al., 2001). Consonant with
previous studies from the IRD belt (e.g., Bond et al., 1992; Broecker, 1994; Cortijo et al., 1997; Hemming, 2004), core SU90-09 has a maximum increase in surface water dilution with isotopically light freshwater from 17.4 to 16.7 ka BP, including a $\delta^{18}O_{\text{Opl}}$ shift of $-0.9^{\circ}$ in just over 100 years at $\sim 17.3$ ka BP. This rapid shift followed more gradual freshening ($-0.7^{\circ}$ in 900 years), which started at $\sim 18.3$ ka BP. The number of lithic grains $g^{-1}$ (IRD concentration) gradually increases at around the same time as the surface water dilution, with a pronounced maximum associated with the maximum freshwater dilution signal focussed at $\sim 16.8$ ka BP. This shows a ‘typical’ Heinrich event signature with meltwater derived from a massive iceberg discharge (e.g., Hemming, 2004). The number of planktonic foraminifera $g^{-1}$ decreases from 8000 to a few hundred, with a minimum centred at the interval of maximum surface water dilution ($17.4-16.7$ ka BP), when $N$. pachyderma (left-coiling) became the dominant surface dweller (Grousset et al., 2001). Compared to core SU90-09, and to descriptions of Heinrich layers (e.g., Bond et al., 1992, 1999; Cortijo et al., 2000, 2001; review of Hemming, 2004), records of $\delta^{18}O_{\text{Opl}}$, IRD and planktonic foraminiferal fluxes from core TTR-451 (Eirik Drift) have distinctly different patterns and timings. Based on the generally low IRD flux during the main $\delta^{18}O_{\text{Opl}}$ shift (Fig. 4d, e), this change does not seem to be primarily derived from an iceberg event. Instead, we interpret this $\delta^{18}O_{\text{Opl}}$ shift during the H1 interval to have originated from iceberg-free freshened surface waters, similar to suggestions by van Kreveld et al. (2000).

On the basis of consistent $\Delta R$ estimates for the Younger Dryas from TTR-451 with those from the Norwegian margin (Bondevik et al., 2006), as well as the excellent agreement between the $\delta^{18}O_{\text{Opl}}$ records on a purely radiocarbon chronology (see Supplementary Content 1), we use inferred TTR-451 $\Delta R$ values for H1 to convert ages for core MD95-2010 onto the GICC05 chronology. By using these inferred $\Delta R$ values for MD95-2010, a sharp increase in magnetic susceptibility values (identified by Dokken and Jansen (1999) as the Bølling warming) shifts from 15.5 ka BP on the Calib6.01 chronology, to 14.7 ka BP on the $\Delta R$ corrected age-scale (see Supplementary Content 1), corroborating this method. Consonant with previous studies from the Nordic Seas (e.g., Duplessy et al., 1991; Rasmussen et al., 2002a, b; Rasmussen and Thomsen, 2004, 2008), the MD95–2010 $\delta^{18}O_{\text{Opl}}$ and Cassidulina teretes ($\delta^{13}C_{\text{C4}}$) records show an up to a $-2^{\circ}$, $\delta^{18}O_{\text{Opl}}$ excursion (Dokken and Jansen, 1999) that spans the later phase of H1 ($\sim 16.5-\sim 15$ ka BP) (Fig. 4f). Comparison of $\delta^{18}O_{\text{Opl}}$ records from Eirik Drift and the Nordic Seas (Fig. 4f) indicates that, apart from resolution differences, these records are nearly identical. This strongly indicates direct surface water communication between the Nordic Seas and Eirik Drift during H1. Unfortunately, abundances of benthic foraminifera are too low in TTR-451 to permit a comparable benthic $\delta^{18}O$ record from Eirik Drift.

Light $\delta^{18}O_{\text{Opl}}$ and $\delta^{18}O$ excursions, with nearly identical patterns, timing and magnitude to those in cores TTR-451 and MD95-2010, have been described for cores from the Nordic Seas, the northern North Atlantic (particularly the Faroe-Shetland Gateway), and around Eirik Drift (Vidal et al., 1998; Dokken and Jansen, 1999; Rasmussen et al., 2002a; Rasmussen and Thomsen, 2004; Lekens et al., 2005; Hillaire-Marcel and de Vernal, 2008; Meland et al., 2008). There are four hypotheses to explain this light $\delta^{18}O$ isotopic event. The first involves iceberg derived, low salinity meltwater pulses and halocline deepening (e.g., Hillaire-Marcel and Bilodeau, 2000; Rashid and Boyle, 2007). Based on the low H1 IRD flux in core TTR-451 from Eirik Drift, we consider this first hypothesis to be unlikely. The second hypothesis involves a reversed thermocline as a result of sea-surface capping with freshwater and sea-ice, and warming of the subsurface layer (e.g., Mignot et al., 2007; Peck et al., 2008). However, $\delta^{18}O$ measurements on $N$. pachyderma sub-populations (based on size) from core MD95-2024 (Labrador Sea) reveal a negative temperature gradient along the thermocline, which led Hillaire-Marcel and de Vernal (2008) to reject this second hypothesis. The third hypothesis involves the sinking of isotopically light brines as a result of intense sea ice formation (e.g., Vidal et al., 1998; Dokken and Jansen, 1999; Risebrobakken et al., 2003; Millo et al., 2006; Hillaire-Marcel and de Vernal, 2008; Meland et al., 2008). The fourth hypothesis suggests penetration of relatively warm (4–8 $^\circ$C) waters into the Nordic Seas at intermediate depths (<1700 m), which is suggested to represent the North Atlantic Drift (NAD) that flowed below the ocean surface into the Nordic Seas (Rasmussen and Thomsen, 2004). We use our new records from Eirik Drift to evaluate these latter two hypotheses.

4.3. $\delta^{18}O$ excursions during H1 in the Nordic Seas and at Eirik Drift

We first consider the sea-ice hypothesis for generating the light $\delta^{18}O$ excursion in planktonic and benthic foraminifera in the Nordic Seas and in planktonic foraminifera at Eirik Drift during H1. LGM reconstructions for the northern North Atlantic suggest that seasonal sea-ice likely extended down to $\sim 40^\circ$N (Mix et al., 2001; Pflaumann et al., 2003). However, a regional $\delta^{18}O_{\text{Opl}}$ of $\sim 4.5^{\circ}$ indicates that the Nordic Seas and northern North Atlantic likely were sea-ice free in summer (e.g., Weiselt et al., 1996, 2003; de Vernal et al., 2002; Meland et al., 2005; Millo et al., 2006). High sedimentation rates, high planktonic foraminiferal fluxes and relatively heavy $\delta^{18}O_{\text{Opl}}$ records from western Fram Strait suggest that seasonally ‘open’ conditions even extended to this region during the LGM, unlike the permanently sea-ice covered central Arctic Ocean region, where both sedimentation rates and planktonic foraminiferal abundances were considerably lower (Nørgaard-Pederson et al., 2003). Because of the relatively high numbers of planktonic foraminifera per gram, despite the high sedimentation rate in core MD95-2010, we suggest that seasonally open conditions likely prevailed throughout the time period of H1in the Nordic Seas.

Dokken and Jansen (1999) proposed that freshwater additions into the North Atlantic and AMOC weakening would have resulted in high-latitude cooling and rapid extension of the sea-ice margin across the Nordic Seas during H1. Sea-ice forms with little oxygen isotope fractionation from freezing surface waters, which at this time were affected by freshwater dilution, so that the rejected
Fig. 5. a. The EDML $\delta^{18}O$ record (black) on the GICC05 time scale (EPICA Community Members, 2006) and the northern hemisphere summer (July) insolation (grey) at 65°N (Berger, 1991, 1999). b. The GRIP + GISP2 composite/mean $\delta^{18}O$ ice core record in unit standard deviation on the GICC05 timescale. c. The concentration of Ca$^{2+}$ (p.p.b.) in the Greenland ice cores. GRIP is shown in black, GISP2 in dark grey and NGRIP in light grey (Fuhrer et al., 1993; Mayewski et al., 1994; Bigler, 2004; Rasmussen et al., 2006, 2008). d. CFRS is the combined far-field relative sea-level record and in panel (e) the DFRS record shows the first derivative of the sea-level change (rate of sea-level change). These records were constructed using a Monte Carlo statistical analysis after Stanford et al. (in press). The 67%, 95%, and 99% probability intervals for the deglacial sea-level history are shown in blue.
brines would have advected a relatively low $\delta^{18}O$ signal to deeper waters. Dokken and Jansen (1999) and Meland et al. (2008) suggested that brine rejection would result in convection of ‘poled’ isopatically light surface waters to intermediate depths (to the base of the halocline), thereby accounting for the light H1 $\delta^{18}O$ signal in both surface waters and at intermediate depths. On the basis of similar $\delta^{18}O$ trends, Dokken and Jansen (1999) extended this hypothesis to account for $\delta^{18}O$ distributions in the Nordic Seas during other Heinrich events (Fig. 6). However, comparison of $\delta^{18}O_{\text{opt}}$ and $\delta^{18}O_{\text{pl}}$ data from the Nordic Seas for Heinrich events 4 and 6 (Fig. 6) indicates a larger $\delta^{18}O$ shift to lighter values at intermediate depths (benthics) than for surface waters (planktonics). For Heinrich event 6 (H6), this difference in $\delta^{18}O$ shift is approximately double. Furthermore, $\delta^{18}O_{\text{opt}}$ shifts to lighter values some 400 yrs prior to $\delta^{18}O_{\text{pl}}$. This timing offset is even more pronounced for H3 in core ENAM93–21 from the southern Nordic Seas (Dokken and Jansen, 1999). Because sea-ice formation and resultant brine rejection mix the light $\delta^{18}O$ signal in the halocline from the surface down through the water column, greater $\delta^{18}O$ would be expected in $\delta^{18}O_{\text{opt}}$ than in $\delta^{18}O_{\text{pl}}$. Also, if there were any temporal offsets, then $\delta^{18}O_{\text{opt}}$ would shift to lighter values before the $\delta^{18}O_{\text{pl}}$. The observations (Fig. 6) are therefore not consistent with the mechanism invoked by Dokken and Jansen (1999). Furthermore, core TTR-451 records nearly identical absolute $\delta^{18}O_{\text{opt}}$ as core MD95–2010 (Fig. 4f). Therefore, nearly instantaneous and rapid growth of the seasonal sea-ice margin would be needed across the entire Nordic Seas and northern North Atlantic to account for these observations. Such rapid and instantaneous sea-ice growth across such a vast area may have occurred (e.g., Gildor and Tziperman, 2003). However, open ocean convection would have likely resulted in dilution of the $\delta^{18}O$ signal. In order to generate brines dense enough to sink to intermediate depths and with such a low $\delta^{18}O$ signal, seasonal sea-ice is required to have formed in shallow shelf areas, reducing the $\delta^{18}O$ dilution because of limited mixing depths, and the Barents Sea Shelf may have provided such a location (Bauch and Bauch, 2001). However, Bauch and Bauch (2001) considered this scenario to be unlikely since the rates of sea ice formation would have been unrealistically high, and furthermore, the GLAMAP LGM reconstruction (Pflaumann et al., 2003) suggests that the Barents Shelf region was permanently sea-ice covered. A recent study by Rasmussen and Thomsen (2010) in Storfjorden, Svalbard, in the Barents Sea showed that, at present, brines that can attain a high enough density to sink to intermediate depths are those formed from cold, saline waters, not freshwaters. Furthermore, these brines have high $\delta^{18}O$ and $\delta^{13}C$ values, as opposed to the strongly light and low values recorded during H1. On the basis of these arguments, we reject the hypothesis that brine rejection from sea-ice formation caused the transfer of the light $\delta^{18}O_{\text{opt}}$ excursion to $\delta^{18}O_{\text{pl}}$ in the Nordic Seas. We now consider the fourth hypothesis that has been put forward to account for the observed light $\delta^{18}O$ isotopic excursions in the Nordic Seas. Rasmussen et al. (1996a, b) and Rasmussen and Thomsen (2004) found that in a number of cores (including ENAM93–21) from the Nordic Seas and the Faroe-Shetland Gateway, there is a distinct abundance increase in the benthic foraminiferal ‘Atlantic species’ group (e.g., Sigmaillospis schlumbergi, Egerella bradyi, Alabaminella weddellensis, Epistominella decorata, Bulimina costata, Sagrina subsinuosa, Cyrtolina spp.) at the same time as the H1 $\delta^{18}O_{\text{opt}}$ and $\delta^{18}O_{\text{pl}}$ light excursion in the Nordic Seas. Similar to Bauch and Bauch (2001), Rasmussen and Thomsen (2004) suggested that there was a weak subsurface invasion of relatively warm (4–8 °C) Atlantic waters (below 1 km) into the Nordic Seas during H1, which they interpreted as a subsurface expression of the NAD. Abundance differences of the ‘Atlantic species’ group have been used to infer that temperatures at intermediate water depths were greater in the Nordic Seas during the latter part of H1 than during Dansgaard-Oeschger interstadials (Rasmussen and Thomsen, 2004). Rasmussen and Thomsen (2004) suggested that the benthic and planktonic $\delta^{18}O$ shift to lighter values in the Nordic Seas during H events was due to this subsurface invasion of relatively warm water. Similar to Meland et al. (2008), we find the interpretation of Rasmussen and Thomsen (2004) problematic because relatively warm inflowing waters would have needed to become denser to enter the Nordic Seas at water depths greater than 1 km. Meland et al. (2008) also suggested that although model experiments indicate that deep waters may be warmed by a few °C (Weaver et al., 1993; Winton, 1997; Paul and Schulz, 2002), they do not account for the shallow Iceland-Scotland Ridge that these waters would need to cross. Moreover, Meland et al. (2008) noted that benthic foraminiferal Mg/Ca for core MD95–2010 do not indicate increased temperatures at intermediate depths in the Nordic Seas during this time period (Dokken and Clark, unpublished data). Furthermore, relatively large temporal offsets are apparent for nearly all Nordic Sea cores between the peak ‘Atlantic species’ abundances and the $\delta^{18}O$ minima (Rasmussen and Thomsen, 2004). Our observation of identical absolute $\delta^{18}O_{\text{opt}}$ in cores TTR-451 and MD95–2010 strongly suggests that there was direct water mass communication between the Nordic Seas and Eirik Drift during H1. If the light $\delta^{18}O$ excursion in the Nordic Seas resulted from relative warming of waters, then it would be unlikely to be expressed nearly identically outside the enclosed Nordic Seas at Eirik Drift. Finally, it is widely accepted that both the Denmark Strait and Iceland-Scotland Ridge overflowed had ceased (McManus et al., 2004; Hall et al., 2006; Stanford et al., 2006). Therefore, to accommodate inflow, waters are needed to have out-flowed at the surface from the Nordic Seas. Given the large isotopic shift observed across the Nordic Seas, a large body of inflowing waters would have been needed, and we question whether an equivalent volume of out-flowing surface waters would have been likely. Given that the Nordic Seas deep overflows had more or less collapsed from ~17.5 ka BP (Fig. 4g), and that the Bering Strait (at a present water depth of ~50 m) was closed due to low sea level (~110 m), any exchange between the Nordic Seas and North Atlantic, to maintain mass balance in the Nordic Seas, would have been via surface waters (e.g., via the Denmark Strait by means of the EGC). Hence, we suggest that $\delta^{18}O_{\text{opt}}$ signals observed at Eirik Drift resulted from net freshwater (diluted surface water) export from the Nordic Seas, in a configuration reminiscent of modern Arctic outflow through the Fram Strait. This would explain the close signal similarity of absolute values between $\delta^{18}O_{\text{opt}}$ in the Nordic Seas and over Eirik Drift. We conclude that neither the sea-ice nor the subsurface warming hypotheses (Dokken and Jansen, 1999; Rasmussen and
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<th>Paleo-climate description</th>
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<td>~ 26 – 21 ka BP</td>
<td>Ice-sheets at maximum extent</td>
<td>Barbados, Sunda Shelf and Boneparte Gulf sea-level records</td>
<td>Fairbanks (1989), Bard (1990a; b), Hanebuth et al. (2000; 2009), Yokoyama et al. (2000), Lambeck et al. (2002), Fairbanks et al. (2005), Peltier and Fairbanks (2006), Stanford et al. (in press), Fig. 5d, e Bowen et al. (2002), Dyke et al. (2002), Rinterknecht et al. (2006), Fig. 5f</td>
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<td>(LGM)</td>
<td>Relative cooling in Greenland.</td>
<td>GRIP and GISP2 δ¹⁸O ice core records</td>
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<td>Relative warmth in Britain and Scandinavia</td>
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<td>Warming in Greenland</td>
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<td>Southern hemisphere warming</td>
<td>Boulder exposure ages and datings on terminal moraines</td>
<td>Arz et al. (1999), Sachs et al. (2001), Kim et al. (2002)</td>
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<td></td>
<td>British and Laurentide ice sheets underwent the first retreat from their LGM position</td>
<td>Increased branched and isoprenoid tetaether (BIT)-index in the Bay of Biscay at ~ 19.5 ka BP supports the dating of the British ice-sheet retreat.</td>
<td>Bowen et al. (2002), Dyke et al. (2002) and references therein, Fig. 5f</td>
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<td></td>
<td>Glaciers in the Austrian Alps, Apennines, and western North America also started to retreat.</td>
<td>GRIP and GISP2 δ¹³C ice core records</td>
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<td>Rapid cooling in the British Isles</td>
<td>Pollen data</td>
<td>Atkinson et al. (1987)</td>
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<td>~19 – 17.5 ka BP</td>
<td>From around 19 – 18 ka BP: Brief warming in Scandinavia</td>
<td>Disappearance of Coleoptera</td>
<td>Atkinson et al. (1987)</td>
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<td></td>
<td>From 19 ka BP: The Scandinavian ice-sheet undergoes a sustained retreat.</td>
<td>Pollen records</td>
<td>Marks (2002), Rinterknecht et al. (2006), Fig. 5f</td>
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<td></td>
<td>This coincides with the deposition of fine-grained sediments over a &gt; 1600 km² area of the Nordic Seas.</td>
<td>Sediment cores and seismic mapping</td>
<td>Bauch et al. (2001)</td>
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<td>Penecontemporaneous glacial retreats are also documented for the northwest American Cordilleran ice sheet, the southeast Appalachian margin and the Austrian Alps. The ~ 500 yr duration Eerie interstadial is also dated at ~ 18.8 ka BP in the southern Laurentide region.</td>
<td>Boulder exposure ages and datings on terminal moraines</td>
<td>Hjelstuen et al. (2004), Lekens et al. (2005)</td>
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<td>On a global scale, ice volume appears to have begun a first substantial decrease, as suggested by an initial rapid step in sea-level rise. However, between 19.5 and 18.5 ka BP, the British ice sheet re-advanced in response to cooling.</td>
<td>Sunda Shelf and Boneparte Gulf sea-level records</td>
<td>Hanebuth et al. (2000; 2009), Yokoyama et al. (2000), Stanford et al. (in press), Fig. 5d, e Bowen et al. (2002) McCabe and Clark (1998), McCabe et al. (2007) and references therein</td>
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<td>~19.3 ka BP to ~ 17 ka BP</td>
<td>Glacial advance in the Apennines</td>
<td>Lack of ³⁶Cl boulder exposure ages</td>
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<td>~18 ka BP</td>
<td>Initiation of glacial advance within the southern Laurentide region</td>
<td>AMS ¹⁴C datings of marine mud</td>
<td>Atkinson et al. (1987)</td>
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<td>~17.5 ka BP</td>
<td>From around 18.3 ka BP, the rate of sea-level rise decreased</td>
<td>Sediment cores and seismic mapping</td>
<td>Clague and James (2002), Dyke et al. (2002) and references therein, Ivy-Ochs et al. (2006), Fig. 5f</td>
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<td>~17.5 – 15.5 ka BP</td>
<td>From 17.5 ka BP: Rapid intensification of polar atmospheric circulation and/or a change in circulation patterns, occurred along with minor cooling in Greenland</td>
<td>Sharp increase in the Ca²⁺ ion concentration in the Greenland ice cores and decreased Greenland ice core δ¹⁸O</td>
<td>Fairbanks (1989), Bard (1990a; b), Hanebuth et al. (2000; 2009), Yokoyama et al. (2000), Lambeck et al. (2002), Fairbanks et al. (2005), Peltier and Fairbanks (2006), Stanford et al. (in press), Fig. 5d, e Biscaye et al. (1997), Mayewski et al. (1997), Rohling et al. (2003), Andersen et al. (2006), Svensson et al. (2006), Rasmussen et al. (2006; 2008) and references therein, Jullien et al. (2006), Fig. 5a, b Giraudi and Frezzotti (1997), McCabe and Clark (1998), Denton et al. (1999), Dyke et al. (2002) and references therein, Bowen et al. (2002) Clague and James (2002) Liciardi et al. (2004), Ivy-Ochs et al. (2006), Rinterknecht et al. (2006), McCabe et al. (2007), Fig. 5f</td>
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<td>~17.5 ka BP</td>
<td>From 17.5 ka BP: Ice sheet retreat in the British Isles, but it re-advance back near its LGM extent by 16.5 ka BP. In the Apennine region, a shorter duration re-advance occurred at around 15.3 ka BP</td>
<td>Re-appearance of Coleoptera. Species abundances indicate that temperatures were similar to those at the LGM (winter temperatures of around -25 °C)</td>
<td>Atkinson et al. (1987)</td>
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<td>From 17.5 ka BP: Significant glacial retreat occurred on a nearly global scale. Ice-free conditions in the Gulf of St. Lawrence by ~16.7 ka BP indicate the scale of this retreat at the southeastern Laurentide margin, and a drawdown of the ice centre around the Hudson Bay region is thought to reflect important reorganisation of ice-streams Contrary to the cooling trend in Greenland ice cores, slight amelioration in British climate at ~17.5 ka BP</td>
<td>Datings on terminal moraines</td>
<td>McCabe and Clark (1998), McCabe et al. (2007), Ivy-Ochs et al. (2006), Giraudi and Frezzotti (1997)</td>
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<tr>
<td>~17.5 ka BP</td>
<td>Ice sheet retreat in the British Isles, but it re-advance back near its LGM extent by 16.5 ka BP. In the Apennine region, a shorter duration re-advance occurred at around 15.3 ka BP</td>
<td>Gradual decrease in the Ca²⁺ ion concentration in the Greenland ice cores and light Greenland ice core δ¹⁸O</td>
<td>Biscaye et al. (1997), Mayewski et al. (1997), Rohling et al. (2003), Andersen et al. (2006), Svensson et al. (2006), Rasmussen et al. (2006; 2008) and references therein, Jullien et al. (2006), Fig. 5a, b</td>
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14.6 ka BP


The southern Scandinavian ice sheet, however, re-advanced at 14.6 ka BP, most likely due to a positive mass balance related to increased moisture availability in Scandinavia (e.g., Rinterknecht et al., 2006), Lekens et al. (2005) suggested that the sediments had a meltwater origin. Based on the presence of > 2 mm IRD and pristine fully articulated bivalves, Lekens et al. (2005) suggested that they were deposited hemipelagically, and classified them as ‘plumites’ (i.e., sediments deposited during surface freshwater events). Lekens et al. (2005) also tentatively suggested that the laminations, which comprise two units (dark fine-grained laminae and lighter coloured coarser-grained laminae), represent winter sea-ice cover and summer plume deposition, respectively. Based on the detailed sediment description provided by Lekens et al. (2005), and the volume of sediment that was delivered into the Nordic Seas over a relatively short time interval (over 1000 km$^3$ in ~2000 yrs), we question whether the sediments resulted from fall-out from surface waters. Instead, due to the relatively high sediment concentrations within the meltwater plumes, we suggest that it is more likely that the sediments entered the Nordic Seas below the sea surface (e.g., Fohrmann et al., 1998), and that they therefore represent low-velocity hyperpycnal deposits that resulted from glacial melting and outwash.

4.3.1. A new hypothesis for light $\delta^{18}$O excursions during H1 in the Nordic Seas and at Eirik Drift

Seismic mapping and studies of marine cores have revealed that sediment accumulation rates were high within the Nordic Seas during H1 (Hjelstuen et al., 2004; Sejrup et al., 2004; Lekens et al., 2005). Core MD99-2291 from the Voring Plateau, southeastern Norwegian Sea, contains laminated/partly laminated fine-grained clay and silt (individual laminae are between 1 mm and 150 mm thick), with generally low planktonic foraminiferal abundances and occasional IRD (Lekens et al., 2005). Covering an area of ~1600 km$^2$ on the Norwegian Sea floor, these post ~18.6 ka BP sediments have a volume of ~1000 km$^3$ and attain a thickness that can exceed 20 m (Hjelstuen et al., 2004; Sejrup et al., 2004; Lekens et al., 2005). However, since the Storegga Slide (ca 8.2 ka BP; Bugge et al., 1987; Bondedvik et al., 1997, 2003) removed much of this deposit, calculations of total sediment volume are problematic (Hjelstuen et al., 2004; Sejrup et al., 2004; Lekens et al., 2005).

$\delta^{18}$O data from these sediments have the same absolute values as those in MD95-2010 and TTR-451, and the termination of the laminated deposits at 15.1 ka BP is coincident with peak light $\delta^{18}$O$^{\text{HPL}}$ in the Nordic Seas (Lekens et al., 2005). Given the similar timing of emplacement of these sediments and rapid retreat of the Fennoscandian ice sheet (e.g., Rinterknecht et al., 2006), Lekens et al. (2005) suggested that the sediments were deposited in a meltwater plume. Based on the presence of > 2 mm IRD and pristine fully articulated bivalves, Lekens et al. (2005) suggested that they were deposited hemipelagically, and classified them as ‘plumites’ (i.e., sediments deposited during surface freshwater events). Lekens et al. (2005) also tentatively suggested that the laminations, which comprise two units (dark fine-grained laminae and lighter coloured coarser-grained laminae), represent winter sea-ice cover and summer plume deposition, respectively. Based on the detailed sediment description provided by Lekens et al. (2005), and the volume of sediment that was delivered into the Nordic Seas over a relatively short time interval (over 1000 km$^3$ in ~2000 yrs), we question whether the sediments resulted from fall-out from surface waters. Instead, due to the relatively high sediment concentrations within the meltwater plumes, we suggest that it is more likely that the sediments entered the Nordic Seas below the sea surface (e.g., Fohrmann et al., 1998), and that they therefore represent low-velocity hyperpycnal deposits that resulted from glacial melting and outwash.

A hyperpycnal flow is a negatively buoyant flow that travels along the basin floor because its density is greater than that of the ambient water mass into which it is injected, and it attains its density from the entrained sediment load (Mulder and Syvitski, 1995; Stow, 1996). This process means that terrestrial material can be transported directly into marine environments via turbulent flow, initially with a freshwater origin (Mulder et al., 2003). Hyperpycnal flows can carry sediments finer than medium sand over large distances (Mulder et al., 2003). Hyperpycnites (the deposits from hyperpycnal flows) can be generated today from ‘dirty’ rivers, flood events, dam breaks, or Jökulhaups. Lekens et al. (2005) rejected the possibility that these sediments represent subsurface turbidity current generated deposits based on the rhythmic nature of the sediments, a lack of erosional contacts, and the presence of fully articulated bivalves and IRD (Rashid et al., 2003). However, the low-velocity, quasi-steady regime of hyperpycnal flows from low-velocity floods may not have erosional contacts (Mulder et al., 2003), and we suggest that such processes do not preclude the occurrence of IRD or intact bivalves.
meltwater injection into the Nordic Seas, with sediments reflecting more proximal material.

Based on modern rivers that produce at least one hyperpycnal flow each year, Mulder and Syvitski (1995) estimated the critical threshold particle concentration required for a water mass to plunge subsurface. Calculated values range from ~38.9 kg m\(^{-3}\) for low latitude rivers to ~43.5 kg m\(^{-3}\) at high latitudes. Measured annual average suspended particle concentrations are, however, somewhat lower (e.g., 20.7 kg m\(^{-3}\) for the Rioni in Russia) as hyperpycnal flows occur during peak discharge. Mulder et al. (2003) suggested that rivers with initial suspended particle threshold concentrations as low as 5 kg m\(^{-3}\) can produce hyperpycnal flows during flood.

We test the hypothesis that meltwaters may have entered the Nordic Seas hyperpycnally during H1 by calculating average suspended particle concentrations, assuming that the sediments have a particle density of quartz, 2650 kg m\(^{-3}\) (Mulder and Syvitski, 1995). We calculate this for three scenarios, and consider mixing this sediment load with incrementally increasing freshwater volumes. Scenario 1 involves the minimum estimated volume of sediment deposited during H1 in the Nordic Seas (1000 km\(^3\); Hjelstuen et al., 2004; Sejrup et al., 2004; Lekens et al., 2005). Scenario 2 takes into account that the light \(\delta^{18}O\) anomaly in MD99-2291 only spans the upper 1.4 m of the 10.5-m-thick laminated section. We therefore reduce the 1000 km\(^3\) sediment volume by 87%. The Storegga Slide, which post-dated H1 by nearly 8000 years, removed a large proportion of the H1 deposits in the Norwegian Sea and therefore, the cited 1000 km\(^3\) volume for the H1 deposit is a minimum estimate (Hjelstuen et al., 2004; Sejrup et al., 2004; Lekens et al., 2005). Although unrealistic, in scenario 3 we calculate the maximum possible sediment load of the H1 meltwaters by including the sediment volume contained within the Storegga Slide (3500 km\(^3\); Bondevik et al. (2003)). Therefore, scenario 3 has a total sediment volume of 4500 km\(^3\). Results are given in Table 3, and mixing curves for scenarios 1 and 2 are shown in Fig. 7.

Assuming that the \(-1.45\%/\text{C}0_{\text{NS}}\) \(\delta^{18}O\) shift at Eirik Drift \((\delta_{\text{lep}})_{\text{Eirik}}\) represents a mixed signal out of the Nordic Seas, we use a glacial meltwater end member \(\delta^{18}O\) value of \(-35\%/\text{C}0_{\text{MW}}\) to calculate the volume of freshwater that likely entered the Nordic Seas during H1. A volume of 2145900 km\(^3\) \((V_{\text{NS}})\) is used for the Nordic Seas, as we assume that mixing was constrained to the Norwegian Sea, Iceland Sea, and only the southernmost sector of the Greenland Sea (Nørgaard-Pederson et al., 2003), and we subtract the volume associated with a -110 m sea-level change \((\delta_{\text{lep}})_{\text{EIRIK}}\) is solved from the equation below, where \(\delta_{\text{NS}}\) is the initial \(\delta^{18}O\) value of \(\delta^{18}O_{\text{lep}}\) in the Nordic Seas and at Eirik Drift and \(V_{\text{MW}}\) is the volume of...
meltwater. We calculate that $V_{MW}$ equals $89000 \text{ km}^3$, or the equivalent of 0.246 m of global sea-level rise.

We then consider a mixture of the derived volume of meltwater ($V_{MW}$) with sediment loads used in scenarios 1, 2 and 3 (Table 4, with mixing curves for scenarios 1 and 2 in Fig. 7), where the meltwater injection volumes are translated to equivalent global sea-level rise.

For scenario 1, a meltwater volume of $61009 \text{ km}^3$ is required for the flow to acquire the critical threshold particle concentration of $43.5 \text{ kg m}^{-3}$ (Table 3). Mixing $1000 \text{ km}^3$ of sediment with the calculated meltwater volume yields an average suspended particle concentration of nearly $30 \text{ kg m}^{-3}$ (Table 4), which is similar to the observed average suspended particle concentrations of modern high-latitude rivers that produce hyperpycnal flows (Mulder et al., 2003). For this scenario, meltwaters injected into the Nordic Seas during H1 could have done so hyperpycnally.

Scenario 2, with the 87% reduced sediment estimate, requires a meltwater volume of only $8303 \text{ km}^3$ for the sediment load to cause meltwater flow to exceed its critical particle concentration ($43.5 \text{ kg m}^{-3}$) (Table 3). This volume is an order of magnitude less than the calculated meltwater volume ($V_2$). Mixing the relatively small sediment volume for scenario 2 with the calculated meltwater volume gives an average suspended particle concentration of $3.98 \text{ kg m}^{-3}$, which is less than the required $>5 \text{ kg m}^{-3}$ reported by Mulder et al. (2003). However, for scenario 2, a minimum sediment load is mixed with a maximum meltwater volume, and is averaged over $\sim1000$ yrs; therefore, sedimentation rates may have been

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Table 3

<table>
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<th>Scenario</th>
<th>Volume of sediment ($\text{km}^3$)</th>
<th>Volume of freshwater ($\text{km}^3$)</th>
<th>Equivalent global sea-level rise (m)</th>
<th>$C_c$ ($\text{kg m}^{-3}$)</th>
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<tr>
<td>Scenario 1</td>
<td>1000</td>
<td>$61009 \text{ (C1)}$</td>
<td>0.169</td>
<td>43.5 ($A^1$)</td>
</tr>
<tr>
<td>Scenario 2</td>
<td>133</td>
<td>$8303 \text{ (C6)}$</td>
<td>0.023</td>
<td>43.5 ($A^1$)</td>
</tr>
<tr>
<td>Scenario 3</td>
<td>1000 + 3500 (Storegga)</td>
<td>$273999$</td>
<td>0.759</td>
<td>43.5</td>
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</table>

That is, for average suspended particle loads to exceed the critical particle concentration ($C_c$). $C_c = 43.5 \text{ kg m}^{-3}$ (Mulder and Syvitski, 1995; Mulder et al., 2003). In brackets are points denoted in Fig. 7.

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Fig. 7. Meltwater volume, in terms of its contribution to global sea-level rise, versus its average suspended particle concentration values ($C_{sw}$) for scenarios 1 (black) and 2 (blue). The scenarios are defined in the main text. $C_{sw}$ changes with incrementally increased meltwater for these two scenarios. $C_{sw}$ sediment particle density = 2650 kg m$^{-3}$. Points $A^{1-4}$ and $C^{1-4}$ are defined in Tables 2 and 3. $B^{1-4}$ intersect with the two curves.
considerably higher during times of peak discharge (i.e., the short duration seasonal melt).

Scenario 3, with the Storegga slide volume added to the volume of laminated sediments in the Nordic Seas, provides an estimate of the maximum average suspended particle concentration. To achieve a critical suspended particle concentration of 43.5 kg m\(^{-3}\), a freshwater volume of nearly 274000 km\(^3\) would be needed (Table 3), which far exceeds the available meltwater volume (V\(_{MW}\)) derived from the \(\delta^{18}\)O\(_{w}\) of \(~89000\) km\(^3\). When the sediment volume for scenario 3 (4500 km\(^3\)) is mixed with V\(_{MW}\), a Cs\(_{av}\) of 134 kg m\(^{-3}\) is obtained (Table 4). For scenario 3, meltwater would be injected hyperpycnally into the Nordic Seas. Considering all three scenarios, we conclude that the laminated Nordic Seas sediments deposited during H1 likely represent hyperpycnites, originating from Scandinavian ice sheet melt.

At \(~15.5\) ka BP, intense northern high-latitude cooling (Alm, 1993; Fig. 5f) caused re-advance of the Scandinavian ice sheet (Rintenknecht et al., 2006), \(\delta^{18}\)O records from the Nordic Seas and Eirik Drift returned to heavier values, and finely laminated silt and clay were no longer deposited in the Nordic Seas. The gradual returns to heavier \(\delta^{18}\)O\(_{w}\) in the Nordic Seas and at Eirik Drift indicate that freshened waters were being purged out of the Nordic Seas, possibly due to wind-driven circulation and/or to a density gradient between the Nordic Seas and North Atlantic (e.g., Millo et al., 2006). Only a few centuries after the meltwater purging began to decrease, the AMOC recovered sharply at the time of and possibly causing (McManus et al., 2004) the abrupt Bølling warming (Fig. 4f, g).

4.4. A new conceptual model for the Nordic Seas and northern North Atlantic during H1

We propose a new concept for the sequence of events associated with the H1 iceberg/meltwater perturbation in the North Atlantic. We discuss a succession of three phases during H1 from both the marine and terrestrial realms. Phase 1 (\(~19–17.5\) ka BP) represents the onset of AMOC collapse, while phase 2 (17.5–16.7 ka BP) spans the main H1 phase in the IRD belt. Phase 2 is defined as the ‘conventional’ Heinrich event sensu stricto (HE-ss) and represents the intense IRD deposition and freshening event in the IRD belt, at the time of the “Heinrich layers” (Heinrich, 1988; Bond et al., 1992, 1999; Bond and Lotti, 1995; Grousset et al., 2000, 2001; Scourse et al., 2000; Hemming et al., 2000, 2002; Knutz et al., 2001, 2007; Hemming and Hajdas, 2003; Hemming, 2004). Phase 3 (16.7–14.6 ka BP) covers H1 in the Nordic Seas and at Eirik Drift and the termination of H1 cooling and AMOC resumption at the Bølling warm transition. Contrary to previous suggestions (e.g., Dowdeswell et al., 1995; Elliot et al., 1998; Rohling et al., 2003; Hemming, 2004; Roche et al., 2004), the entire sequence of paleoceanographic changes associated with H1, as proposed here, extends over almost 4000 years. We call this longer duration for the Heinrich event sequence, which includes all three phases, Heinrich event sensu lato (HE-sl). HE-sl spans the entire period of collapsed Nordic Seas deep-water formation, from onset to weakening, nearly complete collapse, through to sharp resumption at the Bølling warming (McManus et al., 2004; Stanford et al., 2006).

### Table 4

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Volume of sediment (km(^3))</th>
<th>Volume of freshwater (km(^3))</th>
<th>Equivalent global sea-level rise (m)</th>
<th>Cs(_{av}) (kg m(^{-3}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1000</td>
<td>88806 (C(^{14}))</td>
<td>0.246</td>
<td>29.84 (A(^{4}))</td>
</tr>
<tr>
<td>2</td>
<td>133</td>
<td>88806 (C(^{14}))</td>
<td>0.246</td>
<td>3.98 (A(^{4}))</td>
</tr>
<tr>
<td>3</td>
<td>1000 + 3500 (Storegga)</td>
<td>88806</td>
<td>0.246</td>
<td>134.28</td>
</tr>
</tbody>
</table>

The points denoted in Fig. 7 are indicated with brackets.

4.4.1. Phase 1: onset of AMOC collapse (\(~19–17.5\) ka BP)

From \(~26\) to \(~21\) ka BP, most circum-North Atlantic ice sheets had reached their maximum (LGM) extents, as indicated by sea-level records (e.g., Fairbanks, 1989; Bard et al., 1990a, b; Hanebuth et al., 2000, 2009; Yokoyama et al., 2000; Lambeck et al., 2002; Peltier and Fairbanks, 2006). From \(~21\) ka BP, warming is inferred from Scandinavian pollen records (Alm, 1993) and Greenland ice core \(\delta^{18}\)O (Andersen et al., 2006; Rasmussen et al., 2006, 2008; Svensson et al., 2006), as also noted by Bauch et al. (2001). The AMOC gradually slowed from \(~19\) ka BP, culminating in a collapse from \(~17.5\) ka BP (McManus et al., 2004; Hall et al., 2006; Stanford et al., 2006; Fig. 4g), which coincided with the onset of HE-s1 in the IRD belt (e.g., Bond et al., 1992, 1999; Bard et al., 2000; Grousset et al., 2001; Hemming, 2004; Fig. 4c–e). The AMOC slowdown clearly predates the start of H1, as identified by peak IRD in the IRD belt, by more than 1000 years. Consequently, the widespread ice rafting cannot be invoked as the root cause of AMOC weakening from \(~19\) ka BP. Instead, precursory freshwater events are a more likely mechanism for the AMOC slowdown.

Precursory IRD events have been identified in North Atlantic cores from the IRD belt, up to 1500 yrs prior to the main ice rafting; provenance studies indicate significant contributions of sediment derived from the European, Icelandic and Scandinavian ice sheets (e.g., Bond et al., 1992, 1997; 1999; Bond and Lotti, 1995; Darby and Bischof and Darby, 1999; Grousset et al., 2000, 2001; Scourse et al., 2000; Knutz et al., 2001; Hemming et al., 2000, 2002; Hemming and Hajdas, 2003; Hemming, 2004; Jullien et al., 2006; Peck et al., 2006, 2007a, b, 2007; Walden et al., 2007). A (partly) European/Scandinavian origin has been validated by the contemporaneous increase in fluvial input to the northern Bay of Biscay (Menot et al., 2006). This has led to the suggestion that early surging from European ice sheets may have stimulated a reaction from the Laurentide ice sheet (e.g., Grousset et al., 2000, 2001), which was the ‘key player’ in the main Heinrich events (e.g., Marshall and Koutnik, 2006). Precursor IRD-rich layers have also been identified in the Arctic, with a possible source from the Canadian Arctic Archipelago (Darby et al., 1997, 2002; Stokes et al., 2005). This timing for the onset of northern high-latitude warming (\(~19\) ka BP) coincides with the start of a warming trend in the EPICA Dronning Maud Land (EDML) ice core (EPICA Community Members, 2006), and increased northern summer insolation (Berger, 1991). This warming due to insolation, and changes in greenhouse gas and albedo feedbacks (e.g., Bender et al., 1997; Alley and Clark, 1999) may have promoted reduction of the relatively ‘small’ European and Scandinavian ice sheets (e.g., Stocker and Wright, 1991; McCabe and Clark, 1998; Stocker, 2003). Warming may also have been promoted due to Southern Ocean meltwater perturbations that enhanced AMOC intensity (KoHr and Lohmann, 2007), or alternatively, due to increased heat flow between the Indian and Atlantic Oceans via enhanced ‘Agulhas leakage’ (Peeters et al., 2004). We suggest that warming-induced northern hemisphere meltwater perturbations at \(~19\) ka BP reduced the AMOC intensity.
In response to northern cooling associated with AMOC slowdown (Fig. 4g), a short re-advance of the British and southern Laurentide ice-sheets occurred from 18.2 ka BP (Bowen et al., 2002; Dyke et al., 2002; Fig. 5f).

4.4.2. Phase 2: ‘main’ HE-ss1 phase (17.5–16.7 ka BP)

The main HE-ss1 phase (~17.5–16.7 ka BP) was characterised by maximum freshening and intense IRD deposition in the IRD belt, with consequently reduced numbers of foraminifera (Bond et al., 1992, 1999; Bond and Lotti, 1995; Grousset et al., 2000, 2001; Scourse et al., 2000; Hemming et al., 2000, 2002; Knutz et al., 2001, 2007; Hemming and Hajdas, 2003; Hemming, 2004; Peck et al., 2006, 2008; Fig. 4c–e) and a sustained AMOC collapse from ~17.5 ka BP (e.g., McManus et al., 2004; Hall et al., 2006; Stanford et al., 2006, Fig. 4g). Sarnthein et al. (2007) suggest that at around this time, an estuarine like reverse flow initiated over the Denmark Strait, evidenced by a pronounced increase in benthic foraminiferal ventilation ages in the Icelandic Sea, which were also younger than the ventilation ages of surface waters recorded in the planktonic foraminifera at the same core site. However, the mechanism by which inflowing, old intermediate waters may have overcome the steep topographic gradient of the (relatively shallow) Denmark Strait is problematic. From around 17.5 ka BP, cooling is indicated in Britain (Atkinson et al., 1987), Scandinavia (Alm, 1993), and the North Atlantic (Bard et al., 2000), as well as in Greenland, along with a sharp increase in ice core Ca$^{2+}$ ion concentration, which indicates enhanced polar atmospheric circulation and/or a reorganisation.

Fig. 8. Schematic diagrams of the evolution of H1 in the Nordic Seas and North Atlantic. Black lines indicate surface water currents, and dashed black lines indicate possible reduced flow intensity of thermohaline surface currents. Solid grey lines represent bottom currents and dashed grey lines indicate when bottom current flow intensity was significantly decreased. In Fig. 8a, the surface currents are after Pflaumann et al. (2003). Solid blue arrows – inferred ice sheet surges. Note, we also include the inferred freshening from the Barents ice sheet and from the northwestern Icelandic ice sheet after Sarnthein et al. (2001). Solid green arrow – Nordic Seas/Eirik Drift surface water mass communication (as inferred from this study). Light blue shaded zone – sites of NADW formation. Hatched black zone – IRD belt (after Hemming, 2004). Dot/dashed line represents the summer sea-ice margin. Blue hashed area – subsurface freshening in the Nordic Seas. Brown shaded area – high rates of sediment accumulation (Lekens et al., 2005). Orange hashed area represents inferred increased surface water salinity (Schmidt et al., 2004).
of circulation patterns (Biscaye et al., 1997; Mayewski et al., 1997; Rohling et al., 2003; Andersen et al., 2006; Svensson et al., 2006; Jullien et al., 2006; Rasmussen et al., 2008). Coeval with the onset of the main H1 phase, at ~17.5 ka BP, widespread and significant glacial retreats occurred on a nearly global scale (e.g., Giraudi and Frezzotti, 1997; Denton et al., 1999; Dyke et al., 2002 and references therein; Bowen et al., 2002; Clague and James, 2002; Licciardi et al., 2004; Ivy-Ochs et al., 2006; Menot et al., 2006; Rinterknecht et al., 2006), and a draw-down of the ice centre around the Hudson Bay region led to re-organisation of northeastern Laurentide ice-streams (Dyke et al., 2002; Fig. 8b). Significantly increased sedimentation rates indicate that large volumes of freshwater were delivered into the Nordic Seas at this time (Hjelstuen et al., 2004; Sejrup et al., 2004; Lekens et al., 2005). We re-interpret these deposits as hyperpycnal, which formed due to settling from deep sediment-laden meltwater injections (Fig. 8b). Note also that freshwater dilution has been documented in sediment cores that would have been located in front of the Barents ice shelf (Sarnthein et al., 2001). Following previous work (e.g., Broecker, 1991, 1994; 2000; Rahmstorf, 1994; Manabe and Stouffer, 2000; Ganopolski and Rahmstorf, 2001; McManus et al., 2004; Rahmstorf et al., 2005), we suggest that significant freshwater perturbations associated with iceberg discharges into the North Atlantic (as inferred from light planktonic 18O isotope values and thick IRD layers in the IRD belt) (e.g., Hemming, 2004, Fig. 4d,e) likely kept the AMOC in a collapsed state (Fig. 4g), which caused significantly reduced northern hemisphere temperatures and possible changed patterns of northern hemisphere seasonality (Fig. 5a, g).

### 4.4.3. Phase 3: H1 ‘clean-up’ and AMOC resumption (16.7−14.6 ka BP)

The final phase of H1 started at 16.7 ka BP with rapid reduction in surface-water freshening and IRD deposition in the IRD belt (Fig. 5c, d, 8). However, sustained cooling is apparent in Greenland ice core δ18O records (Fig. 5a), and in Britain and Scandinavia (Fig. 5f), while Greenland Ca²⁺ ions suggest that atmospheric polar circulation remained relatively intense (Fig. 5c). The AMOC remained collapsed (Fig. 4g). The British ice sheet and Apennine glaciers re-advanced at around this time (Giraudi and Frezzotti, 1997; McCabe and Clark, 1998; McCabe et al., 2007). Unlike the decreased freshening inferred from IRD and δ18Oplm records from the IRD belt, there are no signs of decreased surface water freshening at Eirik Drift or in the Nordic Seas (Fig. 4e, f). On the contrary, surface water freshening in both benthic and planktonic records started to increase at ~16.7 ka BP (Fig. 4e, f). Given the low IRD flux at Eirik Drift (i.e., freshwater had a non-iceberg source (Fig. 4d)) and that δ18Oplm records have the same absolute values and variability as those of the Nordic Seas (Fig. 4f), we suggest that there was direct surface water communication between the Nordic Seas and Eirik Drift during this latter phase of H1.

Maximum surface freshening at Eirik Drift is suggested at ~15.1 ka BP (Fig. 4f). Thereafter, surface δ18Oplm progressively returned to heavier values, coincident with re-advance of the Scandinavian ice sheet (Marks, 2000; Rinterknecht et al., 2006; Fig. 5f) and termination of exceptionally rapid deposition of fine-grained sediments in the Nordic Seas (Lekens et al., 2005; bright green line in Fig. 5f). Above, we interpreted these fine-grained sediments as hyperpycnal flows that injected significant freshwater into intermediate depths of the Nordic Seas, followed by transfer (mixing) of the light δ18O signal from intermediate depths to the surface. As the AMOC was collapsed during the main phase of H1 (Fig. 4g), these relatively freshwaters would have ‘pooled’ in the Nordic Seas, which may explain why the AMOC remained in its collapsed state (Fig. 4g) despite the fact that iceberg-produced freshwater forcing in the open North Atlantic had ceased already (Fig. 4d and e). Mass balance dictates that these freshened Nordic Seas waters would have been expelled at the surface, notably through Denmark Strait, given that the Bering Strait was closed and that no deep waters exited from the Nordic Seas (Fig. 4g). Hence, δ18Oplm signals at Eirik Drift appear to record the largely iceberg-free meltwater admixture as it was purged from the Nordic Seas. Heavier δ18O after 15.1 ka BP suggests reduced freshwater admixture in the Nordic Seas, due to reduced meltwater input into the basin.

From ~15.5 ka BP, significant climate amelioration is recorded in the British Isles (Atkinson et al., 1987) and accelerating retreat from the H1 glacial maximum extent was well underway on a nearly global scale (Giraudi and Frezzotti, 1997; Benson et al., 1998; McCabe and Clark, 1998; Clague and James, 2002; Dyke et al., 2002 and references therein; Ivy-Ochs et al., 2006; McCabe et al., 2007; Hendy and Cosma, 2008). However, no warming occurred at northern high-latitudes (Alm, 1993; Rasmussen et al., 2006, 2008) or in well-dated lower latitude records (e.g., Hughen et al., 1996; von Grafenstein et al., 1999; Wang et al., 2001; Lea et al., 2003), which show excellent signal comparison with ice core 18O changes in Greenland (Rasmussen et al., 2006; Rohling et al., 2009), until the sharp Bølling (~15 °C) warming at 14.6 ka BP (Severingham and Brook, 1999; Rasmussen et al., 2006). Denton et al. (2005) suggested that records with close similarity to Greenland δ18O may be biased toward winter conditions, and speculated that improving summer conditions caused the apparent ‘mismatch’ with datings of snowline variations. Denton et al. (2005) also suggested that increased winter sea-ice cover at northern high latitudes may have caused increased seasonality, consonant with suggestions by Broecker (2001) and Seager and Battisti (2007).

Model investigations suggest that Southern Ocean perturbations may have caused the AMOC ‘switch on’ at the Bølling warming (e.g., Weaver et al., 2003; Knorr and Lohmann, 2003). Ocean–circulation model results indicate that low latitude North Atlantic salt retention during AMOC collapse may have preconditioned the AMOC to kick start at the Bølling warming by advecting more saline surface waters to areas of NADW formation (Knorr and Lohmann, 2007). This model outcome agrees with sea surface salinity estimates for the Caribbean during the final H1 phase (Schmidt et al., 2004) and with Mediterranean salinity input (Rogerson et al., 2004). Liu et al. (2009) use an ocean-climate model to suggest that a complete cessation of freshwater forcing alone may have switched the AMOC back on at the Bølling warming. However, such a complete cessation of North Atlantic freshening is not validated by the paleo-data (Fig. 5d, e).

We suggest that after 15.1 ka BP, termination of freshwater injection from the Scandinavian ice sheet into the Norwegian Sea (Lekens et al., 2005), and purging of accumulated freshwater out of the Nordic Seas, would have (additionally) allowed gradually increased salinity. A few centuries after the freshwater purging flux began to decrease, the AMOC recovered sharply (McManus et al., 2004; Stanford et al., 2006; Fig. 4g) at the abrupt Bølling warming (14.6 ka BP) (Fig. 4a). We therefore suggest that increasing the salinity of the Nordic Seas prepared this key region of NADW formation (e.g., Dickson and Brown, 1994; Bacon, 1998, 2002) for the sharp AMOC recovery (Fig. 8d).

### 5. Conclusions

We identify four key oceanographic changes during HE-s1, namely: (1) from 19−17.5 ka BP, a slowdown of NADW formation and small-scale precursor, iceberg-produced, meltwater events; (2) between 17.5 and 16.7 ka BP, a large-scale iceberg release and melting in the open North Atlantic IRD belt and a near shutdown of Nordic Seas deep-water formation (HE-s1); (3) pooling of
hypercannelly injected freshwater in the Nordic Seas, and sustained AMOC collapse between 16.7 and 15.1 ka BP, and (4) after 15.1 ka BP, a subsequent purging of pooled meltwater out of the Nordic Seas. Termination of this final meltwater ‘clean-up’ phase was nearly coincident with the AMOC recovery that accompanied Balling warming. The entire sequence of events extends the duration of the ‘wider Heinrich event 1 sequence’ (HE-s11) to almost 4000 years, rather than to several centuries as previously suggested (e.g., Dowdeswell et al., 1995; Elliott et al., 1998; Rohling et al., 2003; Hemming, 2004; Roche et al., 2004), which only represent HE-s1. This longer total duration of the H1 episode (HE-s11) now agrees with the entire previously established period of collapsed Nordic Seas deep-water formation (McManus et al., 2004; Stanford et al., 2006).

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Appendix. Supplementary data

Supplementary data associated with this article can be found in the on-line version, at doi:10.1016/j.quascirev.2011.02.003.

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