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Review paper

A review of the deep and surface currents around Eirik Drift, south of Greenland: Comparison of the past with the present

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ABSTRACT

The global Thermohaline Circulation (THC) is primarily driven by the cooling and sinking of northward flowing North Atlantic surface waters in the Nordic Seas to form North Atlantic Deep Water (NADW) that flows southward as a component of the Deep Western Boundary Current (DWBC). It is widely accepted that major freshwater injections have disrupted the formation of NADW in the past, causing widespread cooling over the North Atlantic. Eirik Drift, a contourite south of Greenland, was formed from deposition of sediments carried in the DWBC, so contains information about DWBC variability. Before now, the spatial and temporal variability of the surface and deep water currents, and their relationship with the associated sedimentation have not been fully understood. Here, we present a review of the key findings from the RAPID Cape Farewell project at Eirik Drift, a multi-disciplinary study which included hydrographic profiles, sub-bottom and sea-floor geophysical data, and multi-proxy analyses of a marine sediment core. We use these previously published results to further elucidate the oceanographic processes above Eirik Drift and relate these results to the sedimentation patterns. We also resolve, using a down-core record of NADW flow intensity, how bottom currents in this region changed in association with freshwater forcing during the last deglaciation.

Contents

1. Introduction

The oceanic Thermohaline Circulation (THC), or the 'Great Ocean Conveyor Belt' after Broecker (1991), globally redistributes heat, and is driven by a combination of wind stress, convection, and (tidal) mixing/turbulence, resulting in a process of high-latitude freshening, cooling and sinking of relatively saline surface waters to form intermediate and deep water masses, which subsequently upwell at low latitudes and in the Pacific Ocean (e.g. Broecker, 1991; Dickson and Brown, 1994; Wunsch, 2002). The Meridional Overturning Circulation (MOC) is a closely related concept, which refers to the north–south flow as a function of depth and latitude (Rahmstorf, 2003).

Relatively warm and saline surface waters are transported northeast across the North Atlantic, from the tropics to the Nordic Seas and Arctic, via the Gulf Stream and the North Atlantic Current (e.g. Schmitz and McCartney, 1993). Transporting up to 10^{15} W of heat (Ganachaud and Wunsch, 2000), the Gulf Stream/North Atlantic Current (NAC) maintains relatively high temperatures over Northwest Europe. At high latitudes within the northern North Atlantic, wind-driven surface cooling releases the heat carried within the NAC to the atmosphere, and drives deep convective mixing, whereby surface water masses become

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colder and denser, and sink to form intermediate and deep water (e.g. Broecker, 1991; Dickson and Brown, 1994; Bacon et al., 2002; Rudels et al., 2002; Wunsch, 2002). The deep waters spread out at depth, particularly in the Deep Western Boundary Current (DWBC), and include Denmark Strait Overflow Water (DSOW), Iceland–Scotland Ridge Overflow water (ISOW) and Labrador Sea Water (LSW) (e.g. Bacon, 1998; Yashayaev and Dickson, 2008; Fig. 1). Sediments carried in suspension by the DWBC are deposited at Eirik Drift, south of Greenland, a contourite formed as the DWBC slows as it rounds the southern tip of Greenland (e.g. Chough and Hesse, 1985; Hunter et al., 2007a,b).

Deep convective mixing occurs when the surface waters are sufficiently cooled so that their density becomes greater than the layers of water below. Gravity causes the dense water to descend and mix with the underlying waters. Convection is therefore more likely to be deep if the vertical density gradient is low (also referred to as low stratification). Density gradients are determined by salinity as well as temperature, and a body of water with a low vertical salinity gradient will be much easier to mix convectively than one that has a substantial layer of light, fresh water at the surface. It is for this reason that during periods of intense inflow of surface freshwater to the North Atlantic, deep convection may be inhibited, and the strength of the MOC reduced (e.g. Rahmstorf, 1995; Rahmstorf and Ganopolski, 1999; Manabe and Stouffer, 2000; Rahmstorf, 2002; Rahmstorf et al., 2005). In the event of a substantial reduction or even shut-down of the MOC, models suggest that the MOC eventually recovers with increasing salinity along a different pathway, forming a hysteresis loop (e.g. Stommel, 1961; Rahmstorf, 1995; Rahmstorf et al., 2005). Coupled climate models indicate that if the THC was 'switched off', then cooling would occur over the northern North Atlantic, with maximum temperature decreases in the Nordic Seas where sea-ice margin expansion and positive feedbacks exacerbate the cooling (Schiller et al., 1997; Rahmstorf, 2000; Seager et al., 2002; Rahmstorf, 2003). A weakening or shutdown of the THC/MOC due to freshwater forcing and cutting off the northern North Atlantic heat supply, has been widely speculated as the cause of past abrupt cold events observed in palaeo-records (e.g. Broecker, 1991; Stocker et al., 1992; Rahmstorf, 1995, 2000, 2002; Manabe and Stouffer, 1997; Schiller et al., 1997; Manabe and Stouffer, 2000; Ganopolski and Rahmstorf, 2001; Vellinga and Wood, 2002; Schmittner et al., 2003; Rahmstorf et al., 2005; Liu et al., 2009).

Observations of periodic surface freshening around the Arctic and the Nordic Seas (e.g. Lindsay and Zhang, 2005; Rignot and Kanagaratnam, 2006), dramatic rates of permanent and seasonal sea-ice loss (Kwok and Rothrock, 2009; Spreen et al., 2009; Cox et al., 2010) and inferences of reduced NADW formation (e.g. Bryden et al., 2005), underline the need

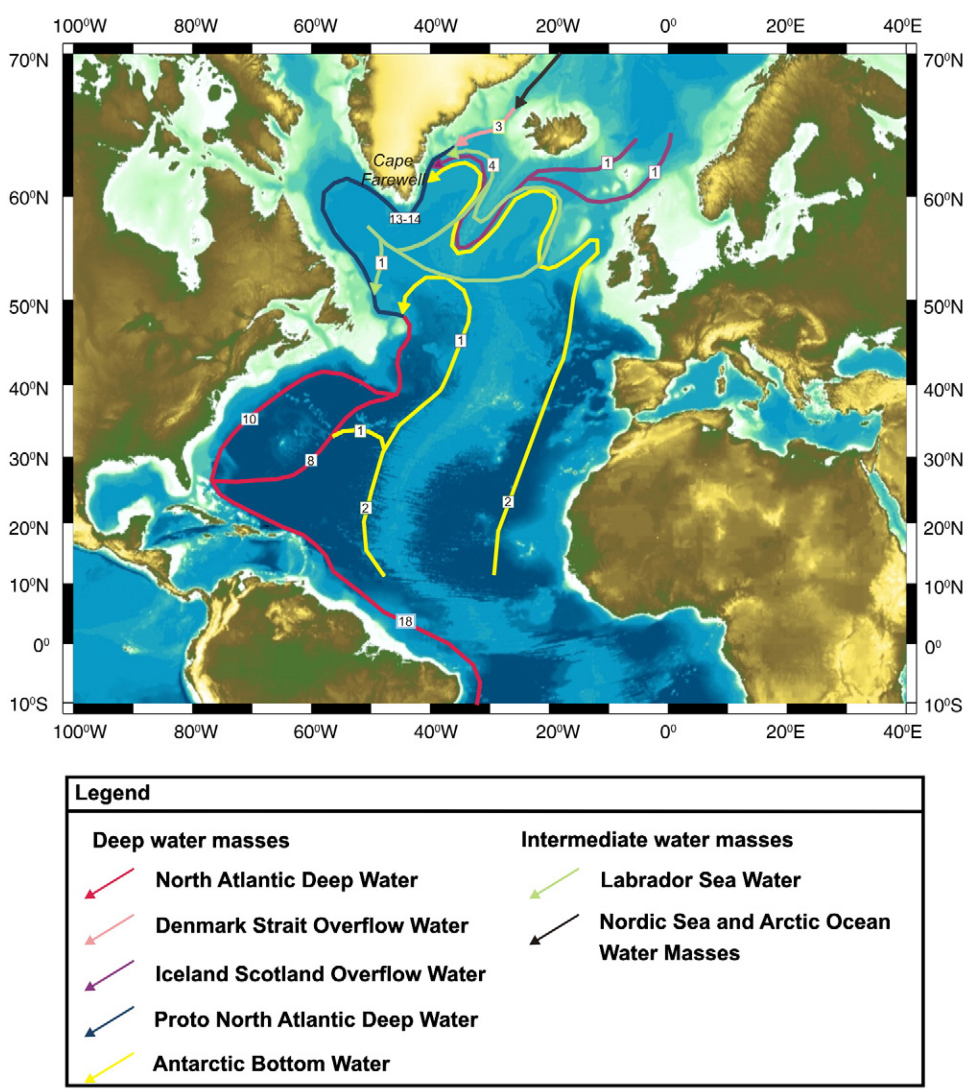

Fig. 1. Map of the North Atlantic region showing the water masses contributing to the formation of North Atlantic Deep Water (modified from Hunter et al., 2007a). Boxed numbers indicate the volume flux in Sverdrups.

to better understand the ocean/climate relationship in both the past and the present day context. This is crucial for the improvement of forecasting future ocean/climate changes, model validation, as well as impact assessment.

The Cape Farewell region, south of Greenland, is the optimum site for observing NADW/DWBC flow variability (Fig. 1), and rates of Arctic freshwater export carried in the East Greenland Current (EGC) and the East Greenland Coastal Current (EGCC) (Fig. 2) (e.g. Dickson and Brown, 1994; Bacon, 1998, 2002). The RAPID project at Cape Farewell, which was funded in 2004, addressed RAPID's objectives 2 (long-term direct observations of transports at critical locations in the North Atlantic) and 3 (construction of well-calibrated and time-resolved palaeo data records of past climate change). This was achieved through three inter-related studies from Eirik Drift and around the Cape Farewell region. The first quantifies the velocities, transport and properties of the modern surface and deep currents in the vicinity of Cape Farewell. The second aims to understand how these deep water changes, both past and present, have affected the sedimentation rates and patterns on Eirik Drift. The third assesses how the regional hydrography changed during past rapid climate transitions, with particular focus on the time period of the last deglaciation and variations in NADW flow intensity. This manuscript provides a review and synthesis of the key results previously published in Wilkinson and Bacon (2005), Stanford et al. (2006), Hunter et al. (2007a,b), Holliday et al. (2007, 2009), Lauderdale et al. (2008) and Bacon and Saunders (2010). Some of these results are also included in three PhD theses (Hunter, 2008; Stanford, 2008; Wilkinson, 2008). Additional work, closely tied to this project, describes the $\delta^{18}O$ composition of surface waters around Eirik Drift, and hence the freshwater contributions to the EGC and EGCC (Cox et al., 2010). However, these results lie outside of the scope of this paper.

2. Material and methods

In 2005, a hydrographic survey was combined with a series of mooring operations on *RRS Discovery* during cruise D298 (Bacon, 2006). Conductivity, temperature and depth (CTD) and Lowered Acoustic Doppler Current Profiler (LADCP) measurements were made at 63 stations in a series of transects in the Cape Farewell region (Fig. 3), using a SeaBird 911 plus CTD and RD Instruments 300 kHz Workhorse LADCP, respectively (Bacon, 2006). This hydrographic survey of the modern current system above Eirik Drift was combined with sub-bottom profiling and the recovery of piston cores, targeted on sediments which showed continuous parallel reflectors in sub-bottom profiles, in order to recover thick Holocene sedimentary sequences. Used in this study were D298-P1 (3452 m water depth; 57°30.227′N

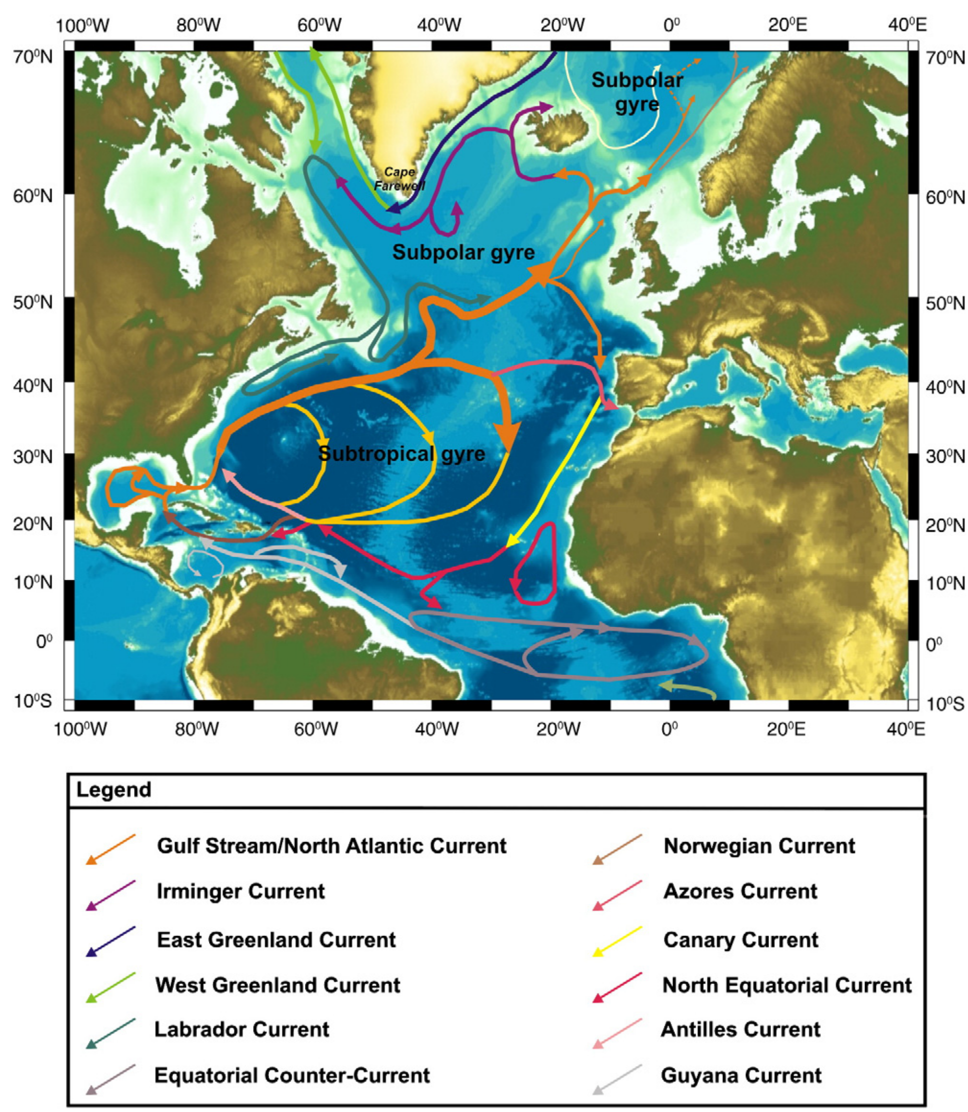

Fig. 2. Map of the North Atlantic region showing the location of surface water currents and surface water circulation (after Schmitz and McCartney, 1993).

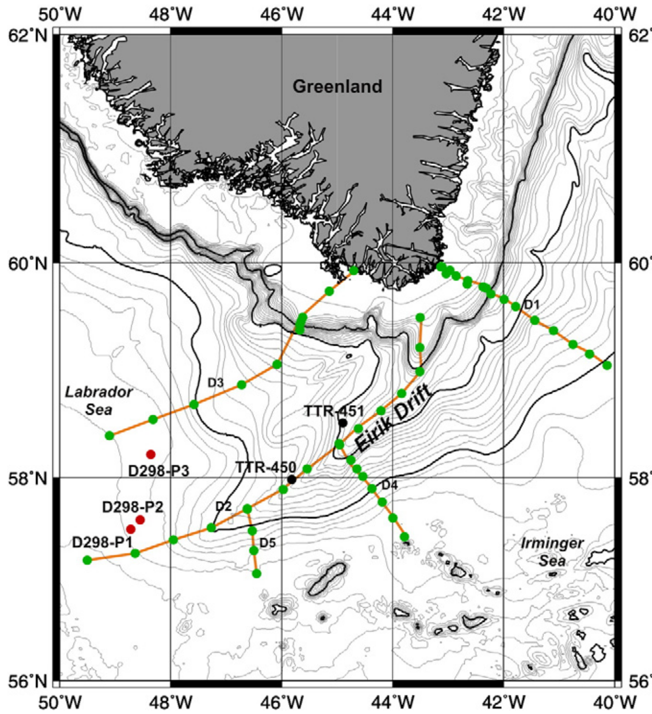

Fig. 3. Map showing the locations of marine sediment cores TTR-450, TTR-451, D298-P1 and D298-P3. Plotted in green are the CTD stations for the 2005, D298 cruise and the orange lines show the survey lines (Bacon, 2006). Bathymetry is from the GEBCO Digital Atlas published by the British Oceanographic Data Centre on behalf of the Intergovernmental Oceanographic Commission of UNESCO (IOC) and the International Hydrographic Organisation (IHO) (2003).

and 48°43.369′W) and D298-P3 (3430 m water depth; 58°13.025′N and 48°21.765′W) (Fig. 3). To increase the data coverage over Eirik Drift, we used additional geophysical data and sediment cores, which were collected during the TTR-13 cruise (2003) onboard the *RV Professor Logachev*. Cores used included TTR13-AT450G from 2326 m water depth and at 57°59.998′N, 45°49.002′W; and TTR-13-AT451G from 1927 m water depth and at 58°30.886′N, 44°54.333′W (hereafter referred to as TTR-450 and TTR-451, respectively) (Fig. 3). Data from a subsequent cruise on *RRS Charles Darwin*, CD-159 (2004), were also used, and the specific methods are presented in Hunter et al. (2007b). By study of core top sediments and geophysical images, the near present-day sedimentary regime could be directly related to the modern current transport system in the Cape Farewell and Eirik Drift region. For further details on the oceanographic material and data collection and processing, see Bacon (2006).

Stanford et al. (2006) determined the ratios of susceptibility of anhysteretic remanent magnetisation (κ_{ARM}) to low-field magnetic susceptibility (κ) κ_{ARM}/κ on discrete samples from core TTR-451. The κ measurements were made using a Kappabridge KLY-4 magnetic susceptibility metre, and the κ_{ARM} 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. The κ_{ARM}/κ ratio represents a measure of titanomagnetite grain size and is used as a proxy for NADW flow intensity (Stanford et al., 2006).

3. Results and discussion

By quantifying the present day current velocities and mapping their pathways, and in tandem mapping the contemporary surface sediments, we can test the hypothesis that surface sediments accurately reflect the intensity of the western boundary currents and by implication, the intensity of the THC. In this section we review the findings of the modern hydrographic survey and recent analysis of short-term variability (interannual to decadal), and link that information to the surface sediments. Finally the subsurface core analysis is reviewed to show NADW flow intensity in the past.

3.1. The East Greenland and the East Greenland Coastal Current hydrography and interannual to decadal variability

Located along the East Greenland margin, the southward flowing EGC and EGCC together form one of the main transport routes of freshwater out of the Arctic (Aagaard and Carmack, 1989; Bacon et al., 2002) and provide a transport route for Greenland glacial run-off and sea ice melt (Bacon et al., 2008; Sutherland and Pickart, 2008; Cox et al., 2010). The EGCC freshwater flux had been previously estimated at 0.06 Sv in 1997 (Bacon et al., 2002) with the current carrying ~30% of the net Arctic freshwater gain. Study of the total transport and short- and long-term variability of these currents is important in understanding not only how to monitor for future changes in high-latitude freshening of the oceans, but in the broader sense, to assess the impact of freshening upon the THC and climate.

Using historical hydrographic (salinity and temperature) data collected between 1932 and 1997, which included 147 stations and 40 sections in the Cape Farewell region, Wilkinson and Bacon (2005) calculated that the EGCC had a baroclinic transport of between 0.5 and 2 Sv. They found that both the long-term and latitudinal variabilities were masked by short-term fluctuations in transport. A weakening of the EGCC was observed between the 1930s and 1950s, which may have been related to the 1930s Greenland warming, identified by Box (2002). However, Wilkinson and Bacon (2005) suggested that significant short-term variability rendered the relationship statistically insignificant.

Wilkinson and Bacon (2005) also showed that the lateral spread of the EGCC is controlled by the shelf width (and hence, is related to latitude), although the current depth showed no clear latitudinal relationship. A linear relationship between salinity and ADCP velocity data (from Bacon et al., 2002) provided a potential proxy for total transport. Wilkinson and Bacon (2005) showed that, surprisingly, there was no increased total transport with decreased latitude, as one might have expected with an increased proportion of Greenland run-off. The freshwater component of the EGCC was found to exhibit a wide range, and showed large short-term variability. Therefore, they suggested that a previous spot measurement of 0.06 Sv (Bacon et al., 2002) represented a summer freshwater flux.

Using CTD and LADCP data from 2005, Holliday et al. (2007) analysed the velocity and water mass properties for three hydrographic sections; one in the Irminger Sea, one along Eirik Drift, and another in the Labrador Sea (D1–D3 in Fig. 3). The location of the section along Eirik Drift allowed, for the first time, the major currents to be observed as they flowed from the Irminger Sea to the Labrador Sea. Results showed that the waters in the EGC and EGCC merge directly south of Cape Farewell (Fig. 4), forming the West Greenland Current. Around one-third of the water transported by the shallow currents (5.1 ± 0.5 Sv of a total of 15.5 ± 1.0 Sv at Eirik Drift) diverges from the main boundary pathway and spreads southwards over Eirik Drift. From there it circulates into the interior of the subpolar gyre. This retroflection feature was also observed in a general ocean circulation model, but whether this it is a long term feature of the Cape Farewell circulation is not presently known.

3.2. DWBC hydrography and variability

Components of the DWBC include Denmark Strait Overflow Water (DSOW) at the bottom with high oxygen and low salinity, modified Iceland–Scotland Ridge Overflow water (ISOW) which is slightly more saline and less dense, and Labrador Sea Water (LSW) which is characterised by low salinity and low stratification (e.g. Bacon, 1998, 2002; Hunter et al., 2007a,b; Holliday et al, 2009). Note that the ISOW

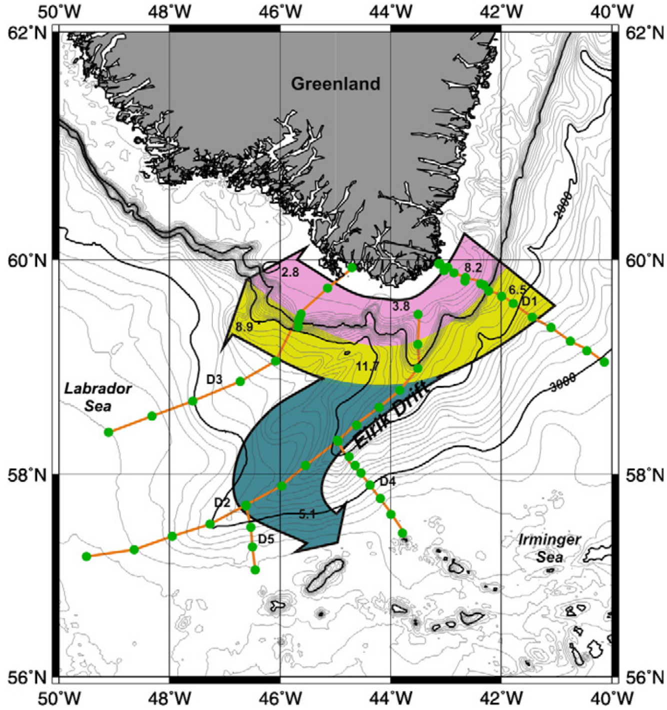

Fig. 4. Schematic diagram of the circulation pathway of the EGC and EGCC around Cape Farewell after Holliday et al. (2007). Plotted in green are the CTD stations for the 2005, D298 cruise and the orange lines show the survey lines (Bacon, 2006). Values show total transport EGC and EGCC after Holliday et al. (2007). The pink shading demarcates the cold fresh, Arctic sourced EGCC and EGC, and the yellow shading indicates Atlantic-origin EGC. The dark green shaded area represents the retroflection of the EGC and shows its southward pathway.

is greatly modified from its original properties at the Iceland–Scotland sill because it has passed through and mixed with a variety of water masses en route to Cape Farewell. For this reason, the ISOW-derived water mass at this location is sometimes called North East Atlantic Deep Water (NEADW) or Charlie-Gibbs Fracture Zone Water, reflecting the particular route followed through the Mid Atlantic Ridge from the Iceland Basin to the Irminger Basin. This water mass will be referred to as ISOW below.

Using the D298 CTD/LADCP datasets, Holliday et al. (2009) described a (~30%) reduction in transport of the DWBC as it rounds the southern tip of Greenland, similar to the divergence and retroflection of part of the EGC. Holliday et al. (2009) suggest that the barotropic DWBC diverged due to the sharply changing topography. The divergent branch contains the most saline components of the ISOW. The divergence of the DWBC can also be detected in historical data. McCartney (1992) showed a water mass with high silicate concentration (characteristic of ISOW), turning southward from Eirik Drift, whilst more DSOW type waters continue westward (Holliday et al., 2009; Fig. 5). Stramma et al. (2004) concluded that ISOW may be sourced more directly from the Charlie Gibbs Fracture Zone, with the westward flowing current lying outside of the DWBC.

Bacon and Saunders (2010) analysed the results from an almost year-long moored array of current metres spanning the breadth of the DWBC south-east of Cape Farewell between the summers of 2005, from deployment on D298, and subsequent recovery on a second cruise on RRS *Discovery* in summer 2006. For waters of density (σ_θ) greater than 27.8, the mean transport was 9.0 Sv, with standard error 1.0 Sv. The range of observed 12-hour mean values was high: from a value of 0.9 Sv, observed during May 2006, to 16.4 Sv, observed during November 2005. Most of the variability was in periods between 10 and 50 days. This long-term mean of 9.0 Sv was notably (and significantly) lower than the previously-accepted value of 13 Sv at this location, using measurements

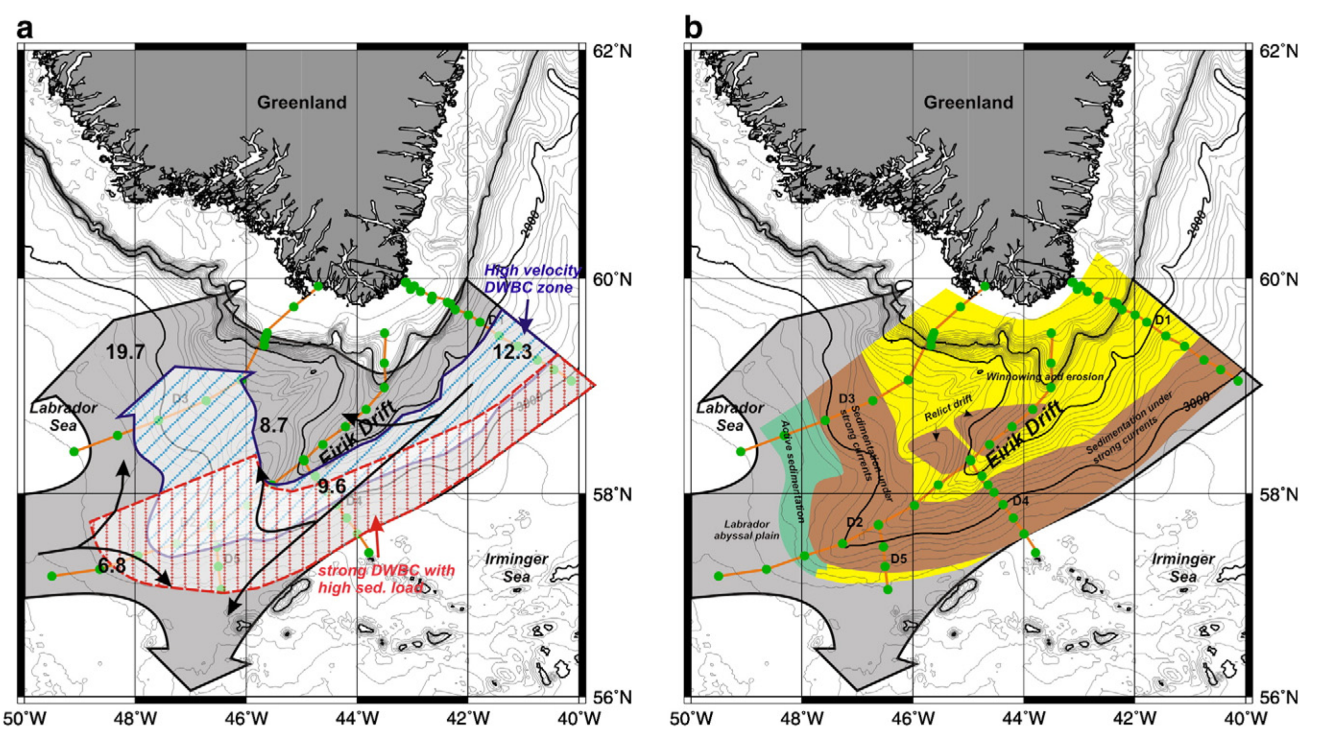

Fig. 5. a. Schematic of the DWBC system. Shaded grey arrow represents the spatial extent of the DWBC, and total transport is denoted (after Holliday et al., 2009). The red hatched area represents the zone of high sediment load and the blue hatched area, the strong DWBC flow (>7.5 cm s^{-1}) after Hunter et al. (2007b). b. Schematic of the sedimentation on Eirik Drift after Hunter et al. (2007b). The coloured shaded areas represent the different echo-characteristics mapped. Note we only show here the sediment zones relevant to the DWBC deposition. The grey shaded arrow demarcates the area of DWBC spatial extent after Holliday et al. (2009).

made in 1978 (Clarke, 1984; Dickson and Brown, 1994). Bacon and Saunders (2010) make a case for the earlier value actually being an underestimate through lack of upper-ocean absolute current measurements. They suggest a more likely value for the 1978 transport to be 16 Sv. There may be a link to the multidecadal DWBC baroclinic (i.e. partial, hydrographic) transport record of Bacon (1998), which shows much stronger DWBC transport at this location in the 1970s–1980s than before or afterwards.

Using data from the *Discovery* 2005 cruise, as well as historical data from the World Ocean Database (WOD) (Conkright et al., 2002), Hunter et al. (2007b) built upon their previously published work (Hunter et al., 2007a) to better understand the sedimentary history of Eirik Drift and how the present-day deep current system has caused the drift morphology to evolve. Velocity profiles were used in combination with oxygen concentration (indication of ventilation age), light transmittance (indication of sediment load), and silicate concentration (allows estimation of LSW contribution) to generate five detailed profiles across Eirik Drift (Hunter et al., 2007b; Fig. 4, D1–D5). The profiles indicated that the DWBC deepened with bathymetry as it rounded Eirik Drift and extended beyond the survey area. Hunter et al. (2007b) showed that the primary current core was situated towards the bottom of the slope, at a depth of around 2000 m, and that a second current strand representative of ISOW (higher oxygen saturation and lower salinity) was present. Associated with a decrease in the transport of the DBWC over the Eirik Drift is a decrease in bottom current velocity. The LADCP profiles showed slowing of the deepest 300 m of water from 0.16 ms^{-1} at the Irminger Sea section to 0.12 ms^{-1} over the Eirik Drift and in the Labrador Sea (Holliday et al, 2009).

Beam attenuation data suggested that below 2800 m, a 300 m thick nepheloid layer was well developed (Hunter et al., 2007b; Fig. 5). The oxygen saturation, salinity and beam attenuation data, when cross-plotted, indicate that DSOW (high oxygen saturation, low salinity) carries the main sediment load. However, Hunter et al. (2007b) suggest that the ISOW may at times, also be a source of sediment supply. These results validate previous suggestions by Innocent et al. (1997) that regions of current with highest suspended sediment loads do not necessarily coincide with the fastest current velocities.

3.3. Sedimentary regimes on Eirik Drift: Past and present

The acoustic character of the sediments were described by Hunter et al. (2007a,b) primarily using 3.5 and 5.1 kHz profiler data collected in 2003 (TTR-13), 2004 (CD-159) and 2005 (D298), and sidescan sonar images from 2003 (TTR-13). The characteristics were related to the deep water current regime above the seafloor and were combined with core-top grain size data from cores TTR-451, TTR-450 and D298-P3, along with previously published bottom photographs (Rabinowitz and Eittreim, 1974) which aided the interpretation of the geophysical images (Hunter et al., 2007b). Also shown were clay mineral percentages by Fagel et al. (1997), which Hunter et al. (2007b) interpret as a tracer for DWBC sediment transportation.

Hunter et al. (2007b) found high-amplitude seabed echo returns and high acoustic backscatter on the southeastern flank of the main ridge and southwestern flanks of the upper two secondary ridges above ~2800 m, which they interpret as representing a coarse seabed after Damuth (1980) and Stow et al. (2002). This was confirmed by the sandy top of core TTR-450 and from seabed photographs. This area underlies the upper region of DWBC where up to 25 cm s^{-1} velocities were recorded (Hunter et al., 2007b) and therefore these sediments are likely winnowed.

Similar echo characteristics were identified by Hunter et al. (2007b) on the northeastern flank of the main ridge and on secondary ridges, but with deeper acoustic penetration of parallel sub-bottom reflectors, typical of contourite deposition (Damuth, 1980; Stow et al., 2002) Within this zone, winnowed sand at the top of TTR-451 indicates that there is very little modern active deposition, but the D298 hydrographic survey suggests that there is today a lack of DWBC activity capable of achieving such winnowing. Conversely, spot velocity measurements and surface lineations in the sidescan sonar images would support the presence of significant flow across the ridge, and high near-bottom turbulence, indicative of strong bottom flow across the ridge, was documented by Lauderdale et al. (2008).

Beam attenuation data would suggest that little sediment is carried by the DWBC above 2800 m water depth (Hunter et al., 2007b) and a high proportion of illite in core-top sediments from this region indicates that ice-rafted debris (IRD) forms the main deposition (Fagel et al., 1997; Hunter et al., 2007b). Hunter et al. (2007b) suggested that the patterns of sedimentation and erosion, along with the historic hydrographic data, indicate the presence of separate flow strands across Eirik Drift. Hunter et al. (2007b) question whether the drift crest was a zone of erosion during the Holocene, whereas active sedimentation occurred during the Plio–Pleistocene (e.g., Hillaire-Marcel et al., 1994) despite the presence of strong DWBC activity (Arthur et al., 1989; Hunter et al., 2007a). They conclude that this likely resulted from higher sediment loads in the DWBC during the Plio–Pleistocene due to the uplift of the Greenland–Scotland Ridge (Wold, 1994), forming a sediment source.

On the other hand, the lower flanks and toe of Eirik Drift are characterised by relatively thick (<12 m) Holocene deposits of relatively fine-grained, clay/silt sediments (Hunter et al., 2007b; Fig. 5). It is at this water depth (below 2800 m) that Hunter et al. (2007b) identified the distinct nepheloid layer associated with DSOW, and it is likely that the high sediment current load combined with lower current velocities result in such high deposition rates. These sediment characteristics, only seen to the north of the ridge flank likely indicate an enhancement of deposition as the flow slows to round the southern tip of Eirik Drift (Hunter et al., 2007b).

3.4. Meltwater injections and their impact on AMOC and climate during the time period of Heinrich event 1 and the last deglaciation

Since it is speculated that freshwater additions into the North Atlantic may result in decreased rates of NADW formation, and reduced poleward heat transport (e.g. Stommel, 1961; Broecker, 1991), the remit of this study was to further elucidate the role of past freshwater forcing on the rate of NADW formation and its impact upon climate. This has been achieved through multi-proxy reconstruction from marine sediment core TTR-451 (Stanford et al., 2006), and comparison of these findings with other, well-dated palaeo-proxy records. This study has focussed upon the time period of the last deglaciation (20–11 ka BP), which encompasses the climate deterioration and iceberg discharge event of Heinrich event 1, the abrupt Bølling warming (when Greenland temperatures rose by more than 10 °C in only a couple of centuries (Severinghaus and Brook, 1999)), meltwater pulse (mwp)-1a, (a sea-level rise of around 20 m in around 500 yr (e.g. Bard et al., 1996; Hanebuth et al., 2000; Peltier and Fairbanks, 2006)), and the sharp climatic deterioration of the Younger Dryas.

Stanford et al. (2006) showed that the much debated timings of the Bølling and mwp-1a were conclusively resolved by comparison of the GRIP ice core $\delta^{18}O$ record on the new layer-counted GICC05 time scale (Rasmussen et al., 2006) with the well constrained U/Th-dated Barbados fossil coral sea-level record (Fairbanks et al., 2005; Peltier and Fairbanks, 2006). This record comparison confirmed previous suggestions (e.g. Bard et al., 1996; Liu and Milliman, 2004) that mwp-1a occurred at around 14.1 ka BP, culminating in a meltwater peak at around 13.9 ka BP. Therefore, mwp-1a was coincident with the abrupt “Older Dryas” (GI-1d) (Björck et al., 1998) cooling event and not the Bølling warming (14.6 ka BP). Even when considering the 2σ uncertainties to these ages, Stanford et al. (2006) suggest that mwp-1a lagged behind the Bølling warming by over 3 centuries. On this basis, the hypothesis was rejected that mwp-1a coincided with, and may have triggered, the Bølling warming (e.g. Hanebuth et al., 2000; Clark et al., 2002; Kienast et al., 2003; Weaver et al., 2003).

However, at question remained an accelerated mass spectrometric (AMS) ^{14}C dated sea-level record from the Sunda Shelf, which suggested an earlier (Bølling warming) timing for mwp-1a than at Barbados (Hanebuth et al., 2000). We address this issue in a separate paper within this volume (Stanford et al., 2011) by combining multiple far-field sea-level records to assess the uncertainties within these datasets. Data used included records from Sunda Shelf (Hanebuth et al., 2000, 2009), Bonaparte Gulf (Yokoyama et al., 2000, 2001), Huon Peninsula (Chappell and Polach, 1991; Cutler et al., 2003), Florida Keys and the Caribbean Sea region (Toscano and Macintyre, 2003 and references therein), Tahiti (Bard et al., 1996, 2010) and Barbados (Fairbanks, 1989; Bard et al., 1990a,b; Fairbanks et al., 2005; Peltier and Fairbanks, 2006). A Monte Carlo statistical analysis was used to produce 99, 95 and 67% probability envelopes for the combined sea-level reconstruction, and by careful scrutiny of specific data points that lay outside of the 99% confidence of the combined reconstruction, for the first time, discrepancies between sea-level records, both in time and in palaeo-sea-level depth, were reconciled. Stanford et al. (2011) showed that the main phase of mwp-1a occurred between 14.3 and 12.8 ka BP, with the highest acceleration of sea-level change at around 13.8 ka BP (Fig. 6).

Stanford et al. (2006) also presented a record from marine sediment core TTR-451 of NADW flow intensity derived from the ratio of the susceptibility of anhysteretic remanent magnetisation (κ_{ARM}) versus low-field magnetic susceptibility (κ), a proxy for average (titano)magnetite grain size (Banerjee et al., 1981; Verosub and Roberts, 1995). Since previous studies have suggested that the magnetic mineral content of sediments along the flowpath of NADW originates from the Nordic basaltic province (Kissel et al., 1999, 2009), and deposition on Eirik Drift is dominated by suspended matter

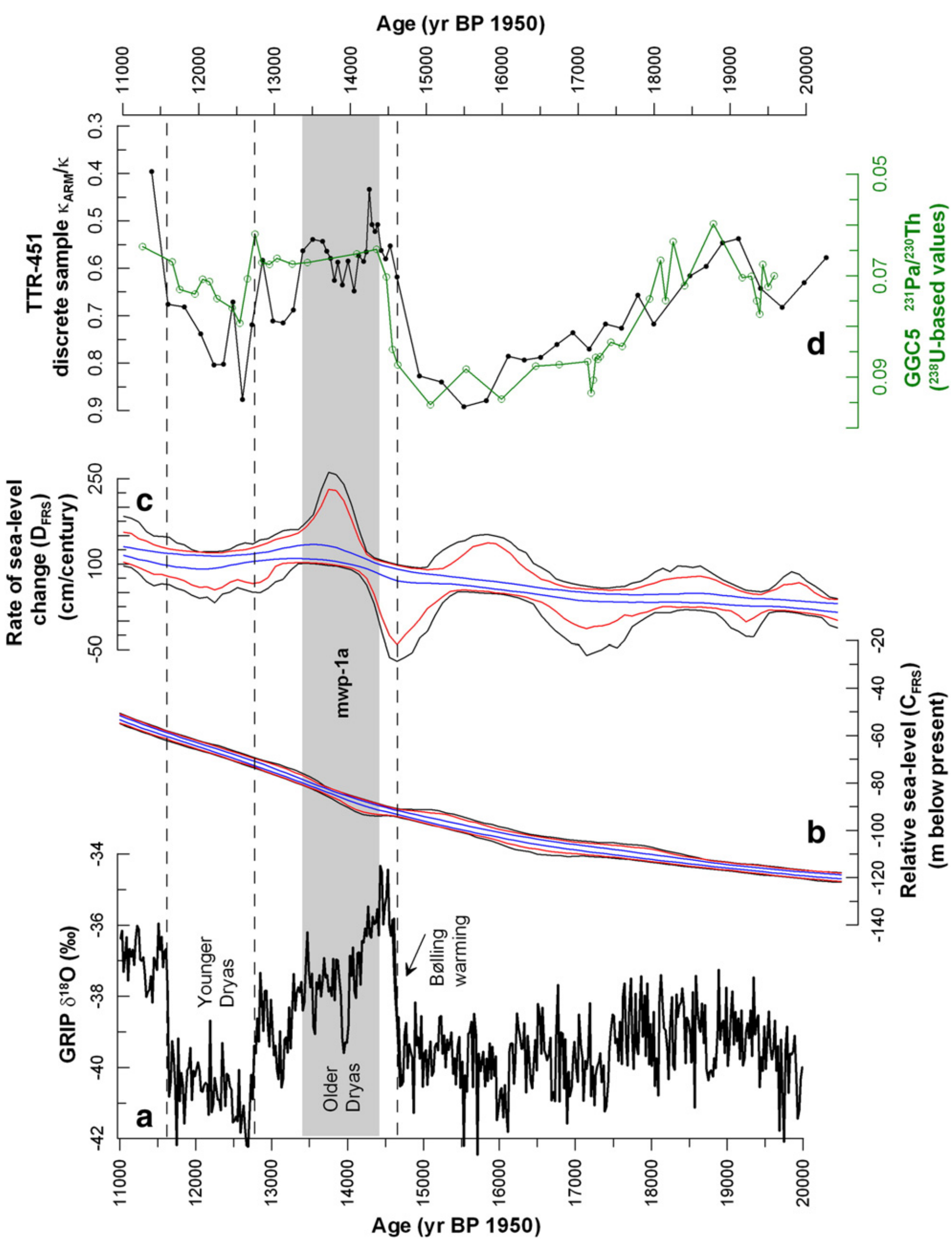

Fig. 6. a. The GRIP ice core $\delta^{18}O$ record on the GICC05 timescale on the basis of layer-counting (Andersen et al., 2006; Rasmussen et al., 2006; Svensson et al., 2006; Rasmussen et al., 2008). b. Combined far-field relative sea-level record (C_{FRS}) and c, the first derivative of it (D_{FRS}) (i.e., the rate of sea-level change) after Stanford et al. (2011) The blue line represents the 67% confidences, the red line, the 95% confidence limits, and the black line, the 99% confidence limits. d. The record of κ_{ARM}/κ for Eirik Drift core TTR-451 from discrete samples (in black) is presented alongside the $^{231}Pa/^{230}Th$ record of core GGC05 from Bermuda Rise (in green) (McManus et al., 2004).

transported via deep-water overflow through Denmark Strait (Hunter et al., 2007a,b), the κ_{ARM}/κ record from Eirik Drift was interpreted to reflect variation in the size of magnetic grains that can be carried by DSOW and ISOW in NADW, and which settle out on Eirik Drift as the water mass rounds the southern tip of Greenland. The interpretation of κ_{ARM}/κ values from Eirik Drift as a magnetic grain size indicator was further tested and validated by additional environmental magnetic measurements, scanning electron microscopy (SEM) studies and sortable silt grain size analyses (Stanford, PhD thesis 2008). It is puzzling however, that today these currents lie well below the position of TTR-451 and yet the core contains a condensed/winnowed Holocene top, suggesting that vigorous current activity has eroded the site during the last few thousand years (Hunter et al., 2007b).

By combining the (κ_{ARM}/κ) proxy record of NADW flow intensity from Eirik Drift with the Barbados sea-level and Greenland ice-core $\delta^{18}O$ (temperature) records, Stanford et al. (2006) showed that at the time of mwp-1a there was a ~500 yr reduction in NADW flow intensity. This slowdown was coincident with the brief Older Dryas cold period. In Fig. 6, we present the combined far field relative sea-level reconstruction (C_{FRS}), and its first time derivative (D_{FRS}) Stanford

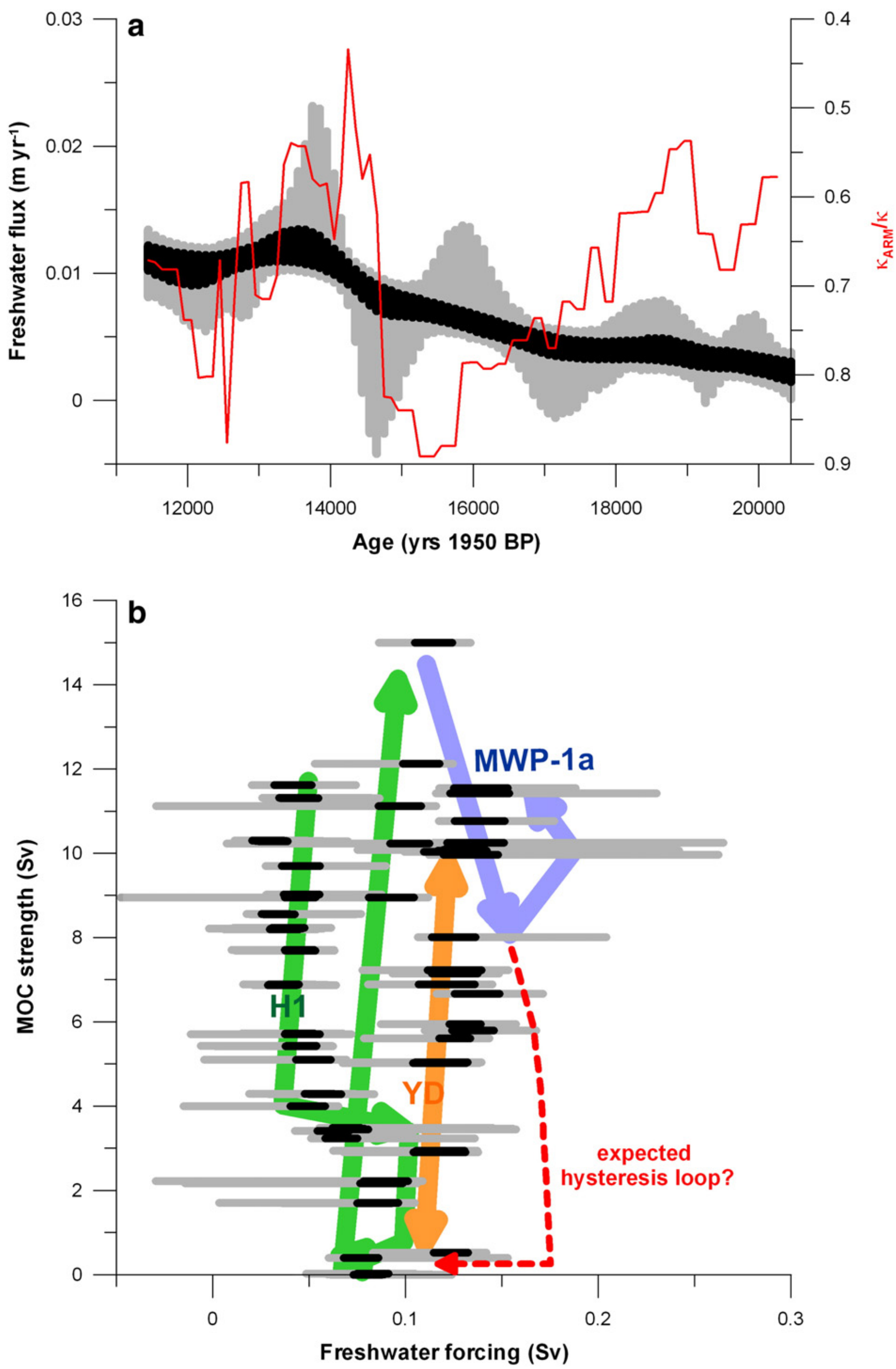

Fig. 7. a. In red is the κ_{ARM}/κ record for NADW intensity from TTR-451, interpolated at the same time intervals as the record of D_{FRS} (Stanford et al., 2011), which is shown in grey (95% confidence) and black (67% confidence). b. A cross-plot of the derived AMOC intensity (scaled assuming that the Bølling AMOC intensity equals 75% of the modern AMOC strength (~15 Sv; McManus et al., 2004)) versus freshwater forcing in Sv. The horizontal black lines represent the 67% uncertainty in the D_{FRS} reconstruction, and the grey lines represent the 95% confidence. In red is the extrapolated hysteresis loop for mwp-1a. H1 = Heinrich event 1 and YD = Younger Dryas.

et al. (2011) alongside the GRIP $\delta^{18}O$ ice core record and the AMOC flow intensity proxies. The new time constraints now place mwp-1a at the termination of the Bølling warming, and possibly initiating within the Bølling period as a direct response to high-latitude warming during that period (McManus et al., 2004; Stanford et al., 2006). This compilation of records also serves to demonstrate the characteristic near-collapsed NADW formation during more extreme cooling events which were not associated with meltwater pulses large enough to significantly affect the sea-level record (Heinrich event 1 and the Younger Dryas). Stanford et al. (2011) suggest that mwp-1a likely caused the only brief reduction in NADW formation and cooling associated with the minor reductions in temperatures at the Older Dryas.

In order to better assess how the AMOC varied in response to total freshwater forcing of the world ocean, we cross-plot the calculated flux of injected freshwater (in Sv) derived from the D_{FRS} reconstruction (Stanford et al. (2011)), against the κ_{ARM}/κ record from TTR-451, which is scaled to AMOC strength after McManus et al. (2004) (Fig. 7). As previously suggested by Stanford et al. (2006), Fig. 7 shows a distinctly complex behaviour between the rate/magnitude of meltwater injections and the AMOC responses. The results in Fig. 7 show that during the AMOC collapses of the Younger Dryas and H1, the freshwater perturbation was actually diminishing (see orange and green arrows in Fig. 7). This suggests that it might not be the instantaneous influx of freshwater that is important to invoke an AMOC collapse, but its cumulative impact over time. For mwp-1a (>0.15 Sv of freshwater forcing), only a minor reduction of AMOC intensity is observed. If we extrapolate this curve to 'collapse' or to close a typical hysteresis loop (e.g. Rahmstorf et al., 2005), this would suggest that something in the order of 0.175 Sv or more of freshwater would be required. This suggests that the ocean–climate system is much less sensitive than ocean–climate models might imply, with values for a complete AMOC collapse ranging from 0.1 to 0.5 Sv of freshwater forcing/perturbations (e.g., Manabe and Stouffer, 1997; Rahmstorf et al., 2005; Liu et al., 2009). Alternatively, Lohmann and Schulz (2000) used an ocean general circulation model with deepwater formation (unlike previous models) parameterised north of the Greenland–Scotland Ridge, and showed that mwp-1a did not result in a complete AMOC collapse. Instead, they suggest that the mwp-1a freshwater forcing may have preconditioned the North Atlantic for the Younger Dryas AMOC collapse (Fig. 7).

The complete AMOC collapse for H1 and the YD and the more minor collapse during mwp-1a suggest that the ocean–climate system displays non-linear behaviour (Fig. 7). Explanation for such non-linear behaviour may be found in model simulations which demonstrate these characteristics for different quasi-stable climate states (Rahmstorf, 1995). Reconciliation between the palaeo-reconstructions and ocean circulation models may also be gained if the ocean–climate system is shifted to where more than one AMOC state might co-exist (Knorr and Lohmann, 2007). Alternatively, it may be that the depth-distribution or the location of meltwater injection into the oceans are more important for NADW formation than the magnitude or rate (Moore, 2005; Tarasov and Peltier, 2005; Stanford et al., 2006). For example, it has been suggested that mwp-1a (or a considerable component) may have entered the ocean via the Gulf of Mexico at intermediate depths (Flower et al., 2004; Aharon, 2005), and it maybe that strong mixing with ambient seawater could have reduced its impact on NADW formation (Aharon, 2005; Tarasov and Peltier, 2005).

On the other hand, to explain the AMOC collapses during Heinrich events and the Younger Dryas, Moore (2005) and Tarasov and Peltier (2005) proposed that a small meltwater flux, not large enough to discern in sea level records, injected into the Arctic and/or Nordic Seas (critical regions for NADW formation) could have triggered NADW formation collapse. Conversely, an ocean–climate simulation suggests that a freshwater lid would have likely suppressed deepwater formation irrespective of the location of the freshwater injection, with a sustained collapse resulting in secondary feedbacks and sea-ice growth (Otto-Bliesner and Brady, 2010).

It is clear from this study and from previous model investigations (e.g. Knorr and Lohmann, 2007; Huang and Tian, 2008) that climatic impacts are not simply governed by the magnitude and/or rate of meltwater addition. If climate forcing was indeed dependent on freshwater input, then (small) freshwater additions targeted on the Arctic/Nordic Seas (a key region of NADW formation today — see e.g. Dickson and Brown, 1994; Bacon, 1998, 2002) and/or during stadial modes of AMOC (e.g. Rahmstorf, 2002) may increase the risk of disruption to NADW formation. Alternatively, secondary feedback mechanisms such as sea-ice growth need to be considered (e.g. Wunsch, 2006). To evaluate these possibilities, new palaeo-proxy records are required that constrain other potential aspects of ocean–climate interaction, such as sea-ice feedbacks and seasonality changes (e.g. Broecker, 2001; Gildor and Tziperman, 2003; Denton et al., 2005; Li et al., 2005; Wunsch, 2006; Peck et al., 2008).

4. Conclusions

The Cape Farewell project at Eirik Drift set out to address three key issues; to describe the modern hydrography around Eirik Drift, to assess how these hydrographies affect sedimentation, and how past freshwater forcing affected the DWBC and hence climate. This has been achieved through a multi-disciplinary study of both new and old hydrographic surveys, along with geophysical images of the sea-floor and sub-bottom sediments and multi-proxy analyses of marine sediment cores.

The results indicate that the EGC, EGCC and DWBC that flow around Cape Farewell as western boundary currents are complex in terms of how they interact with sea-bottom topography, surrounding water masses, and their short- and long-term variability, which have influenced sedimentary deposition on Eirik Drift. By combining a record of NADW strength from Eirik Drift with palaeo sea-level and other well dated proxy records (namely, the Greenland $\delta^{18}O$ ice core records of temperature), we conclude that there is a fundamental non-linear behaviour to how the ocean–climate system responds to freshwater forcing during the last deglaciation. These studies form part of ongoing projects in which we hope to build upon the results obtained from the Cape Farewell project.

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