North Atlantic climate and deep-ocean flow speed changes during the last 230 years

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[1] Variations in the near-bottom flow speed of Iceland-Scotland Overflow Water (ISOW) are documented in a 230-year-long deep-sea sediment record from the eastern flank of Reykjanes Ridge in the subpolar North Atlantic at (sub)decadal time scales. For recent decades, the ISOW palaeocurrent reconstructions show similarities with observational hydrographic data. Furthermore, recent ISOW flow changes fall mostly within the range of its variability of the past 230 years. The record also reveals a hitherto unrecognized coupling of deep flow speeds in the subpolar North Atlantic with the North Atlantic Oscillation (NAO) index, with more (less) vigorous ISOW flow during negative (positive) phases of the NAO. Our results suggest that the changes in ISOW vigor are largely controlled by the transport and characteristics of Labrador Sea Water rather than variations in the overflow itself, with implications for the meridional overturning of the Atlantic Ocean and climate. **Citation:** Boessenkool, K. P., I. R. Hall, H. Elderfield, and I. Yashayaev (2007), North Atlantic climate and deep-ocean flow speed changes during the last 230 years, *Geophys. Res. Lett.*, *34*, L13614, doi:10.1029/2007GL030285.

1. Introduction

[2] The Atlantic meridional overturning circulation (MOC) is responsible for a substantial component of the meridional heat transport in the Atlantic Ocean [*Ganachaud and Wunsch*, 2000] and models have shown that alterations in its strength and spatial structure can lead to abrupt climate change [*Rahmstorf*, 2002]. Iceland-Scotland Overflow Water (ISOW) and Denmark Strait Overflow Water (DSOW) jointly comprise the cold, dense overflows from the Greenland, Iceland and Norwegian (Nordic) Seas across the Greenland-Scotland Ridge (GSR) [*Hansen and Østerhus*, 2000]. They are the precursors of lower North Atlantic Deep Water (NADW) and form part of the downwelling limb of the MOC. There is some evidence that the strength of the GSR overflows is modulated by the North Atlantic Oscillation (NAO) [*Biastoch et al.*, 2003; *Dickson et al.*, 1996, 2000], which is the dominant mode of atmospheric variability in the North Atlantic sector [*Hurrell*, 1995]. The NAO is usually expressed as an index of the normalized sea-level pressure difference between Iceland and Lisbon [*Hurrell*, 1995] or Gibraltar [*Jones et al.*, 1997], and its influence is most pronounced during winter.

[3] As ISOW descends into the deep North Atlantic its volume nearly triples due to entrainment of salty, warm Atlantic subpolar mode water (SPMW) and, subsequently, at intermediate depths, Labrador Sea Water (LSW), which is relatively fresh [*Hansen and Østerhus*, 2000]. Observations suggest that ISOW, and the water masses it entrains, freshened [*Dickson et al.*, 2002], and that its flow weakened [*Hansen et al.*, 2001] over the last four decades of the 20th century. However, it is not known if these changes fall within the natural variability of the system.

[4] Here we present a 230-year record of variations in the near-bottom flow speed of ISOW at subdecadal timescales from a deep-sea sediment core from the eastern flank of Reykjanes Ridge (hereafter $ISOW_{RR}$). We compare these palaeocurrent reconstructions with observational hydrographic data from the same area and corroborate our findings with geostrophic calculations of $ISOW_{RR}$ transport. Furthermore, we explore the relationship between the deep flow speed changes of $ISOW_{RR}$ and the NAO index, and reveal a hitherto unrecognized inverse relation.

2. Material and Methods

[5] The flow of $ISOW_{RR}$ has built up extensive sediment drifts from remobilized sediments originating from the SE Iceland slope along the eastern slope of Reykjanes Ridge [*Bianchi and McCave*, 2000]. Interactions between the bottom currents and sea-bed topography have led to generally high sedimentation rates, exceptionally up to $\sim$120 cm·ka^{-1} on certain areas of the Gardar Drift through the Holocene [*Ellison et al.*, 2006]. We present data obtained from a 54.3-cm-long sediment box-core, RAPID-21-12B (57°27.09′N, 27°54.53′W, 2630 m water depth), recovered from the southern limb of the Gardar Drift during *RRS Charles Darwin* cruise 159. The core was dated using the ^{210}Pb method (see auxiliary material[1]), providing a maximum linear estimated sedimentation rate of 2.3 ± 0.2 mm·a^{-1}. It was sampled continuously at 0.5 cm intervals, each sample representing an integrated signal of 2.2 ± 0.2 years. Changes in the near-bottom flow speed of $ISOW_{RR}$ through the interval 1770–2004 AD (*Anno Domini*; all years in AD) were reconstructed using the palaeocurrent proxy 'sortable' silt mean grain size ($\overline{SS}$), the mean grain size of the 10–63-μm terrigenous silt fraction [*McCave and Hall*, 2006; *McCave et al.*, 1995]. This proxy varies independently of sediment supply in current-sorted and

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[1]Auxiliary material data sets are available at ftp://ftp.agu.org/apend/gl/2007gl030285. Other auxiliary material files are in the HTML.

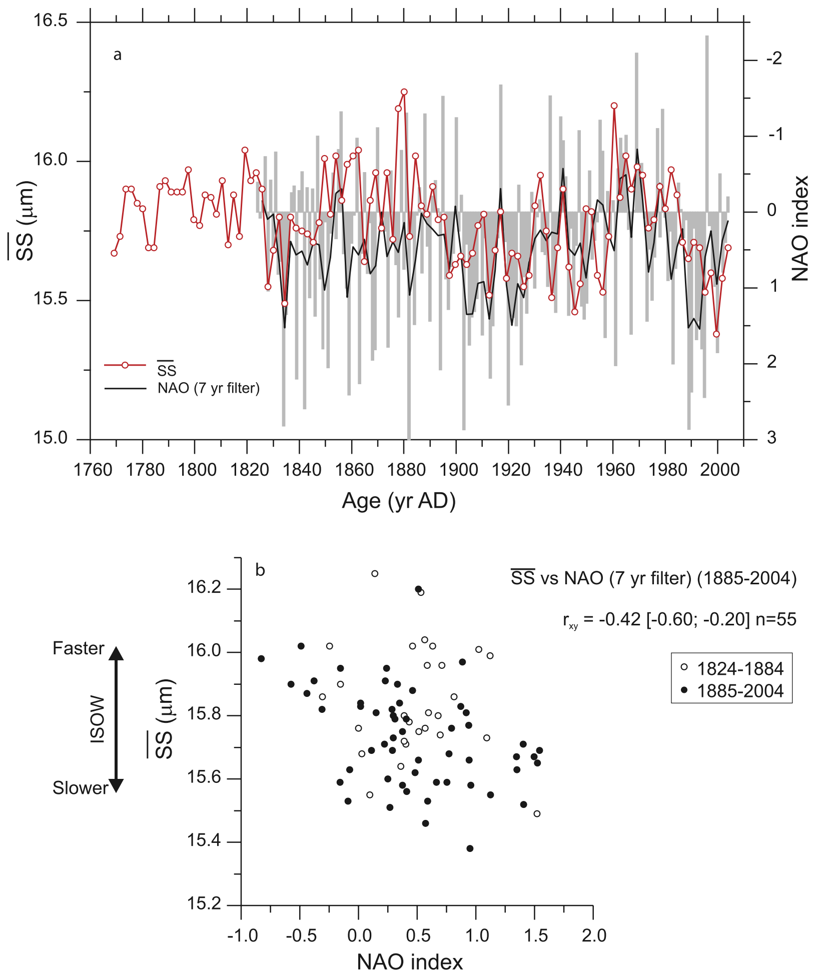

Figure 1. Evidence of a linear relationship between near-bottom flow speed changes of $ISOW_{RR}$ and the NAO index. (a) $\overline{SS}$ record of core RAPID-21-12B (in μm, red) recording near-bottom flow speed changes in Iceland-Scotland Overflow Water on the eastern flank of Reyjanes Ridge ($ISOW_{RR}$; see Figure 3, inset), larger grain sizes correspond to faster near-bottom flow); winter index of the North Atlantic Oscillation (NAO) [*Jones et al.*, 1997] (grey bars, reverse scale), 7-year smoothed NAO index (black). (b) Scatter plot of the raw $\overline{SS}$ record versus the smoothed NAO index for 1824–1884 (open circles) and 1885–2004 (filled circles).

deposited muds; higher values represent relatively greater near-bottom flow speeds. The $\overline{SS}$ was measured using a Coulter Multisizer III (analytical precision 1–4% [*Bianchi et al.*, 1999]). The $\overline{SS}$ data in Figures 1 and 2 are averages of triplicate measurements.

[6] We used the winter (December–March) index of the NAO based on the normalized sea-level pressure difference between Gibraltar and southwest Iceland [*Jones et al.*, 1997] (http://www.cru.uea.ac.uk/cru/data/nao.htm), hereafter "NAO index". To quantify the interrelation between the $ISOW_{RR}$ flow speed record and the NAO index [*Jones et al.*, 1997] the latter was smoothed, using a 7-year Gaussian filter, since the atmosphere is generally prone to higher-frequency variability than the ocean because of the ocean's thermohaline inertia or "memory" [*Curry and McCartney*, 2001]. A 7-year filter was applied to retain sub-decadal variability in the NAO index. Subsequently, the filtered NAO record was resampled in the time domain to match the average sampling interval of the $\overline{SS}$ record (2.2 years). Pearson's correlation coefficient (r_{xy}) was estimated employing a nonparametric stationary bootstrap confidence interval with an average block length proportional to the maximum estimated persistence (serial dependence) time of the data [*Mudelsee*, 2003].

[7] The total transport of $ISOW_{RR}$ was estimated based on hydrographic stations on or in close proximity of trans-Atlantic section AR7E [*Yashayaev et al.*, 2007] using the classical dynamic method, the simplest and most widely employed method for estimating large-scale ocean current transports from observed seawater density fields. Based on geostrophic and hydrostatic dynamics, it provides the relative current in reference to a chosen level, such as a seawater density surface. Here, the isobaric surface of the deep LSW core in the Iceland Basin (~1500 decibars) was chosen as the reference level.

3. Results and Discussion

[8] The total $\overline{SS}$ range in RAPID-21-12B (0.87 μm; Figure 1a) is small compared to, e.g., the ~2 μm excursion

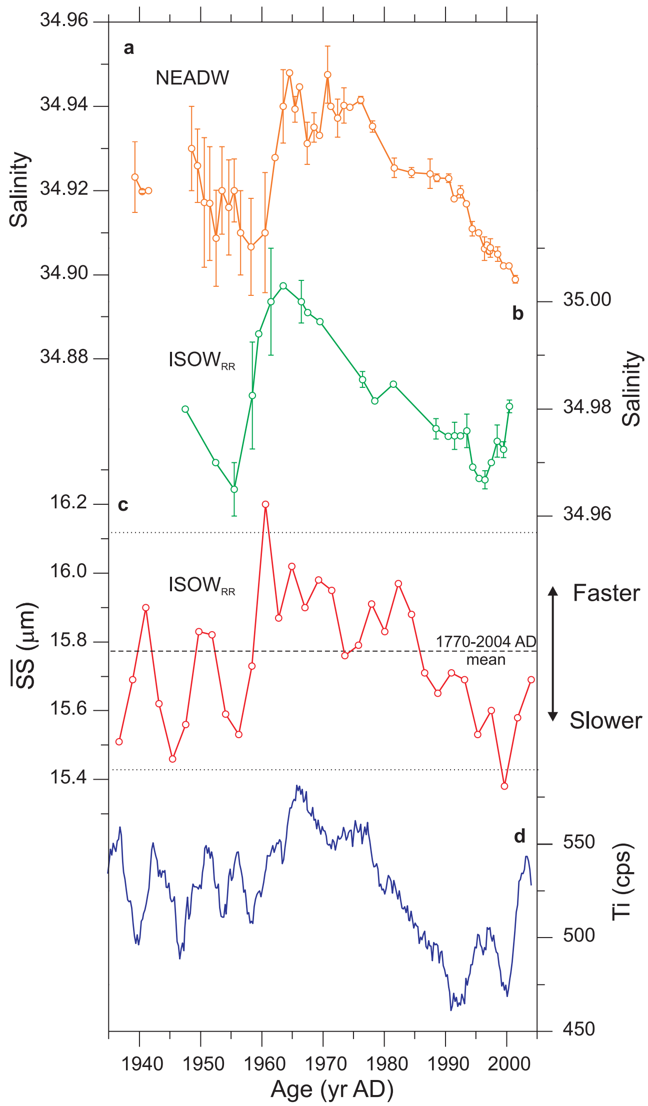

Figure 2. Recent changes in ISOW salinity and vigor along its spreading pathway to the Labrador Sea. Salinity time series at key location in the North Atlantic (updated from *Dickson et al.* [2002]): (a) North East Atlantic Deep Water (NEADW) in the Labrador Sea and (b) ISOW$_{RR}$ on the eastern flank of Reykjanes Ridge. (c) $\overline{SS}$ record of core RAPID-21-12B (μm); the dashed line depicts the 1770–2004 $\overline{SS}$ average, dotted lines indicate the 2σ interval of standard deviation. The core location lies within the area of salinity record in Figure 2b. (d) Ti content of core RAPID-21-12B (in counts per second (see auxiliary material)).

representing the reduction in ISOW$_{RR}$ flow speed during the "8.2 kyr cold event" in neighboring sediment core MD99-2251 [*Ellison et al.*, 2006], almost certainly reflecting the relatively stable nature of sedimentation over the past centuries. Nonetheless, the fluctuations are significantly greater than the precision of the method [*Bianchi et al.*, 1999]. Therefore we suggest that the observed decadal trends shown in Figure 1a reflect marked changes in ISOW$_{RR}$ vigor at the core site.

[9] For the period 1885–2004 (n = 55), the r_{xy} correlation between the ISOW$_{RR}$ $\overline{SS}$ and the NAO index was estimated at −0.42 with a 95% bias-corrected and accelerated confidence interval of [−0.60; −0.20] (Figure 1b). For the period 1824–2004 (n = 83) r_{xy} was estimated at −0.33 [−0.49; −0.14]. Both correlations are significant and strongest at zero lag, suggesting that, on decadal time scales, ISOW$_{RR}$ flow speed is significantly inversely correlated with the NAO index; i.e. the flow of ISOW$_{RR}$ has been persistently strongest (weakest) during low (high) NAO index phases. The fact that the correlation attained for the interval including the 1824–1885 period is weaker is probably due to the decreased (sub)decadal persistency of the NAO index during this time. This might be related to a poorer quality of the historic observations [*Jones et al.*, 1997], but could also reflect a change in the large-scale oceanic and atmospheric climate regimes [*Crowley*, 2000].

[10] Several studies have shown a strong link between the NAO and North Atlantic sea surface temperatures (SSTs) on annual time scales [*Deser and Blackmon*, 1993]; the observed SST pattern is likely a response to local air-sea heat flux changes. Nevertheless, the relationship between the NAO and the MOC is less well understood. We explore the link between ISOW$_{RR}$ flow speed variations and the NAO index further by concentrating on the time period for which hydrographic time series of ISOW are available. In particular, we focus on the pronounced NAO reversal from an extremely negative index state in the 1960s to dominantly positive values through the 1990s (Figure 1a), which is unprecedented in the instrumental record [*Hurrell*, 1995; *Jones et al.*, 1997]. During this period the $\overline{SS}$ record (Figure 2c) declines significantly, suggesting a marked slowdown of ISOW$_{RR}$ flow. This trend is mirrored in the decreased bulk sediment Ti concentration (Figure 2d), which suggests a reduction in the transport of Icelandic sourced low-Ti content magnetite to the core site by the southbound ISOW flow [*Ballini et al.*, 2006]. Comparison of the ISOW$_{RR}$ palaeocurrent speed record with the extended salinity time series of ISOW at two locations along its spreading pathway (Labrador Sea and eastern Reykjanes Ridge; Figures 2a and 2b [*Dickson et al.*, 2002]) suggests that changes in flow speed and salinity are coordinated. *Saunders* [1994] reported reduced volume transport of ISOW passing through Charlie-Gibbs Fracture Zone (CGFZ) from 4.6 Sverdrups (1 Sv = $10^6 \cdot m^3 \cdot s^{-1}$) in 1964 [*Worthington and Volkmann*, 1965] to 2.4 ± 0.5 Sv in 1988/89 coeval with a salinity decrease of 0.02. *Saunders* [1994] ascribed the transport reduction to overestimation of the hydrography-based 1964 value, the 1988/89 value being based on current meter records. However, hydrographic calculations by *Saunders* [1994] yielded an estimate of 3.5 Sv for 1988/89, suggesting a 24% reduction of ISOW transport through CGFZ since 1964. The sustained decreases in ISOW$_{RR}$ palaeocurrent speed and salinity into the 1990s (Figure 2) might imply that volume transport reduced even further toward the mid-1990s.

[11] Repeat trans-Atlantic hydrographic sections at the time of the NAO and ISOW$_{RR}$ flow speed extremes not only confirm that ISOW$_{RR}$ was saltier, but also show that it was denser in 1966 (Figure 3a) than it was in 1994 (Figure 3c) and 2001 (see auxiliary material). Furthermore, our geostrophic transport estimates suggest that the total transport of the deep southward flow over the eastern flank of Reykjanes Ridge decreased from 5.5–6 Sv in 1966 to 4–4.5 Sv in 1994 (Figures 3b and 3d) and 3–3.5 Sv in 2001

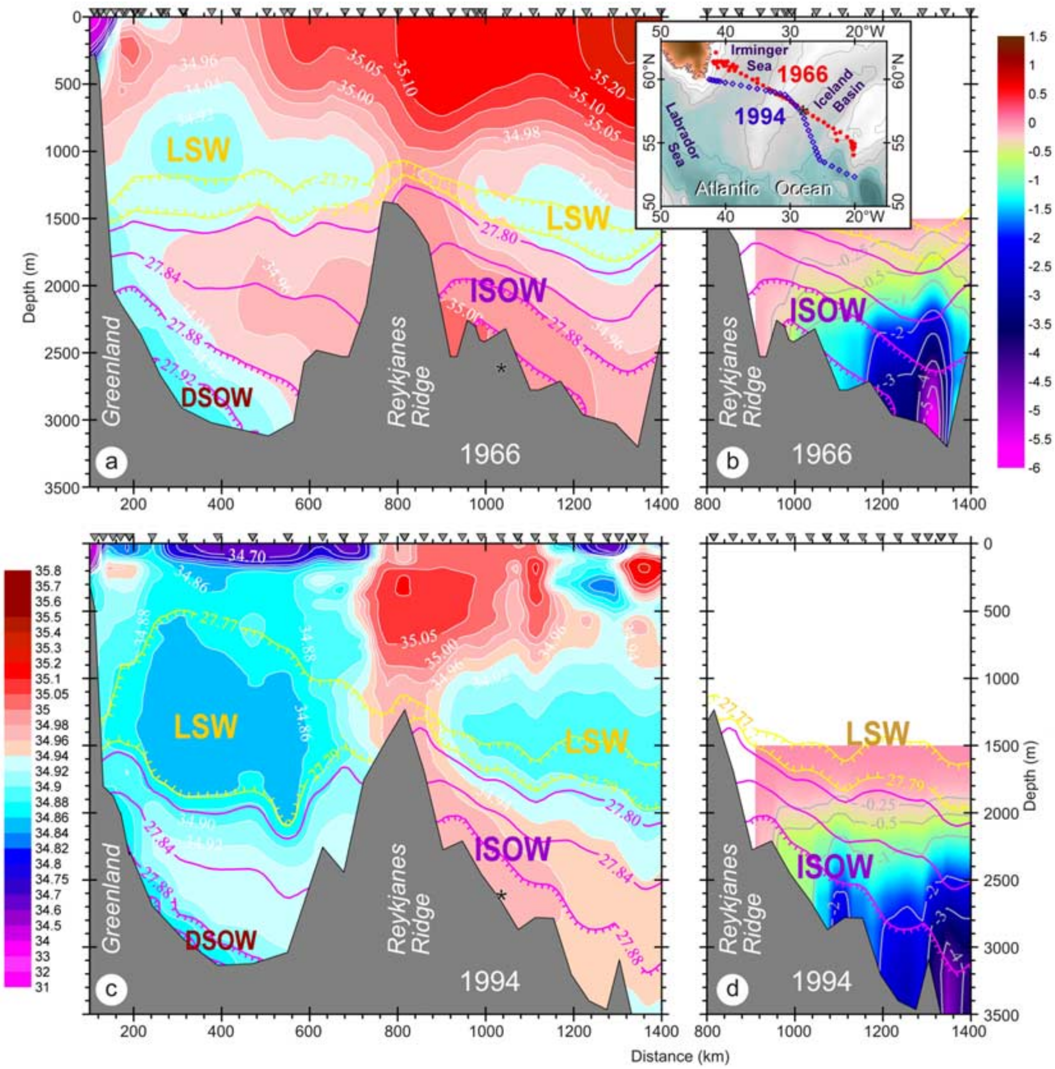

Figure 3. Trans-Atlantic hydrographic sections for 1966 and 1994. Distributions of salinity, density (isopycnal surfaces) and geostrophic transport on sections shown on the map insert for 1966 and 1994 (red and blue symbols); the locations of hydrographic stations are marked as triangles above each panel; the asterisks mark the (projected) location of core RAPID-21-12B; (a, c) salinity fields (scale left of Figure 3c) for 1966 and 1994; (b, d) geostrophic transport fields for 1966 and 1994 (scale right of Figure 3b, in Sverdrups; 1 Sv = $10^6 \cdot m^3 \cdot s^{-1}$; negative values indicate southward flow). LSW - Labrador Sea Water, DSOW – Denmark Strait Overflow Water. Yellow and magenta contours are isopycnals (kg m^{-3}) indicating changes in vertical stratification and characteristic densities of intermediate (yellow, LSW) and deep (magenta, ISOW and DSOW) waters. Estimates of transport are relative to a 1500-decibar (db) reference surface (i.e. the deep core of LSW) and show integrated total transports from the 1500-db surface and ~900 km to the east and toward the bottom (i.e. from top left to bottom right in Figures 3b and 3d). These transport estimates reveal a strong flow associated with the deep salinity maximum in 1966 and reduced total transports during 1994. The absolute magnitude of this difference may in part reflect varying LSW transport at the 1500-db surface. However, as it is likely that LSW was more voluminous and its northward transport greater in 1994 than in 1966 the actual decrease in total ISOW transport during this period therefore would likely exceed the estimated decrease relative to 1500 db.

(see auxiliary material). Since the flow of LSW is expected to weaken after crossing Reykjanes Ridge, and because it is partially compensated by the southward flow of the LSW component entrained into ISOW, these deep-water transport estimates relative to the deep LSW core are likely to represent the changes in $ISOW_{RR}$ itself.

[12] Variations in $ISOW_{RR}$ flow speed are likely caused by changes in any of its constituents: ISOW, SPMW and LSW, or by interactions with surrounding water masses. For the four decades spanning the NAO reversal, property changes in each of these water masses have been interpreted as the ocean's responses to NAO forcing through heat flux and wind stress anomalies. Although model studies suggest that the flux of ISOW is less responsive to NAO forcing than DSOW [*Biastoch et al.*, 2003; *Zhang et al.*, 2004], a slowdown of ISOW was observed between the 1960s and the 1990s [*Hansen et al.*, 2001]. Over the same period, ISOW [*Dickson et al.*, 2002], SPMW [*Curry and

Mauritzen, 2005] and LSW [*Lazier et al.*, 2002] freshened (Figures 2a and 2c). However, while ISOW and SPMW became less dense, LSW reached is freshest, but also coldest and densest, state ever recorded in 1993/94 [*Dickson et al.*, 1996; *Lazier et al.*, 2002], after several high-NAO-index winters during 1990–1994 led to exceptionally vigorous and deep convection in the Labrador Sea. An unprecedented large, dense body of LSW was formed and exported, and spread within 2–5 years to fill the intermediate depths of the subpolar North Atlantic (Figure 3c [*Yashayaev et al.*, 2007]). Contrary to the 1990s, LSW production, density and export were greatly reduced during the 1960s (Figure 3a) through diminished deep convection due to mild winters with low heat loss to the atmosphere, and weaker winds, which are commonly associated with NAO minima [*Dickson et al.*, 2002].

[13] Some model studies suggest convective changes in the Labrador Sea may be more important in the production of NADW than variations in the GSR overflows [*Cheng and Rhines*, 2004; *Eden and Jung*, 2001]; particularly, the modeled volume transport of the overflows weakens when LSW formation intensifies. We suggest that over the 1960s-to-1990s NAO reversal, the increasing volume of denser and fresher LSW (Figures 3a and 3c) possibly slowed the deep flow of $ISOW_{RR}$ through reducing the density gradient in the Iceland Basin, increased entrainment, and/or increasing deep hydrostatic pressure. A reduced density gradient in the Iceland Basin would likely diminish the geostrophic transport across the Iceland-Scotland Ridge. A decrease in volume transport across the ridge has been observed during this period [*Hansen et al.*, 2001]. Such a reduction in the southward flow would in turn weaken the ability of ISOW to entrain saline SPMW, resulting in a freshening of $ISOW_{RR}$. In addition, the expansion of denser LSW probably increased the amount of LSW entrained by ISOW, thus reducing $ISOW_{RR}$ salinity, density and transport further (Figures 2c, 3b, and 3d). Recent evidence [*Yashayaev et al.*, 2007] suggests that the large body of LSW formed during the early-mid 1990s filled the upper intermediate layers of the Labrador and Irminger basins with denser waters than previously present. This likely increased hydrostatic pressure on the levels below LSW (including ISOW) in these basins, and reduced the tilt of isobaric surfaces across Reyjanes Ridge, in turn decelerating ISOW transport from the Iceland to the Irminger Basin.

[14] Diminished LSW production and export after 1999 [*Lazier et al.*, 2002] would be expected to have reversed the trends in ISOW volume, flow speed, and salinity. Such reversals have indeed been observed in the overflow volume [*Curry and Mauritzen*, 2005] and are confirmed from the ISOW salinity time series (Figure 2b) and proxy flow speed record (Figure 2c).

[15] Our results have implications for monitoring future MOC stability. Most oceanographic observational time series started in the 1960s and mainly cover the transition from the 1960s negative-NAO phase to the positive NAO period of the 1990s. Although pre-1960s hydrographic data are sparse, the $\overline{SS}$ record (Figure 2c) again closely mirrors the extended observational salinity time series from along the spreading pathway of ISOW (Figures 2a and 2b). It is only the maximum of 1961 and the minimum of 2000 in the $ISOW_{RR}$ flow speed record that lie outside of the 2σ range of the 230 year record, suggesting that most of the changes in $ISOW_{RR}$ flow during the last decades of the 20th century cannot be distinguished from natural variability in the deep flow. Average values for both $ISOW_{RR}$ flow speed and salinity time series were recorded during the mid-1970s, suggesting that this period coupled with the 1960–2000 range might provide useful reference levels for future oceanographic changes.

[16] In conclusion, our palaeocurrent speed record suggests that decadal scale changes occurred in $ISOW_{RR}$ vigor during the past 230 years that are significantly, inversely correlated with the NAO index. Variability in hydrographic properties of each of the constituents of $ISOW_{RR}$ (ISOW; SPMW; LSW) has previously been attributed to NAO forcing. Many of these changes theoretically work in the same direction either slowing down or speeding up $ISOW_{RR}$ flow. Therefore, it is difficult to ascribe the observed flow speed changes to a single cause. However, our data suggest that, apart from governing the major North Atlantic SST patterns, the NAO influences deep-ocean current speeds in the area. We suggest that LSW plays a major role in transferring NAO forcing to the deep ocean.

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