Accelerated drawdown of meridional overturning in the late-glacial Atlantic triggered by transient pre-H event freshwater perturbation

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[1] Abrupt decreases of the Atlantic meridional overturning circulation (MOC) during the Late Pleistocene have been directly linked to catastrophic discharges of glacimarine freshwater, triggering disruption of northward marine heat transport and causing global climate changes. Here we provide measurements of excess sedimentary $^{231}Pa/^{230}Th$ from a high-accumulation sediment drift deposit in the NE Atlantic that record a sequence of sudden variations in the rate of MOC, associated deep ocean ventilation and surface-ocean climatology. The data series reveal a sequential decrease in the MOC rate at ~18.0 ka BP ago that coincides with only transient and localized freshwater inputs. This change represents a substantial, though not total, cessation in MOC that predates the major Heinrich (H1) meltwater event by at least 1,200 years. These results highlight the potential of targeted freshwater perturbations in promoting substantial MOC changes without a direct linking with catastrophic freshwater surges. **Citation:** Hall, I. R., S. B. Moran, R. Zahn, P. C. Knutz, C.-C. Shen, and R. L. Edwards (2006), Accelerated drawdown of meridional overturning in the late-glacial Atlantic triggered by transient pre-H event freshwater perturbation, *Geophys. Res. Lett.*, *33*, L16616, doi:10.1029/2006GL026239.

1. Introduction

[2] ^{231}Pa (half life $t_{1/2} = 32.5$ ka) and ^{230}Th ($t_{1/2} = 75.2$ ka) are particle-reactive radionuclides produced in the oceans at a constant initial $^{231}Pa/^{230}Th$ activity ratio of 0.093 and deposited in underlying sediments by attachment to settling particles. Interest in the application of excess, or unsupported, sediment $^{231}Pa/^{230}Th$ (decay-corrected to the time of deposition; herein referred to as $^{231}Pa_{xs}/^{230}Th_{xs}$) as a palaeocirculation proxy in the Atlantic is based on the similar residence time of ^{231}Pa (~100–200 yrs) [*Nozaki and Nakanishi*, 1985] and Atlantic deep waters [*Broecker*, 1979], whereby the rate of meridional overturning circulation (MOC) directly affects the lateral export of ^{231}Pa [*Yu et al.*, 1996; *Marchal et al.*, 2000]. Presently, approximately half of the water column production of ^{231}Pa is exported in North Atlantic Deep Water (NADW) to the Southern Ocean [*Yu et al.*, 1996; *Marchal et al.*, 2000] by comparison, lateral transport of ^{230}Th is minimized due to its shorter residence time (~20–40 yr) [*Nozaki and Nakanishi*, 1985; *Yu et al.*, 1996; *Marchal et al.*, 2000]. Thus, changes in Atlantic MOC are recorded down-core as variations in sediment $^{231}Pa_{xs}/^{230}Th_{xs}$, with vigorous rates of MOC causing a reduction in $^{231}Pa_{xs}/^{230}Th_{xs}$ from the production ratio. The $^{231}Pa_{xs}/^{230}Th_{xs}$ proxy therefore enables the reconstruction of deep ocean circulation changes that goes beyond what traditional proxies such as $\delta^{13}C$ from benthic foraminifera can provide. Benthic $\delta^{13}C$ is driven by end-member variation, water mass chemical "aging" and mixing while $^{231}Pa_{xs}/^{230}Th_{xs}$ does depend more directly on lateral water mass advection, i.e., the physical vigor of the MOC.

2. Material and Methods

[3] We measured sedimentary $^{231}Pa_{xs}/^{230}Th_{xs}$ along sediment core DAPC2 recovered from a high-accumulation drift deposit located in the northern Rockall Trough (58°58.10′N, 09°36.75′W, 1709 m water depth) (Figure 1). The site is presently influenced by Wyville-Thomson Overflow Water (WTOW) from the Norwegian Sea, a precursor to North Atlantic Deep Water (NADW), and recirculated upper NADW [*New and Smythe-Wright*, 2001]. This core provides high-resolution, multi-proxy, paleoceanographic records [*Knutz et al.*, 2002] of iceberg discharge and meltwater variability that indicate a rapid response of ice sheet surges and meltwater from NW Europe over the last deglaciation.

[4] Radiochemical analyses of ^{231}Pa and ^{230}Th were made by isotope dilution using a thermal ionization mass spectrometer for ^{231}Pa and a high-resolution magnetic sector inductively coupled plasma mass spectrometer for ^{230}Th [*Shen et al.*, 2002, 2003]. Diatom abundance as a possible modulator of $^{231}Pa_{xs}/^{230}Th_{xs}$ was quantified following *Scherer* [1995]. The chronology of DAPC2 is constrained by 10 calibrated accelerator mass spectrometry ^{14}C dates (age scale of *Knutz et al.* [2002] augmented by 4 further ^{14}C-AMS dates) that provide a tight temporal framework for the abrupt MOC changes indicated in the records and their linking with sudden changes of surface ocean climatology. The resulting time step along our stable isotope, faunal and ice rafted debris (IRD) records during the last glacial maximum (LGM) and H1 is less than 50 years, mean time step along the $^{231}Pa_{xs}/^{230}Th_{xs}$ is less than 120 years. At this temporal resolution, and because of its geographic location, DAPC2 provides an ideal sedimentary record in which to investigate the sequence of events

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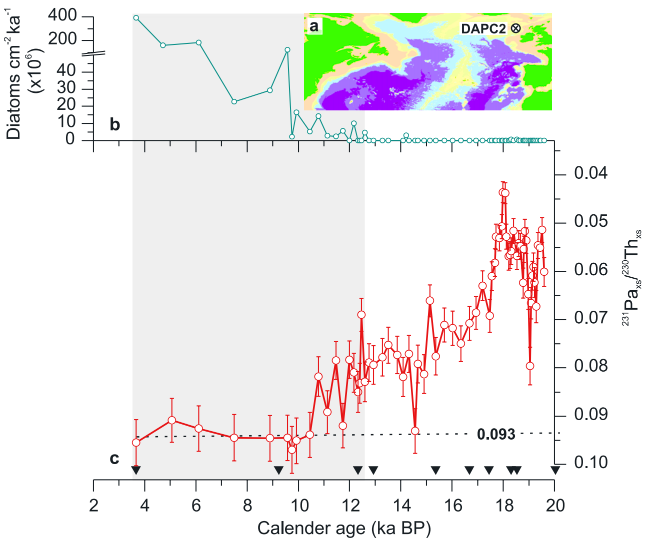

Figure 1. Study area and core DAPC2 data. (a) Map of the North Atlantic showing the location of core DAPC2 from Rockall Trough (58°58.10′N, 09°36.75′W, 1709 m water depth). (b) ^{230}Th normalized biogenic silica flux (diatoms), note axis break (c) sedimentary record of $^{231}Pa_{xs}/^{230}Th_{xs}$. Horizontal dashed line represents the $^{231}Pa/^{230}Th$ production ratio. AMS ^{14}C ages denoted by solid black triangles. Error bars represent uncertainties in ^{231}Pa and ^{230}Th data (2σ), which range from 3–7% and are generally <5%. Shaded area highlights data not considered here for the reconstruction of paleocirculation due to the likely effects of preferential removal of ^{231}Pa by the increased biogenic silica flux.

surrounding a major past freshwater perturbation, the H1 meltwater event, and its consequences for the MOC.

3. Results and Discussion

[5] The $^{231}Pa_{xs}/^{230}Th_{xs}$ record displays (Figures 1 and 2) minimum values during the LGM and a multi-step transition to higher values that includes the Heinrich 1 (H1) meltwater event (H-1$_{MW}$). As non-carbonate clay and carbonate fluxes were high during the LGM (see auxiliary material[1] Figure S1), conditions typically associated with elevated $^{231}Pa_{xs}/^{230}Th_{xs}$ ratios [*Bacon*, 1988], the low $^{231}Pa_{xs}/^{230}Th_{xs}$ ratios prior to ∼18 ka BP imply a vigorous MOC rate and export of ^{231}Pa from this location. At this time δ^{18}O values in DAPC2 of surface (*G. bulloides* [*Schiebel et al.*, 1997]) and deep-dwelling (thermocline depths, *N. pachyderma* sinistral [*Kohfeld et al.*, 1996]) planktonic foraminifera converge suggesting a weakly developed thermocline and decreased vertical stability, indicative of a well mixed ocean upper layer, plausibly marking conditions favorable for convective overturn in the region, in line with elevated benthic δ^{13}C in our core [*Knutz et al.*, 2002]. In contrast, following the Younger Dryas stadial and throughout the Holocene, $^{231}Pa_{xs}/^{230}Th_{xs}$ ratios remain close to the production ratio which we attribute to preferential removal of ^{231}Pa by biogenic silica [*Guo et al.*, 2002; *Luo and Ku*, 2004; *Siddall et al.*, 2005], as evidenced by enhanced diatom accumulation in the uppermost 50 cm (∼12 ka BP) of DAPC2 (Figure 1c). The absence of diatom accumulation at the DAPC2 site during the LGM and early deglaciation indicates that $^{231}Pa_{xs}/^{230}Th_{xs}$ is not dominated by variations in biogenic silica flux during this period.

[6] Fine scale variability is observed in DAPC2 on centennial to sub-centennial time scales (Figure 2). These changes are directly coupled to atmospheric and surface

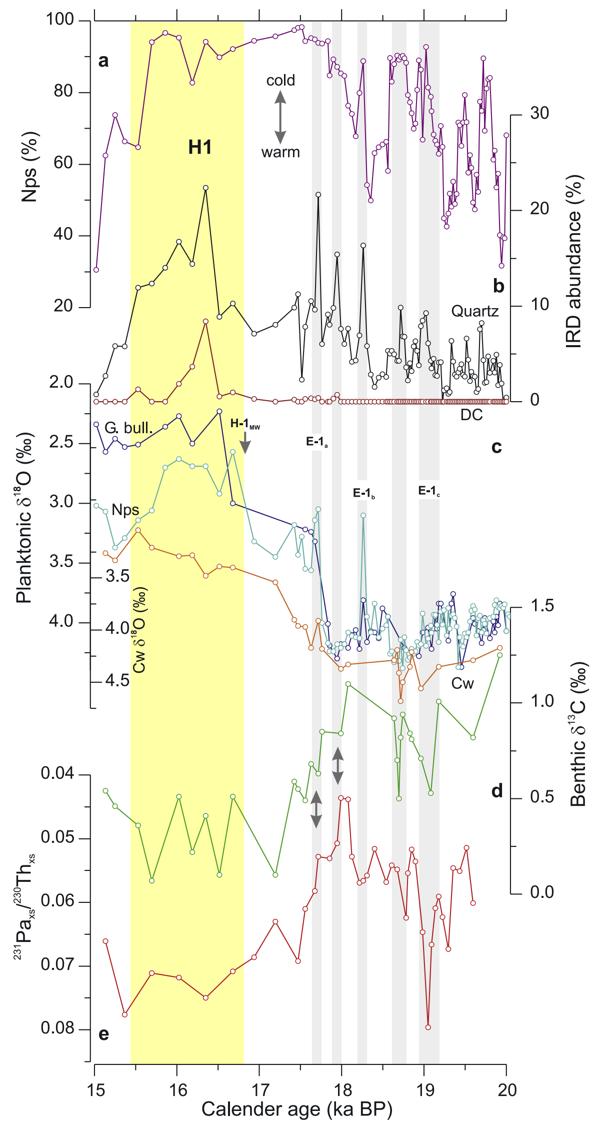

Figure 2. Paleoceanographic time-series from DAPC2 (a) *Neogloboquadrina pachyderma* sinistral (Nps) abundance. (b) Abundance of ice-rafted debris, quartz grains (Quartz, black) and detrital carbonate (DC, brown), in the >250 μm fraction (c) *N. pachyderma* s. (Nps., cyan), *Globigerina bulloides* δ^{18}O (*G. bull.*, blue) and benthic δ^{18}O (*Cibicidoides wuellerstorfi*, Cw, orange). (d) Benthic δ^{13}C (Cw). (e) Sediment $^{231}Pa_{xs}/^{230}Th_{xs}$. Acronyms are: H-1$_{mw}$, Heinrich 1 meltwater event; E-1$_{a}$-$_{c}$, European sourced glacimarine events [*Knutz et al.*, 2002]. These E-events and additional IRD pulses that do not display measurable planktonic δ^{18}O meltwater excursions at this location are highlighted by the grey bands. Heinrich Event 1 (H1) indicated by yellow band. Arrows between d and e mark onset of a multi-step MOC slowdown.

[1]Auxiliary materials are available in the HTML. doi:10.1029/2006GL026239.

ocean (buoyancy) forcing revealed by the planktonic records that document a sequence of transient warming events followed by cooling/freshwater pulses, displayed in brief $\delta^{18}O$ and *N. pachyderma* (sin) abundance anomalies, prior to and during the onset of deglaciation (Figures 2a and 2c). Concomitant IRD peaks (Figures 2a–2c) suggest that the latter are linked with brief but intense pulses of icebergs discharged from the EIS (Figure 2b), previously labeled as E-events [*Knutz et al.*, 2002], most likely triggered by short-lived collapses of marine-based ice margins caused by the slow but steady global warming at the end of the LGM. Similar lithic ''precursor'' events preceding Heinrich events have been reported from North Atlantic core sites and inferred to originate from European and Icelandic glacial sources [*Grousset et al.*, 2000; *Scourse et al.*, 2000]. The DAPC2 $^{231}Pa_{xs}/^{230}Th_{xs}$ record indicates that these events went along with measurable slowdown in MOC. In particular the event centered at $\sim$19.1 ka BP stands out that is associated with a cooling and freshwater pulse, labeled as E-1_c event (Figure 2). The transient MOC slow-down during E-1_c that is evident in $^{231}Pa_{xs}/^{230}Th_{xs}$ is around half the magnitude of the major MOC drawdown starting at $\sim$18 ka BP, although a value similar to those around H1 is reached in a single data point at 19.0 ka. MOC rapidly recovered from the slow down during the E-1_c event, within $\sim$200 years returning to similar rates as before. A sustained step-wise increase in $^{231}Pa_{xs}/^{230}Th_{xs}$ then starts at $\sim$18.0 ka BP, within $\sim$170 yr after the warm pulse that followed the European sourced glacimarine E-1_b event [*Knutz et al.*, 2002]. This signifies the onset of reduced lateral ^{231}Pa escape and a slow-down of deep water export from the region (Figures 2a and 2e). The first step of apparent MOC slow-down is coincident with a centennial scale IRD pulse that suggests a coeval EIS surge event. About one third of the total MOC drawdown, relative to the $\sim$18.0 ka BP $^{231}Pa_{xs}/^{230}Th_x$ maxima, is accounted for in this initial step. Approximately 270 years later, a second step of increasing $^{231}Pa_{xs}/^{230}Th_{xs}$ directly coincides with the incursion of freshwater/IRD pulse E-1_a. This phase of further MOC slow-down is accompanied by a coeval stepwise benthic $\delta^{13}C$ decrease that is also recorded at other core sites in the wider North Atlantic region and has been used, in conjunction with benthic trace element ratios, to infer decreasing chemical ventilation from northern sources and increased advection of southern hemisphere water masses [*Zahn et al.*, 1997; *Willamowski and Zahn*, 2000; *Zahn and Stüber*, 2002; *Rickaby and Elderfield*, 2005]. The lack of equivalent negative benthic $\delta^{18}O$ excursions during the E-$1_{b\text{-}a}$ events supports the view that these events did not cause globally significant changes in ice volume and sea level. The main H1 event lithologically identified in DAPC2 by the presence of a discrete detrital (dolomitic) carbonate IRD layer that is embedded within the quartz IRD deposit [*Knutz et al.*, 2002] and the associated meltwater surge then occur as MOC rates approach peak minimum rates, some 1,200 yr after the onset of the MOC drawdown (Figure 2, H-1_{MW}). $^{231}Pa_{xs}/^{230}Th_{xs}$ ratios throughout the H1 interval remain below production ratio indicating that MOC never achieved a total cessation.

[7] The changes in deep water export traced by $^{231}Pa_{xs}/^{230}Th_{xs}$ in DAPC2 are also recorded at deep subtropical Atlantic sites (core OCE326-GGC5, lower Bermuda Rise, 4550m water depth [*McManus et al.*, 2004]; SU81-18, western Iberian margin, 3135 m water depth [*Gherardi et al.*, 2005]). The inference drawn at these sites was a significant slow-down in MOC coincident with the catastrophic iceberg discharge H1 [*McManus et al.*, 2004; *Gherardi et al.*, 2005] dated to 16.8 ka BP in the North Atlantic [*Hemming*, 2004]. The fine-scale time series of complementary proxies of ice rafted detritus, planktonic $\delta^{18}O$, benthic $\delta^{13}C$, and $^{231}Pa_{xs}/^{230}Th_{xs}$ (Figure 2) in DAPC2 from within the more immediate region of iceberg drift reveals that such MOC decrease occurred significantly before the H-1 event. The fact that all our records are from the same single sediment core ensures that the sequence of events and the temporal offsets between them are robust features independent of age modeling. In particular, we emphasize the similarity between the $^{231}Pa_{xs}/^{230}Th_{xs}$ and benthic $\delta^{13}C$ signals that underscores the significance of the changes in the rate of overturning inferred from $^{231}Pa_{xs}/^{230}Th_{xs}$. Based on our age model the MOC decrease precedes the H1 event by as much as $\sim$1,200 ($^{231}Pa_{xs}/^{230}Th_{xs}$) to $\sim$1,300 yrs (benthic $\delta^{13}C$).

[8] Transient $^{231}Pa_{xs}/^{230}Th_{xs}$increases associated with the E-$1_{a\text{-}b}$ meltwater events suggest a direct link between circulation changes and local European sector derived meltwater perturbations. Taking the E-$1_{a\text{-}b}$ planktonic $\delta^{18}O$ anomalies at face value suggests a salinity reduction of $\sim$0.3–1.0 (freshwater endmember $\delta^{18}O$ -35‰ VSMOW), probably higher if accounting for the effects of sea surface cooling on planktonic $\delta^{18}O$. While salinity anomalies appear close to the magnitude of the $H1_{mw}$ perturbation recorded at the DAPC2 site, at $\sim$90–150 year duration these localized episodes are extremely short-lived. The possibility of additional sources of freshwater input at the time of MOC collapse remains for instance, from the small Iceland Ice Sheet and the much larger Laurentide and Greenland Ice Sheets [*Hagen and Hald*, 2002; *Voelker et al.*, 2000] but the European provenance for the E-$1_{a\text{-}b}$ meltwater perturbations is consistent with the reconstruction of a widespread salinity reduction west of Ireland at 15.0–17.7 ^{14}C ka BP (17.3–20.4 ka BP) that likewise has been linked to European melt sources [*Sarnthein et al.*, 1995]. Moreover, the close linkage of the multi-step $^{231}Pa_{xs}/^{230}Th_{xs}$ increase documented in DAPC2 with the sequence of IRD and negative planktonic $\delta^{18}O$ events in the same core make a compelling case that while additional freshwater sources may have existed elsewhere in the region it were these brief freshwater pulses that played a decisive role in the sustained, but not total, drawdown of meridional overturning. Thus the H1 event and its associated major meltwater surge (H-1_{MW}) appear to have been embedded into an already slowed-down MOC and acted to sustain rather than initiate minimum MOC rates. Such lead of MOC slow down is supported by a similar temporal offset between the initial $^{231}Pa_{xs}/^{230}Th_{xs}$ increase after 19 ka BP in the deep subtropical Atlantic [*McManus et al.*, 2004] and the initiation of surface ocean heat being retained in the tropical [*Rühlemann et al.*, 1999] and subtropical [*Flower et al.*, 2004] Atlantic that each significantly precede the age of H1 event. Continued surface ocean stratification is suggested in DAPC2 in the enhanced offset between the paired planktonic $\delta^{18}O$ records following H1 (Figure 2c) [*Knutz et al.*, 2002] most likely indicating enhanced poleward transport of warm surface waters to the Nordic Sea, via the North Atlantic Drift, and that the location of deep convection has

shifted to north of the Iceland-Scotland Ridge [*Rahmstorf*, 1995], leaving the DAPC2 site under the influence of WTOW and an increasing recirculation of upper deep water [*Knutz et al.*, 2002].

4. Conclusions

[9] Our data demonstrate that catastrophic freshwater surges, e.g., during the past from the disintegrating Laurentide Ice Sheet, are not the sole component involved with major MOC change. Model simulations suggest that the MOC is sensitive to small-scale freshwater perturbations, on order of 0.1 Sv [*Stocker and Wright*, 1991; *Manabe and Stouffer*, 1997]. Our data from core DAPC2 indeed demonstrate that transient localized freshwater episodes occurred between 1,200–1,300 years prior to the large-scale H1 event and coincided with likewise transient phases of increasing $^{231}Pa_{xs}/^{230}Th_{xs}$ and decreasing benthic $\delta^{13}C$, both fully consistent with decreasing deep water ventilation and export. The events formed part of a multi-stepped shift toward a collapsed MOC, lending credence to numerical models that demonstrate the role of targeted freshwater events in forcing the MOC toward and across a threshold from where on convection enters a collapsed mode [*Rahmstorf*, 1994, 1995]. From this it appears that accelerated (yet non-catastrophic scale) melting of the Greenland Ice Sheet [*Zwally et al.*, 2002], may indeed bear significance for future MOC stability and climate in the wider North Atlantic region.

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