Quaternary magnetic and oxygen isotope stratigraphy in diatom-rich sediments of the southern Gardar Drift (IODP Site U1304, North Atlantic)

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Abstract

The sediment sequence from Integrated Ocean Drilling Program (IODP) Site U1304 (53°03.40'N, 33°31.78'W; water depth, 3024 m) on the southern Gardar Drift, North Atlantic Ocean, covers the last ~1.8 Myr with a mean sedimentation rate of ~15 cm/kyr. At Site U1304, paleomagnetic directional and relative paleointensity (RPI) records have been generated for the last ~1.5 Myr, and benthic oxygen isotope data for the last ~1 Myr. The age model for Site U1304 was established by matching $\delta^{18}\text{O}$ and RPI data to calibrated reference records. Prominent intervals of diatom mats at Site U1304 are associated with the latter stages of interglacial marine isotope stages (MIS) 9, 11, 13, 15, 21, 27 and 35, and with the weak glacial of MIS 14. The deposition of diatom mats, and associated diluted magnetic mineral concentrations, leads to weak magnetizations and

susceptibilities in these intervals. Comparison of histograms of paleomagnetic directions and RPI from diatom-rich sediments and from surrounding silty clays indicates that, although results from diatom-rich sediments are more scattered, the occurrence of diatom mats does not appreciably distort the paleomagnetic directional and RPI records. Site U1304 sediments record the Matuyama/Brunhes boundary (~772 ka), the Jaramillo Subchron (~993-1071 ka) and the Cobb Mountain Subchron (~1193-1219 ka), as well as three apparent geomagnetic excursions in the Matuyama Chron, interpreted as the “Kamikatsura/Santa Rosa”, “Punaruu”, and “Gardar” excursions, at ~888 ka, ~1124 ka, and ~1463 ka, respectively. The Site U1304 RPI record can be correlated with the PISO-1500 RPI stack and with other high-resolution RPI records from the North Atlantic Ocean. Wavelet analyses performed on the Site U1304 RPI record and a new high-resolution North Atlantic paleointensity stack for the last 1500 kyr (HINAPIS-1500), comprising Ocean Drilling Program (ODP) Sites 983, 984 and IODP Sites U1306 and U1304, did not reveal significant orbital power.

Keywords: Quaternary; North Atlantic; Relative Paleointensity; Magnetic Excursions; Oxygen Isotopes; Diatoms; IODP Site U1304

1. Introduction

Site U1304 (53°03.40'N, 33°31.78'W; water depth, 3024 m) was drilled in October 2004 from the R/V *JOIDES Resolution* during Integrated Ocean Drilling Program (IODP) Expedition 303. The objective was to obtain a high-resolution environmental record of the last few million years (Myrs) within the central Atlantic ice-rafted debris (IRD) belt,

at a water depth sufficient to monitor past North Atlantic Deep Water (Expedition 303 Scientists, 2006). The site lies in a semi-enclosed basin at the southern end of the Gardar Drift, to the north of the Charlie Gibbs Fracture Zone (Figure 1). Four holes (U1304A, U1304B, U1304C, and U1304D) were cored at Site U1304 to obtain an apparently continuous ~270-m composite sediment sequence covering the last ~1.8 Myrs. With an average sedimentation rate of ~15 cm/kyr for the ~270-m stratigraphic section, Site U1304 represents one of the higher resolution Quaternary deep-sea sediment archives available from the North Atlantic Ocean, although sedimentation rates are highly variable (~5 to >40 cm/kyr) due to fluctuating biogenic (mainly diatom) abundance.

Sediments at Site U1304 consist of inter-bedded, gray-green calcareous nannofossil and diatomaceous oozes with clay and silty clay (Expedition 303 Scientists, 2006). The elevated sedimentation rates at Site U1304 are attributable to the remarkable and unusual occurrence of laminated and massive diatom mats on scales of a few centimeters to tens of meters. The diatomaceous ooze deposition comprises mainly the planktic, araphid, needle-like species *Thalassiothrix longissima*. The diatom-rich intervals are episodic and discontinuous, but are present throughout the entire recovered section back to ~1.8 Ma (Expedition 303 Scientists, 2006; Shimada et al., 2008; Romero et al., 2011). The only analogous documented occurrence of diatom mats in the North Atlantic Ocean is during the last interglacial, marine isotope stage (MIS) 5, in conventional piston Core EW9303-17 (57°N, 37°W; water depth 3233 m), located ~500 km NW of Site U1304 (Figure 1). The occurrence of diatom mats in MIS 5 of Core EW9303-17 was attributed by Bodén and Backman (1996) to its proximity to the Sub-arctic Convergence (SAC), which is a

71 distinct oceanic front between the warm and saline North Atlantic Current and the
72 Irminger Current, and the cold and less saline East Greenland Current and the waters that
73 flow out of the Labrador Sea (Ruddiman and Glover, 1975). Shimada et al. (2008), based
74 on principal component analysis of the diatom assemblage, concluded that the switching
75 between warm- and cold-water diatom assemblages occurred concurrently with diatom
76 mat deposition, which implies that diatom mat deposition was related to migrations of the
77 SAC over the site, analogous to the interpretation of Bodén and Backman (1996). Diatom
78 occurrence at Site U1304 is not simply related to glacial-interglacial cycles (Shimada et
79 al., 2008) and, during the last interglacial, productivity apparently peaked at the MIS
80 5d/5e boundary (Romero et al., 2011). According to Romero et al. (2011), surface waters
81 alone would not have provided the necessary nutrients to sustain the observed diatom
82 productivity at Site U1304. The unique occurrence of diatom mats throughout the
83 Quaternary section at Site U1304 implies episodic incursion of silicate- and nitrate-rich
84 Sub-Antarctic Mode Water (SAMW), as a result of slow-down of Iceland-Scotland
85 Overflow Water (ISOW), and sporadic convective mixing of the water column (Romero
86 et al., 2011).

87
88 During the last few decades, accumulation of relative paleointensity (RPI) data from
89 deep-sea sediments and the construction of RPI stacks (e.g. Guyodo and Valet, 1999;
90 Stoner et al., 2002; Laj et al., 2004; Valet et al., 2005; Channell et al., 2009), as well as
91 the documentation of brief geomagnetic excursions (e.g. Laj and Channell, 2007;
92 Roberts, 2008), have added new information pertaining to the geodynamo, and offer
93 potential for high-resolution stratigraphic correlation within polarity chrons. Sedimentary

RPI records covering the last ~1.5 Myr with mean sedimentation rates above 10 cm/kyr are rare, and include Ocean Drilling Program (ODP) Sites 983 and 984 (Channell et al., 1997, 2002; Channell, 1999; Channell and Kleiven, 2000) and IODP Site U1306 (Channell et al., 2014). In this paper, we report high-resolution (mean sedimentation rate ~15 cm/kyr) Quaternary paleomagnetic directional and RPI records for the last ~1.5 Myr, and benthic oxygen isotope ($\delta^{18}\text{O}$) data for the last ~1 Myr, from diatom-rich deep-sea sediments recovered at IODP Site U1304 (Figure 1). In view of the variable lithology and sedimentation rates at Site U1304, the site provides a rigorous test of the use of RPI records in stratigraphic correlation.

2. Samples and Methods

2.1 Magnetic measurements

U-channels are samples enclosed in plastic containers with an internal $\sim 1.8 \times 1.9 \text{ cm}^2$ cross-section and the same length as the core sections (usually 150 cm), with a clip-on plastic lid that constitutes one of the sides that seals the samples to minimize dehydration (Tauxe et al., 1983; Weeks et al., 1993; Nagy and Valet, 1993). U-channel samples were collected from the archive halves of core sections within the composite splice derived from Holes U1304A, U1304B, and U1304D (Table T21 in Expedition 303 Scientists, 2006), except for occasional intervals (i.e. 74.76-77.76 m composite depth (mcd), and 177.74-182.18 mcd) where sampling switched to core sections not on the composite splice, because the splice in these intervals was found to be affected by coring disturbance. Our u-channel sampling covered the 0-225 mcd interval, except for a ~3.5-m gap in the 197.9-201.4 mcd interval where a thick diatom mat was impossible to sample

without extensive sediment disturbance. Compared with the shipboard data acquired using split half-core sections, u-channel measurements have a number of advantages: (1) u-channels are taken from the center of the split core to avoid disturbed core margins; (2) complete stepwise alternating field (AF) demagnetization can be carried out to define magnetization components; (3) the smaller diameter sample access and the narrower response function of the u-channel magnetometer introduce less convolution/smoothing of sequential measurements; (4) u-channel measurements are made at higher spatial resolution (typically 1-cm spacing) than shipboard measurements (typically 5-cm; Weeks et al., 1993); and (5) laboratory-induced magnetizations can be measured to develop RPI proxies.

Remanent magnetization measurements for Site U1304 u-channel samples were carried out on a 2G Enterprises pass-through superconducting rock magnetometer (SRM) designed to measure u-channel samples at the University of Florida. Measurements for each u-channel were made at 1-cm intervals with a 10-cm “leader” interval and a 10-cm “trailer” interval, immediately before/after the ends of the sample reached/passed the center of the magnetometer sensor region. The “leader” and “trailer” measurements serve the dual functions of monitoring the background magnetic moment and allowing for future signal deconvolution. The natural remanent magnetization (NRM) of u-channel samples was measured before demagnetization, and then after progressive AF demagnetization at peak fields of 20-60 mT in 5 mT increments, 60 to 80 mT in 10 mT increments, and at 100 mT (12 steps in total). For each 1-cm interval, characteristic remanent magnetization (ChRM) directions were calculated using the principal

component analysis (PCA, Kirschvink, 1980) and the UPmag software (Xuan and Channell, 2010). PCA calculations used NRM data acquired during the 20-80 mT demagnetization interval without anchoring directions to the origin of orthogonal projections, and is associated with maximum angular deviation (MAD) values that monitor the quality of the ChRM directions.

After completion of NRM measurements for each u-channel sample, volume magnetic susceptibility (κ) measurements were made at 1-cm interval resolution using a susceptibility bridge designed for u-channel samples. The susceptibility bridge is equipped with a square-shaped ($\sim 3 \times 3$ cm² cross-section) Sapphire Instruments loop sensor to maximize measurement resolution (Thomas et al., 2003). An anhysteretic remanent magnetization (ARM) was then imparted in a 100-mT peak AF and a 50- μ T direct current (DC) bias field, which were both oriented along the long-axis of the u-channel sample. The acquired ARM was measured prior to demagnetization and after stepwise AF demagnetization using the same peak fields used for NRM in the 20-60 mT range. Partial ARMs were then measured during acquisition as the peak AF was increased in 5 mT increments in the 20–60 mT peak field range, in a constant 50- μ T DC bias field. After ARM demagnetization and acquisition experiments, an isothermal remanent magnetization (IRM) was acquired for each u-channel sample in a 300-mT DC field, followed by measurements before and after AF demagnetization with the same peak fields applied to the NRM in the 20-60 mT range. IRMs were also acquired in a 950-mT DC field followed by measurement without demagnetization. Subsequently, an IRM was acquired in a 300-mT DC back-field, opposite to that for the 950-mT IRM, and

measured. The IRM acquired in a 950-mT DC field and in a subsequent 300-mT back-field was used to calculate the S-ratio (Bloemendal et al., 1992): $S\text{-ratio} = [(IRM_{300mT}/IRM_{950mT}) + 1]/2$. S-ratios were calculated for each 1-cm measurement interval, and provide information on concentration changes of high coercivity magnetic components.

NRM intensity, normalized by the intensity of laboratory-induced magnetizations, is often used as a proxy for relative paleointensity (Levi and Banerjee, 1976; King et al., 1983; Tauxe et al., 1993). Normalization is usually carried out using ARM and IRM intensities, or less commonly magnetic susceptibility, to compensate for changes in concentration of remanence carrying grains down-core. Our estimates of RPI proxies for Site U1304 utilize NRM intensity data as well as ARM intensities (both for demagnetization and acquisition) and IRM intensity data from the 20-60 mT demagnetization steps. For each 1-cm interval, RPI proxies were calculated using the slopes of best-fit lines between NRM lost versus ARM lost, NRM lost versus ARM acquired, and NRM lost versus IRM lost (where the slopes of best-fit lines between NRM lost and ARM acquired are multiplied by -1). RPI estimates were determined using the UPmag software (Xuan and Channell, 2009) and each slope calculation was accompanied by a linear correlation coefficient (r-value) that monitors the quality of the line fit.

Excess sediment collected during u-channel sampling was sealed in plastic bags. One bag of bulk sediment was collected for each u-channel sample. These bulk samples were used for magnetic property measurements on a Princeton Measurements Corporation vibrating sample magnetometer (VSM) at the University of Florida. For each bulk sediment

sample, we measured a hysteresis loop and backfield remanence curve to determine ratios of hysteresis parameters: M_r/M_s and H_{cr}/H_c , where M_r is saturation remanence, M_s is saturation magnetization, H_{cr} is coercivity of remanence, and H_c is coercive force.

2.2 *Diatom mat occurrence*

We recorded the occurrence of diatom mats in Site U1304 sediments during sampling by visual inspection of u-channel samples, and from shipboard digital image scans. Diatom-rich sediments have characteristic “oatmeal” texture and often have sharp contacts with inter-bedded silts and clays. For each 1-cm interval of sediment, we assigned a diatom richness value (DRV) of “1” to intervals that comprise predominantly diatom mats. A DRV of “0.5” was assigned to 1-cm intervals that comprise sediment containing disseminated diatoms. For sediment intervals that comprise silts and clays without macroscopic evidence for diatoms, “0” was assigned as the DRV.

2.3 *Oxygen isotopes*

Oxygen isotope ratios ($\delta^{18}\text{O}$) measured on the benthic foraminifera *Cibicidoides wuellerstorfi* have been reported previously for the 23.6-m long (0 to ~125 ka) piston Core KN166-14-JPC13 (hereafter referred to as JPC-13) from the same location as Site U1304 (Fig. 1, Hodell et al., 2010), and for the 15-25 mcd interval (MIS 5) of Site U1304 (Hodell et al., 2009). Oxygen isotope ($\delta^{18}\text{O}$) data for the 25-166 mcd interval at Site U1304 were also measured on specimens of *Cibicidoides wuellerstorfi* at ~5-cm stratigraphic resolution. Specimens of *Cibicidoides wuellerstorfi* were selected from the >150 μm size fraction. Isotope measurements at the University of Florida were made

using a Finnigan-MAT 252 isotope ratio mass spectrometer coupled with a Kiel III carbonate preparation device, and are reported in standard delta notation relative to Vienna Pee Dee Belemnite (VPDB) using NBS-19 for calibration. The estimated analytical error for $\delta^{18}\text{O}$ data is better than $\pm 0.1\text{‰}$. Subsequently, additional stable isotope analyses were performed on a VG SIRA mass spectrometer at the University of Cambridge with a Multicarb system for samples of $>80\text{ }\mu\text{g}$ mass. Analytical precision for the measurements is estimated to be $\pm 0.08\text{‰}$ for $\delta^{18}\text{O}$. For smaller samples ($<80\text{ }\mu\text{g}$) analyzed in Cambridge, measurements were performed on a Thermo Finnigan MAT253 mass spectrometer fitted with a Kiel IV carbonate device. Analytical precision for these samples is estimated to be $\pm 0.08\text{‰}$ for $\delta^{18}\text{O}$. No offset in $\delta^{18}\text{O}$ were observed among the various instruments employed in Florida and Cambridge.

3. Age Model

The age model was constructed for the top ~ 225 mcd of Site U1304 by correlating volume susceptibility (κ), $\delta^{18}\text{O}$, and the RPI proxy, to reference records. For the top ~ 22 mcd, susceptibility data from Site U1304 were correlated to those from Core JPC-13 from the same location (Figure 1). The chronology for the 0-80 ka interval of JPC-13 (Hodell et al., 2010) was based on correlation of millennial scale features in the benthic $\delta^{18}\text{O}$ record of JPC-13 to that of Core MD95-2042 from the Portuguese Margin, which was synchronized with Greenland and Antarctic ice cores (Shackleton et al., 2004). Ages for the 80-125 ka interval of JPC-13 were based on correlation of benthic $\delta^{18}\text{O}$ data of JPC-13 to the LR04 benthic $\delta^{18}\text{O}$ stack (Lisiecki and Raymo, 2005). An unambiguous correlation between the susceptibility records of Site U1304 and JPC-13 was established

using 19 tie points (Table 1) to transfer the age model of JPC-13 to Site U1304 for the last ~125 kyr (Figure 2a, b). The resulting chronology results in a convincing match (Figure 2c) of the JPC-13 RPI record (Evans, 2006) to the Site U1304 RPI record (estimated using the slopes of best-fit lines of NRM lost versus ARM acquired). For the ~22-166 mcd interval, benthic $\delta^{18}\text{O}$ data for the last interglacial at Site U1304 (Hodell et al., 2009) were combined with newer Site U1304 $\delta^{18}\text{O}$ data (Figure 3b), and were correlated to the LR04 stack (Lisiecki and Raymo, 2005) to construct a chronology for Site U1304 that extends to ~1 Ma (Figure 2a). Beyond ~1 Ma, we correlated the Site U1304 RPI record, estimated based on slopes of best-fit lines between NRM lost and ARM acquired, to the PISO-1500 RPI stack, which was constructed by tandem matching and stacking of 13 paired RPI and $\delta^{18}\text{O}$ records from worldwide locations (Channell et al., 2009). The RPI correlation of Site U1304 to PISO-1500 was achieved using the automated matching algorithm (Match) developed by Lisiecki and Lisiecki (2002). The Site U1304 and the PISO-1500 RPI records were normalized to have zero mean and a standard deviation of 1, prior to the “matching”. Site U1304 depth-age tie points, acquired using the above protocol, are listed in Table 1 and plotted on Figure 2a. According to the age model, sedimentation rates at Site U1304 varied from ~5 cm/kyr to >40 cm/kyr over the last ~1.5 Myr (Figure 2a).

4. Results and Discussion

4.1 NRM and Geomagnetic Reversals and Excursions

Site U1304 cores from which the u-channel samples were taken are shown in Figure 3a. ChRM declinations and inclinations, with associated MAD values, are plotted versus

depth (mcd) in Figures 3c-e together with the benthic $\delta^{18}\text{O}$ data (Figure 3b). The MAD values are largely below 1° and have a mean of $\sim 0.8^\circ$ for samples from the non-diatom-rich sediment intervals (DRV=0). For samples from diatom-rich sediment intervals (DRV=0.5 or 1, vertical gray bars in Figure 3), MAD values are generally below 5° and have a mean of $\sim 1.7^\circ$. These generally low MAD values suggest that ChRM directions are well defined in both non-diatom-rich and diatom-rich sediments, although diatom-rich sediment intervals appear to yield generally noisier ChRM directions (larger MAD values). Representative demagnetization behaviors for non-diatom-rich and diatom-rich sediment intervals are depicted in the top two rows of Figure 4. AF demagnetization using a maximum peak field of 100 mT typically reduced initial NRM intensities of non-diatom intervals to $<5\%$ of the initial value. Orthogonal projections of non-diatom-rich sediments (top row in Figure 4) indicate single magnetization components carried by a low-coercivity magnetic mineral. For diatom-rich sediments, occasionally $>10\%$ NRM intensity remained after 100 mT peak AF demagnetization (second row in Figure 4), which implies the presence of a high-coercivity remanence carrier. Although NRM data from diatom-rich intervals are apparently more scattered than from the non-diatom intervals, probably due to their lower magnetization intensities, NRM demagnetization data from both diatom-rich and non-diatom intervals still fall largely on a straight line that defines a single magnetization component (Figure 4). Magnetization intensities and susceptibility are usually at least one order of magnitude greater for non-diatom-rich versus diatom-rich intervals (Figures 4 and 5).

277 ChRM declinations (light blue points in Figure 3c) were corrected by uniform rotation of
278 each core so that the mean of the corrected declinations (green points in Figure 3c) is
279 oriented north or south for positive and negative inclination intervals, respectively.
280 ChRM inclinations and corrected declinations of Site U1304 can be correlated
281 unambiguously to the geomagnetic polarity time scale (GPTS), to recognize the
282 Matuyama/Brunhes boundary (~138.98 mcd), the Jaramillo Subchron (164.63-176.04
283 mcd), and the Cobb Mountain Subchron (188.17-191.54 mcd) (Figure 3c and d). In
284 addition, three possible magnetic excursions were identified at Site U1304 during the
285 Matuyama Chron at ~149.23-149.34 mcd (Excursion I), ~182.96-183.02 mcd (Excursion
286 II), and ~220.75-221.18 mcd (Excursion III). In Table 2, we list the depths for the
287 polarity reversals and excursions recorded at Site U1304 (including top, bottom, and
288 center mcd) together with their corresponding age estimates according to the age model
289 (Table 1) and the marine isotope stages in which they occur. To identify possible
290 excursion intervals, the following criteria were used: (1) NRM demagnetization data of
291 the samples plotted on orthogonal projections show magnetization component with
292 inclination deviating $>\sim 45^\circ$ from the expected values within a polarity chron (bottom
293 three panels in Figure 4); (2) there is no record of sampling disturbance or visible
294 changes in lithology; (3) similar paleomagnetic directional changes can be observed in
295 shipboard data and/or in samples from other holes over the same depth (mcd) interval; (4)
296 RPI estimates across the excursion interval have low values compared to surrounding
297 intervals; and (5) deconvolution of measurement data using UDECON (Oda and Xuan,
298 2014; Xuan and Oda, 2015) suggests that anomalous directions are not due to artifacts of
299 negative lobes in the SRM sensor response (Roberts, 2006). An example comparison of

ChRM directions (for core section U1304B-18H-3A) calculated using original and deconvolved measurement data is shown in Supplementary Figure 1.

Excursion I (149.23-149.34 mcd) corresponds to the MIS 22/23 transition (Figure 3) and is dated to between 885.84 and 889.22 ka (center age ~887.97 ka) (Table 2). This part of the Site U1304 age model is based on correlation of benthic $\delta^{18}\text{O}$ to the LR04 stack. Excursion I is recorded in core sections U1304A-14H-5 (at ~20 cm) and U1304B-15H-2 (near section top). ChRM directions (estimated from 20-80 mT interval) are significantly shallower than values from surrounding intervals, and have large MAD values (Figure 3). NRM demagnetization data in this interval have an apparent positive inclination component at ~55-100 mT peak AF demagnetization, with a shallow negative inclination overprint in the ~20-55 mT demagnetization interval (third row, Figure 4). Excursion I occurs between the top of the Jaramillo Subchron and the Matuyama/Brunhes boundary. Only two geomagnetic excursions have been documented within this time interval (Laj and Channell, 2007): the Santa Rosa excursion and the Kamikatsura Excursion with reported $^{40}\text{Ar}/^{39}\text{Ar}$ ages of ~936 ka and ~899 ka (Singer et al., 2004), respectively. The Site U1304 benthic $\delta^{18}\text{O}$ based age of ~888 ka for Excursion I is close to that of the Kamikatsura Excursion, which was first documented in the Kamikatsura Tuff of the Osaka group in SW Japan (Maenaka, 1983), and has been reported in loess sections from China (Yang et al., 2004; Wang et al., 2010; Ao et al., 2012; Liu et al., 2015), and in marine sediments from the equatorial Pacific Ocean (Laj et al., 1996). An interval of anomalously low virtual geomagnetic pole (VGP) latitudes also occurs at ODP Site 983 in MIS 21 at ~850 ka (Channell and Kleiven, 2000). Because of poor age control in

some of the marine sediment records (e.g. Laj et al., 1996), and delayed NRM acquisition in Chinese loess (e.g. Spassov et al., 2003) the existence of the Kamikatsura Excursion remains controversial. Singer et al. (1999) and Coe et al. (2004) considered that intermediate directions in the Haleakala Caldera (Maui), immediately below the Matuyama-Brunhes boundary, record the Kamakatsura excursion, although this is also controversial. The Santa Rosa excursion is, however, better defined both in igneous rocks (Singer and Brown, 2002) and in sedimentary sequences from the North Atlantic Ocean (Channell et al., 2002) and from the western Pacific Ocean (Hornig et al., 2002, 2003).

Excursion II (~182.96-183.02 mcd) occurs between the top of the Cobb Mountain Subchron and the base of the Jaramillo Subchron (Figure 3), with an estimated age of ~1124 ka (Table 2). This excursion is recorded in core section U1304B-18H-3 (at ~80 cm). ChRM inclinations are significantly shallower (fourth row of Figure 4) than values in surrounding intervals and have large MAD values (Figure 3). The estimated age of ~1124 ka for Excursion II is close to the previously reported Punaruu Excursion from basaltic lava flows in the Punaruu Valley, Tahiti (Chauvin et al., 1990). $^{40}\text{Ar}/^{39}\text{Ar}$ methods yielded estimated ages of ~1105 ka (Singer et al., 1999) and ~1122 ka (Singer et al., 2004) for the Punaruu Excursion. Marine sediments from California Margin ODP Site 1021 (Guyodo et al., 1999), and North Atlantic ODP Sites 983 and 984 (Channell et al., 2002) and IODP Sites U1308 (Channell et al., 2008; 2016) also record this excursion, with estimated ages of 1110 ka (ODP Site 1021), 1115 ka (ODP Site 983 and 984) and 1092 ka (IODP Site U1308).

Excursion III (~220.75-221.18 mcd) is recorded below the Cobb Mountain Subchron with an estimated age of ~1463 ka (Table 2). This excursion is recorded in core sections U1304B-22H-2 (~100 cm) (Figure 3; and bottom row, Figure 4), U1304A-22H-1, and U1304D-16H-6. In Supplementary Figure 2a and b, we compare shipboard NRM inclinations and declinations from intervals that record the excursion, along with u-channel data from Core U1304B-22H. Data from Cores U1304A-22H and U1304D-16H were shifted down-section by 20 cm and 5 cm, respectively, relative to Core U1304B-22H to optimize the correlation of shipboard susceptibility (Supplementary Figure 2c) and natural gamma radiation data (Supplementary Figure 2d). The three records correlate satisfactorily, notwithstanding erratic declinations in the excursion interval (Supplementary Figure 2a and b). The estimated age of ~1463 ka for Excursion III is close to the reported age of ~1472-1480 ka for the Gardar Excursion, which was first documented at ODP Sites 983 and 984 (Channell et al., 2002). The Gardar Excursion has also been reported in (paleo) lake sediments from SW China (An et al., 2011) and in a loess section from North China (Ao et al., 2012).

4.2 Diatom Deposition and its Influence on Paleomagnetic records

Rock magnetic data (κ , ARM, and IRM) are shown in Figure 5 together with the spliced core image of the sampled composite section, shipboard counts of diatom percentage per core section (Expedition 303 Scientists, 2006) and abundance of the dominant diatom species *Thalassiothrix longissima* (Shimada et al., 2008). Susceptibility (κ), as well as ARM and IRM intensities indicate the concentration of magnetic minerals (mainly magnetite), and are sensitive to different grain-size populations of magnetic grains.

Variations in these parameters at Site U1304 are closely related to diatom abundance. Susceptibility, ARM, and IRM values from diatom-rich sediment (vertical gray bars in Figure 5) are usually 1->2 orders of magnitude lower than those from non-diatom-rich intervals. In addition, within diatom-rich intervals with higher diatom mat abundances (DRV=1, vertical dark gray bars in Figure 5), we observe lower values of susceptibility, ARM, and IRM. For instance, the lowest values of these magnetic parameters occur at ~44-49, 55-60 mcd, 94-107 mcd and 194-203 mcd where the thickest diatom mats are observed (Figure 5). S-ratios also co-vary with diatom occurrence (Figure 5f), with diatom-rich intervals often having lower values of ~0.9 while non-diatom-rich sediment has values close to 1. The lower S-ratios suggest that diatom-rich sediment contains higher coercivity magnetic minerals. This is consistent with NRM demagnetization behavior (second row, Figure 4), where samples from diatom-rich intervals often have more NRM remaining after 100-mT demagnetization compared to non-diatom-rich intervals (top row, Figure 4).

Hysteresis ratios on a Day et al. (1977) plot are shown in Figure 6a. Data symbols for the hysteresis ratios are color-coded by the mean value of susceptibility (κ) within the core section, which can be used to estimate diatom-richness. All hysteresis ratio data lie in the pseudo single domain (PSD) region on the Day et al. (1977) plot with a distribution elongated along a typical magnetite grain-size mixing trend. No apparent difference in magnetic grain size is observed between samples from diatom-rich (low κ) and non-diatom-rich intervals (high κ). Hysteresis ratios from the low κ (yellow) intervals appear to lie on a slightly different grain-size mixing line toward higher H_{cr}/H_c values, which is

consistent with the presence of a higher-coercivity magnetic mineral in the diatom-rich sediments. Representative hysteresis loops and back-field remanence measurements for non-diatom-rich samples have typical characteristics of a low-coercivity magnetic mineral with a paramagnetic component (Figure 6b, c). Hysteresis loops and back-field remanence data for diatom-rich sediments are noisier due to the low magnetic mineral concentration (Figure 6d, e), and the saturation field and remanent coercivity (H_{cr}) are apparently larger than those of non-diatom-rich sediments.

To test if the low concentration and presence of high-coercivity magnetic minerals in diatom-rich intervals influences paleomagnetic directions, magnetization intensities and RPI records, we compare histograms of paleomagnetic data from non-diatom-rich and diatom-rich sediments (Figure 7). NRM intensities (after 35-mT demagnetization) of diatom-rich samples have distinctly different distribution from non-diatom-rich samples with variable value from below 10^{-4} A/m to $\sim 10^{-1}$ A/m (Figure 7a). Despite the apparent difference in NRM intensity, ChRM directions (Figure 7b, c) from non-diatom-rich and diatom-rich samples are comparable to each other. Corrected ChRM declinations from both non-diatom-rich and diatom-rich intervals are around 0° (or 360°) and 180° with a similar distribution, although data from diatom-rich intervals appear to have slightly greater spread over the data range. Similarly, ChRM inclinations from non-diatom-rich and diatom-rich sediments have comparable distributions for normal (positive inclinations) and reverse (negative inclinations) polarity chrons. ChRM inclinations from diatom-rich intervals appear to be slightly shallower (by $\sim 4^\circ$ - 5°) and more scattered than

for non-diatom-rich intervals (mean values of ChRM inclinations are $\sim 64.7^\circ$ and $\sim 60.6^\circ$ for non-diatom-rich sediments, and $\sim 60.7^\circ$ and $\sim 55.1^\circ$ for diatom-rich sediments).

Histograms of r-values (linear correlation coefficients) associated with RPI estimates for slopes of best-fit lines between NRM intensity lost versus (1) ARM lost, (2) ARM acquired, and (3) IRM lost are shown in Figure 7d. The optimal normalizer for RPI proxies should activate the same population of magnetic grains that carry the NRM (e.g. Levi and Banerjee, 1976), so that demagnetization or acquisition behaviour of the normalizer should mimic that of the NRM demagnetization, with a similar proportion of magnetization lost or acquired between the same treatment levels, resulting in r-values close to 1. RPI estimates using NRM lost versus ARM acquired appear to be associated with the largest r-values that are close to 1 (green vertical bars in Figure 7d). Using only r-values that are larger than 0.98, mean r-values are 0.9948, 0.9984, and 0.9967 for RPI estimates using NRM lost versus ARM lost, NRM lost versus ARM acquired, and NRM lost versus IRM lost, respectively.

RPI estimates and associated r-values using NRM lost versus ARM lost, and NRM lost versus ARM acquired are plotted against mcd in Figure 8. The two RPI estimates are comparable (bottom panel, Figure 8), with r-values for NRM lost versus ARM acquired (red points, top panel in Figure 8) having consistently higher (closer to 1) values for almost the entire studied depth interval. We, therefore, estimated RPI using slopes of best-fit lines of NRM lost versus ARM acquired for Site U1304. Histograms of ARM intensity (after 35-mT demagnetization with 50- μ T DC bias field) and RPI estimates

437 using NRM lost versus ARM acquired for both non-diatom-rich and diatom-rich samples
438 are compared in Figure 7e and f. Similar to the NRM intensity data (Figure 7a),
439 distributions of the ARM intensity data for non-diatom-rich and diatom-rich samples are
440 different, with data from non-diatom-rich samples having a broader distribution of values
441 (Figure 7e). Despite distinct distributions of NRM and ARM intensity data, RPI estimates
442 from the non-diatom-rich and diatom-rich samples have similar distributions and both
443 have mean values around 0.35 (Figure 7f), which suggests that although diatoms decrease
444 the magnetic mineral concentration, RPI proxies are not affected significantly. We also
445 note that RPI from diatom-rich sediments contain slightly more values >1 compared with
446 RPI estimations from non-diatom-rich sediments that contain few such values. These
447 large RPI values in diatom-rich intervals give rise to spikes in Figure 8, and could be
448 related to inadequate normalization. We further refined our RPI record by discarding RPI
449 estimates, based on NRM lost versus ARM acquired, using two criteria: (1) r -values
450 <0.98 ; or (2) raw RPI (slope) values >1 . The resulting RPI record for Site U1304 against
451 age is shown in Figure 9f. The satisfactory correlation between the Site U1304 RPI
452 record and the PISO-1500 RPI stack (Channell et al., 2009) for both the non-diatom-rich
453 and the diatom-rich intervals (Figure 9f) further suggest that RPI estimates are not
454 significantly degraded by diatom-rich intervals. In addition, no apparent correlation is
455 observed between RPI (Figure 9f) and the “normalizer” ARM; Figure 9a).

456
457 The close relationship between magnetic concentration dependant parameters and
458 diatom-richness means that magnetic data can potentially serve as a high-resolution bulk
459 diatom abundance proxy at Site U1304. Reduced magnetic mineral concentration during

diatom-rich intervals is attributed to dilution caused by rapid diatom deposition. Diatom-rich sediments generally have elevated mean sedimentation rates of ~20 to over 30 cm/kyr (Figure 9c). Observation of laminated diatom deposits in the deep-sea was first documented in Neogene sediments of the eastern equatorial Pacific Ocean (Kemp and Baldauf, 1993) and later in Pleistocene sediments at ODP Site 1093 in the South Atlantic (Grigorov et al., 2002). Apart from IODP Site U1304 (Expedition 303 Scientists, 2006; Shimada et al., 2008; Romero et al., 2011), the only report of laminated diatom-rich sediments in the deep North Atlantic Ocean was during MIS 5 in a short piston core EW9303-17 (Bodén and Backman, 1996), located ~500 km NNW of Site U1304 (Figure 1). Accumulations of laminated diatom-rich sediments in the deep-sea are often associated with the convergence of major oceanic frontal systems. Bodén and Backman (1996) explained the occurrence of laminated diatom-rich sediments in Core EW9303-17 as related to the position of the subarctic convergence where the warm and saline North Atlantic and Irminger Currents meet the cold and less saline Labrador Sea water. This hypothesis was supported by Shimada et al. (2008) in explaining the occurrence of laminated diatom-rich sediments at Site U1304. Romero et al. (2011) argued that the subarctic convergence alone is insufficient to provide the necessary nutrients to sustain the observed diatom productivity at Site U1304, and that incursions of silicate- and nitrate-rich subantarctic mode water due to slowdown of Iceland-Scotland overflow water, and sporadic convective mixing of the water column might have played a decisive role.

Based on a chronology constructed using shipboard polarity stratigraphy and biostratigraphy, Shimada et al. (2008) noted that diatom occurrence at Site U1304 is not simply related to glacial-interglacial cycles. For the last interglacial, Romero et al. (2011) observed peak abundance of diatom mats at the MIS 5d/5e boundary at ~116 ka. According to the updated age model (Table 1), diatom mat deposition tends to occur late in interglacial stages and during glacial inception, rather than in peak interglacials, but not in all interglacials, as well as in the weak glacial of MIS 14 (Figure 9). Minor deposition of diatom mats occurred in some interglacials (e.g. MIS 49, 37, 33, 31, 23, 19, and 17). For the last ~1.5 Myr, the most prominent episodes of diatom mat deposition at Site U1304 occurred at around MIS 39, during MIS 35, 27, 21, between MIS 15 and 13, including MIS 14, and during MIS 11 and 9. The increase in the occurrence of diatom mats during the middle Brunhes Chron (between MIS 9 and 15) at Site U1304 may coincide with an increase in pelagic carbonate production at this time (Barker et al., 2006).

Preservation of laminated diatom mats often requires inhibition of benthos by low dissolved oxygen levels either in anoxic basins (e.g. Schimmelmann and Lange, 1996) or beneath areas of coastal upwelling (e.g. Brodie and Kemp, 1994). On the other hand, where the deep-sea was well oxygenated, preservation of laminated diatom mats cannot be attributed to low oxygen levels. The absence of bioturbation in laminated intervals can be explained by physical suppression of benthic faunal burrowers because of the high tensile strength of diatom mats (Kemp and Baldauf, 1993). High rates of sediment accumulation and low bioturbation in diatom mats result in high stratigraphic resolution.

We speculate that the lower density (Expedition 303 Scientists, 2006) and mesh-like structure of the diatom mats, and the physical suppression of benthic activity beneath the laminated intervals promoted a slightly more oxic environment within diatom-rich intervals shortly after deposition, whereby fine PSD magnetite particles experienced certain degree of oxidation that produced higher coercivity magnetic minerals such as maghemite (Kawamura et al., 2012). This process would explain the presence of high coercivity magnetic components in the diatom-rich sediments as indicated by the NRM demagnetization behavior, S-ratios, and hysteresis parameters. Compaction, de-watering and the lack of bioturbation of diatom-rich sediments after deposition may have led to the slightly shallower inclinations ($\sim 4\text{-}5^\circ$) observed in the diatom-rich sediments (Figure 7c).

4.3 High-resolution North Atlantic Paleointensity Stack (HINAPIS-1500) and Wavelet Analyses of RPI Records

In Figure 10a, we compare the Site U1304 RPI record with other individual RPI records that cover the last 1.5 Myr with mean sedimentation rates >10 cm/kyr. Only three such records exist and all are from the northern North Atlantic Ocean (Figure 1): ODP Sites 983 and 984 (Channell et al., 1997, 2002; Channell, 1999; Channell and Kleiven, 2000), and IODP Site U1306 (Channell et al., 2014). RPI records from ODP Sites 983 and 984, and IODP Sites U1304 and U1306 were used to construct a high-resolution stack record. Construction of the RPI stack (here named “HINAPIS-1500”) follows new procedures from those used for PISO-1500 (Channell et al., 2009), SINT-800 (Guyodo and Valet, 1999), and SINT-2000 (Valet et al., 2005). To allow for more realistic uncertainty estimates, individual records are placed on their published age models without

correlation/matching to a target record or interpolation of records onto a common age model. All records were normalized to their means for the same 0-1500 ka time interval. This procedure is necessary to account for differences in normalization parameter/method used for RPI estimation, and for possible differences in paleointensity at different locations. The new stack was constructed for time-steps between 0 and 1500.5 ka at 0.5 kyr intervals (i.e. 0.5, 1, 1.5, ..., 1499, 1499.5, 1500). For each time step center (t_i), RPI values were randomly drawn (10,000 times with a uniform distribution) within the $t_i-0.5$ to the $t_i+0.5$ time interval (1-kyr duration) from the four records (2,500 times from each record to ensure equal weight for each record). This process allows a 0.5 kyr overlapping interval within which RPI values are drawn when constructing the stack for each adjacent time step. Chronological uncertainty certainly exceeds 0.5 kyr. Allowance of a 0.5-kyr overlapping interval for stacking is therefore justified while making available more RPI values for the bootstrap experiment at each time step (usually ~5 to over 20 data points from each record in the stack; Figure 10c). For each time step, the mean of the 10,000 (or less for occasional intervals where some records do not have RPI values) randomly drawn RPI values was used to produce the stack (red curve in Figure 10b). We estimated the 90% confidence level (vertical gray intervals in Figure 10b) for each stacked value using cut-off levels of the top and bottom 5% of randomly drawn RPI values at each time step.

The HINAPIS-1500 RPI stack compares well with the published PISO-1500 RPI stack on time scales of a few thousand to tens of thousands of years (Figure 10b). The apparent mismatch between ~630 and 700 ka is probably caused by age model discrepancies. Age

models for ODP Sites 983 and 984, and IODP Site U1304 during this interval are based on oxygen isotope stratigraphy, while the IODP Site U1306 record for this interval was dated through tandem correlation of both RPI and oxygen isotope to the PISO-1500 and the LR04 stacks (Channell et al., 2014). The ODP Site 983 and 984 RPI records used in the HINAPIS-1500 stack were also used for PISO-1500 RPI stack construction, although the PISO-1500 stack included 11 other records including 2 from the western equatorial Pacific Ocean, 1 from the Indian Ocean, and 1 from the South Atlantic Ocean. Within the 0-1500 ka time interval, 9 out of the 13 records in PISO-1500 spanned ~500 kyr or longer, among which 6 spanned ~1000 kyr or longer. It is possible that spatial bias of the records in PISO-1500 toward the North Atlantic (9 of 13) may contribute to the observed correlation between HINAPIS-1500 and PISO-1500, which are both dominated by common paleointensity variability in the North Atlantic region. The PISO-1500 RPI stack fits within the 90% confidence intervals of the HINAPIS-1500 record (Figure 10b). The observation that a regional (northern North Atlantic) RPI stack (i.e. HINAPIS-1500) can be correlated satisfactorily to another stack built using globally distributed records (i.e. PISO-1500) suggest that RPI variations are global on times scales of a few tens of thousand years, and therefore can be used as a valuable high-resolution and independent global stratigraphic correlation and dating tool.

To further investigate common variability among the individual records and RPI stacks, and the time-frequency domain characteristics of the records, wavelet analyses including local wavelet power spectra (LWPS), global wavelet power spectra (GWPS), and wavelet coherence (WTC) (Torrence and Compo, 1998; Torrence and Webster, 1999; Grinsted et

al., 2004) were performed on the Site U1304 RPI record, and the HINAPIS-1500 and PISO-1500 stacks, and on other individual records used in the HINAPIS-1500 stack. During the last two decades, orbital cycles with ~100 kyr and/or ~41 kyr periods, have been reported in RPI records from marine sediments and have been considered evidence for orbital forcing on the geodynamo (e.g. Channell et al., 1998; Yamazaki, 1999; Yamazaki and Oda, 2002; Thouveny et al., 2004; 2008). On the other hand, wavelet analyses of some of these records indicated that orbital periods in the RPI record are possibly due to lithological variations (Guyodo et al., 2000) or are not statistically significant (Roberts et al., 2003). Based on analyses of multiple records from both the Atlantic and Pacific Oceans, Xuan and Channell (2008) suggested that orbital periods are introduced into NRM records through orbital control on subtle lithologic variations. Our analyses of the Site U1304, HINAPIS-1500, and PISO-1500 RPI data using LWPS (Supplementary Figure 3a-c) indicate little significant (1% level against red noise) power at orbital periods in these records. The apparent high power for the 41-kyr period at ~700 ka and ~1100 ka in the HINAPIS-1500 stack (Supplementary Figure 3b) are likely due to strong obliquity power in RPI data from ODP Site 983 during these time intervals (Figure 4a in Xuan and Channell, 2008). The lack of significant orbital power in the analyzed RPI records is consistent with the suggestion that orbital periods observed in some individual RPI records are due to lithological contamination (Guyodo et al., 2000; Xuan and Channell, 2008).

The definition of squared WTC closely resembles that of a traditional correlation coefficient, and it is useful to think of the WTC as a localized correlation coefficient in

597 time-frequency (or -period) space (Grinsted et al., 2004). Squared WTC between the Site
598 U1304 RPI record and the HINAPIS-1500 stack (Supplementary Figure 3g) indicate
599 significant coherence at periods ranging from ~8 kyr to over 500 kyr during various time
600 intervals. The two records appear to be coherent with one another at periods of ~64-100
601 kyr and at around 256 kyr, for majority of the 0-1500 ka time interval. Similar WTC is
602 observed between the Site U1304 RPI record and the PISO-1500 stack (Supplementary
603 Figure 3i). The fact that the two records are more coherent with each other beyond ~1000
604 ka is probably because the Site U1304 age model beyond 1000 ka was based on
605 automated matching of Site U1304 RPI to the PISO-1500 stack. During the ~0-300 ka
606 time interval, Site U1304 RPI has a larger and more significant coherence with the
607 HINAPIS-1500 stack than the PISO-1500 stack on periods ranging between ~8 kyr and
608 ~20 kyr. This is reflected in the lack of coherence on the WTC map between the PISO-
609 1500 and the HINAPIS-1500 stack (Supplementary Figure 3h) over the same period.
610 Significant coherence between the PISO-1500 and HINAPIS-1500 stacks over the
611 majority of the time-period space at periods above ~40 kyr is consistent with the
612 observed correlation of the two records in the time domain (Figure 10b). Significant
613 WTC of the two stack records can also be seen for periods between ~8 and 40 kyr for
614 multiple time intervals (Supplementary Figure 3h). The lack of significant WTC between
615 any of the three records for periods of <~8 kyr is likely related to chronological
616 uncertainties in these records, which could reach a few kyr and affect analyses of short-
617 term variability. The two period ranges over which all three RPI records appear to be
618 coherent with each other are periods between ~64 and 100 kyr and around 256 kyr. These
619 two ranges are also dominant peaks in the global wavelet power spectra (GWPS) of these

records (Supplementary Figure 3d-e), which sum the wavelet power along the time axis for each frequency (or period) component. The coherent high power of variability over the two ranges of periods appear to be a common feature for RPI records and could potentially relate to mechanisms that drive the geodynamo.

5. Conclusions

An age model for the sediment sequence at IODP Site U1304 for the last ~1.5 Myr was based on: (1) correlation of Site U1304 magnetic susceptibility to that of Core JPC-13, from the same location, for the 0-125 ka interval, and use of the benthic $\delta^{18}\text{O}$ -based age model of Core JPC-13 (Hodell et al., 2010); (2) correlation of the Site U1304 benthic $\delta^{18}\text{O}$ record to the LR04 $\delta^{18}\text{O}$ stack from ~125 ka to ~1 Ma; and (3) automated matching of the Site U1304 RPI record to the PISO-1500 RPI stack for the ~1-1.5 Ma interval. ChRM declination and inclination data from u-channel measurements reveal unambiguously the Matuyama-Brunhes boundary (at ~138.98 mcd), the Jaramillo Subchron (at ~164.63-176.04 mcd), and the Cobb Mountain Subchron (at ~188.17-191.54 mcd), with estimated ages of ~772 ka, ~993-1071 ka, and ~1193-1219 ka, respectively. Three geomagnetic excursions are observed at Site U1304 in the Matuyama Chron at ~888 ka, ~1124 ka, and ~1463 ka. These magnetic excursions may be correlated to previously documented “Kamikatsura/Santa Rosa”, “Punaruu”, and “Gardar” excursions with similar reported ages.

Diatom mats at Site U1304 are associated with elevated sedimentation rates, and generally occur in the latter part of certain interglacial stages and during glacial

inceptions. Prominent episodes of diatom mat deposition occurred in the latter part of MIS 39, 35, 27, 21, between MIS 15 and 13, including the weak glacial MIS 14, and during the latter part of MIS 11 and 9. As expected, rapid deposition of diatom mats at Site U1304 appears to have substantially diluted the concentration of magnetic minerals, leading to much lower volume susceptibility, ARM intensity, and IRM intensity in diatom-rich intervals. A high coercivity magnetic component is preserved in the diatom-rich sediments, possibly maghemite. Maghematization could have occurred due to oxidation of fine PSD magnetite in diatom-rich layers due to a slightly more oxic environment created by a combination of lower density mesh-like diatom mats and physical suppression of benthic activity within diatom-rich layers shortly after deposition. Despite the much weaker magnetizations, and possible occurrence of high coercivity minerals, comparison of histograms of paleomagnetic data from both non-diatom-rich and diatom-rich sediments indicates that the diatom mats do not significantly degrade the fidelity of paleomagnetic directions or RPI proxies. Slightly shallower inclinations (~ 4 - 5°) in the diatom-rich sediments could be related to compaction, de-watering, and lack of bioturbation of diatom-rich sediments.

The Site U1304 RPI record correlates with the PISO-1500 RPI stack and with three other North Atlantic records (ODP Sites 983 and 984, and IODP Site U1306) that span the last 1500 kyr at similar resolution (i.e. mean sedimentation rates >10 cm/kyr). These three records were used together with the Site U1304 RPI record to build a high-resolution reference paleointensity stack (HINAPIS-1500) for the northern North Atlantic region, following a new stacking method. The HINAPS-1500 stack matches the PISO-1500 stack

on time scales greater than a few thousand years. Wavelet analyses of the Site U1304 RPI record, HINAPIS-1500, and PISO-1500, indicate no statistically significant power at orbital periods, in agreement with previous suggestions that orbital periods observed in some individual RPI records are due to lithological contamination rather than to orbital forcing of the geodynamo. Squared wavelet coherence and global wavelet power spectra of individual RPI records and stacks have very high coherence and global power at period bands of ~64-100 kyr and around 256 kyr, which might constitute a common feature associated with RPI variations.

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Figure Caption

Figure 1. Location of IODP Sites U1304 (this study), U1306, and U1308, Core EW9303-17 studied by Bodén and Backman (1996), and ODP Sites 983 and 984. Site U1304 and Core KN166-14-JPC-13 are at the same location. Dashed orange and blue lines depict major surface and deep currents, respectively, in the North Atlantic Ocean. The Subarctic Convergence (Ruddiman and Glover, 1975) is marked by a thick dashed black line. DSOW = Denmark Strait Overflow Water; ISOW = Iceland-Scotland Overflow Water; NAC = North Atlantic Current; CGFZ = Charlie Gibbs Fracture Zone.

Figure 2. (a) Site U1304 depth versus age and resulting sedimentation rates (gray lines). Age models were developed by a combination of correlating Site U1304 volume magnetic susceptibility to that of Core KN166-14-JPC-13 (green triangles), correlating Site U1304 benthic $\delta^{18}\text{O}$ to LR04 stack (red circles), and automated matching of Site U1304 RPI to the PISO-1500 RPI stack (blue symbols). Depth-age ties are listed in Table 1. (b) Correlation of Site U1304 volume magnetic susceptibility (in blue) to that of Core KN166-14-JPC-13 (red), and (c) the resulting comparison of Site U1304 (blue) and Core

KN166-14-JPC-13 (red) RPI proxies. The two RPI records were normalized by their respective mean values for comparison.

Figure 3. (a) Cores from Site U1304 that were sampled using u-channels for this study. (b) Site U1304 benthic $\delta^{18}\text{O}$ (blue) with marine isotope stages labeled by dark red numbering. (c) ChRM declination before (light blue) and after correction (green) by uniform rotation of individual cores so that the mean of corrected declinations is oriented north or south for intervals with positive and negative inclinations, respectively. (d) ChRM inclinations (red) with polarity chrons/subchrons and excursions labeled in black. (e) Maximum angular deviation (MAD; black) associated with the ChRM calculation. In (d): E-I, II, and III = Magnetic excursion I, II, and III, Cobb Mtn = Cobb Mountain Subchron. Vertical dark and light gray bars indicate intervals with predominant and disseminated diatom mats (i.e. DRV = 1 or 0.5, see text), and vertical yellow bar denotes a ~3.5-m long interval dominated by a thick diatom mat and that was not sampled.

Figure 4. Orthogonal projections of NRM demagnetization data for representative samples from non-diatom (clay-rich) intervals and diatom-rich intervals (DRV = 0.5 or 1, see text), and for samples from the three recorded magnetic excursion intervals. Circles (red) and squares (blue) indicate projections onto vertical and horizontal planes, respectively (intensity units are mA/m). The peak AFs are 0 (no demagnetization), 20-60 mT in 5 mT increments, 60-80 mT in 10 mT intervals, and 100 mT. The 20 mT and/or 80 mT demagnetization steps are labeled.

Figure 5. Site U1304: (a) Digital image of the composite splice. (b) Percentage of core section composed of diatom ooze (black, Expedition 303 Scientists, 2006) and relative abundance of *Thalassiothrix longissima* (purple, Shimada et al., 2008). (c) Volume magnetic susceptibility (red). (d) ARM intensity after 35 mT AF demagnetization (blue). (e) IRM intensity after 35 mT AF demagnetization (green). (f) S-ratios (light blue). Vertical dark and light gray bars indicate intervals with predominant and disseminated diatom mats (i.e. DRV = 1 or 0.5, see text), and the vertical yellow bar denotes a ~3.5-m interval dominated by a thick diatom mat that was not sampled.

Figure 6. (a) Hysteresis ratios for Site U1304 samples from each u-channel sample on a Day et al. (1977) plot; and representative hysteresis loops (blue curves) and back field measurements (red curve) from non-diatom (clay-rich) intervals (b and c) and diatom-rich (d and e) sediments. Plotted hysteresis ratios are coloured according to average susceptibility of the u-channel, from which the bulk sediment samples were collected. SD = single domain; PSD = pseudo-single domain; MD = multi-domain.

Figure 7. Comparison of non-diatom (blue) and diatom-rich (red) samples in histograms of: (a) NRM intensity after 35-mT peak field demagnetization, (b) corrected ChRM declinations, (c) ChRM inclinations, (e) ARM intensity after 35-mT peak field demagnetization, and (f) slopes of NRM versus ARM acquisition (RPI proxy). (d) Comparison of linear correlation coefficients (r-values) associated with best-fit lines for NRM lost versus ARM lost (blue), NRM lost versus IRM lost (red), and NRM lost versus

ARM acquired (green). Note that NRM and ARM acquisition intensity data in (a) and (e) are on logarithmic scales.

Figure 8. Relative paleointensity (RPI) slope values (bottom panel) and accompanying linear correlation coefficients (r-values, upper panel) for best-fit lines of NRM lost versus ARM lost data (blue) and for NRM lost versus ARM acquired data (red). Slopes were determined using NRM and ARM data from the 20-60 mT demagnetization/acquisition interval, at 1-cm intervals. Vertical dark and light gray bars indicate intervals with predominant and disseminated diatom mats (i.e. DRV = 1 or 0.5, see text), and the vertical yellow bar denotes a ~3.5-m interval dominated by a thick diatom mat that was not sampled.

Figure 9. Site U1304: (a) ARM intensity after 35 mT demagnetization (green), (b) benthic $\delta^{18}\text{O}$ (red) compared with the LR04 stack (blue, Lisiecki and Raymo, 2005), (c) sedimentation rate (black), (d) interpretation of magnetic polarity stratigraphy, (e) ChRM inclination, and (f) RPI (red) compared with the PISO-1500 RPI stack (blue). Light blue labels in (b) mark marine isotope stages. In (d): E-I, II, and III = Magnetic excursions I, II, and III, Ja. = Jaramillo Subchron, and Cobb Mtn = Cobb Mountain Subchron. Site U1304 and PISO-1500 RPI records were normalized by their mean values for comparison in (f). Vertical dark and light gray bars indicate intervals with predominant and disseminated diatom mats (i.e. DRV = 1 or 0.5, see text), and the vertical yellow bar denotes a ~3.5-m interval dominated by a thick diatom mat that was not sampled.

Figure 10. (a) IODP Site U1304 RPI record (red) compared with RPI records from ODP Sites 983 (blue) and 984 (light blue) (Channell et al., 1997, 2002; Channell, 1999, Channell and Kleiven, 2000), and from IODP Site U1306 (gray) (Channell et al., 2014). (b) The HINAPIS-1500 RPI stack (red) compared with the PISO-1500 RPI stack (black) (Channell et al., 2009) with 90% confidence level estimated through 10,000 bootstrapped populations (gray). (c) Number of RPI values that each RPI record contributes to time intervals in the HINAPIS-1500 stack. Note that all RPI shown in (a) and (b) are normalized by their mean values for the 0-1.5 Ma time interval for comparison.

Table 1. Site U1304 age model.

Depth (mcd)	Age (ka)	Method	Depth (mcd)	Age (ka)	Method	Depth (mcd)	Age (ka)	Method	Depth (mcd)	Age (ka)	Method
0.00	0.00	(a)	136.47	739.97	(c)	183.19	1126.25	(d)	206.01	1325.09	(d)
0.96	2.64	(b)	136.85	746.55	(c)	183.41	1129.24	(d)	206.24	1328.08	(d)
3.66	14.79	(b)	140.88	794.63	(c)	183.63	1132.23	(d)	206.46	1329.57	(d)
4.31	17.42	(b)	142.92	822.43	(c)	183.86	1136.72	(d)	206.68	1331.07	(d)
4.75	24.08	(b)	143.40	826.27	(c)	184.08	1141.20	(d)	206.91	1332.56	(d)
5.89	41.01	(b)	145.24	839.44	(c)	184.31	1144.19	(d)	207.13	1334.06	(d)
7.79	55.33	(b)	148.17	864.14	(c)	184.53	1147.18	(d)	207.35	1335.55	(d)
8.10	57.57	(b)	148.87	871.88	(c)	184.75	1148.68	(d)	207.58	1337.05	(d)
8.36	59.42	(b)	149.30	888.55	(c)	185.65	1153.16	(d)	207.80	1340.04	(d)
9.39	68.62	(b)	150.00	900.35	(c)	185.87	1154.66	(d)	208.92	1346.02	(d)
10.75	79.00	(b)	153.79	916.04	(c)	186.32	1159.14	(d)	209.59	1349.01	(d)
11.83	86.32	(b)	158.15	956.00	(c)	186.54	1162.13	(d)	210.04	1350.50	(d)
12.91	93.48	(b)	158.50	964.19	(c)	186.99	1169.61	(d)	210.71	1352.00	(d)
13.60	98.94	(b)	162.39	978.80	(c)	187.21	1174.09	(d)	211.38	1353.49	(d)
14.39	104.05	(b)	164.00	990.00	(c)	187.44	1178.58	(d)	211.83	1354.99	(d)
14.87	107.53	(b)	165.91	997.99	(c)	187.66	1183.06	(d)	212.05	1356.48	(d)
15.02	108.25	(b)	166.18	1012.63	(d)	187.89	1187.55	(d)	212.50	1360.97	(d)
15.77	110.34	(b)	166.40	1017.12	(d)	188.11	1192.03	(d)	212.95	1368.44	(d)
19.98	119.86	(b)	166.63	1021.60	(d)	188.33	1196.52	(d)	213.17	1371.43	(d)
21.94	123.70	(b)	166.85	1026.09	(d)	188.56	1201.00	(d)	213.40	1372.93	(d)
22.39	126.07	(c)	167.08	1029.08	(d)	188.78	1205.49	(d)	213.84	1374.42	(d)
22.66	134.81	(c)	167.30	1030.57	(d)	189.00	1209.97	(d)	214.51	1375.92	(d)
25.35	188.37	(c)	167.97	1033.56	(d)	189.23	1211.47	(d)	215.63	1378.91	(d)
27.83	204.19	(c)	169.09	1038.05	(d)	189.68	1212.96	(d)	216.08	1380.40	(d)
31.02	216.90	(c)	170.21	1042.53	(d)	190.80	1215.95	(d)	216.30	1381.90	(d)
31.42	223.31	(c)	171.33	1047.02	(d)	191.24	1217.45	(d)	216.53	1384.89	(d)
32.14	229.88	(c)	171.77	1048.51	(d)	191.91	1220.44	(d)	216.75	1389.37	(d)
34.30	240.03	(c)	172.89	1051.50	(d)	192.14	1221.93	(d)	217.20	1396.85	(d)
34.90	245.76	(c)	173.56	1053.00	(d)	192.36	1223.43	(d)	217.87	1402.83	(d)
38.25	277.28	(c)	174.24	1054.49	(d)	192.59	1224.92	(d)	218.09	1405.82	(d)
40.70	293.24	(c)	174.91	1055.99	(d)	193.48	1232.40	(d)	218.32	1410.30	(d)
41.45	299.26	(c)	175.13	1057.48	(d)	194.15	1238.38	(d)	218.54	1414.79	(d)
42.77	305.15	(c)	175.35	1060.47	(d)	194.60	1242.86	(d)	218.77	1419.27	(d)
51.58	332.01	(c)	175.58	1063.46	(d)	195.05	1250.34	(d)	218.99	1423.76	(d)
52.24	341.35	(c)	175.80	1066.45	(d)	195.27	1254.82	(d)	219.21	1428.24	(d)
54.09	374.47	(c)	176.03	1070.94	(d)	195.49	1259.31	(d)	219.44	1432.73	(d)
59.84	391.82	(c)	176.25	1073.93	(d)	195.72	1262.30	(d)	219.66	1437.21	(d)
71.78	423.34	(c)	176.47	1075.42	(d)	196.17	1266.78	(d)	219.89	1441.70	(d)
72.38	432.03	(c)	177.37	1079.91	(d)	197.06	1274.26	(d)	220.11	1446.18	(d)
74.85	474.92	(c)	177.59	1081.40	(d)	197.96	1281.73	(d)	220.33	1450.67	(d)
75.54	483.92	(c)	177.82	1085.89	(d)	198.18	1283.23	(d)	220.56	1455.15	(d)
100.87	530.75	(c)	178.04	1090.37	(d)	198.40	1284.72	(d)	220.78	1459.64	(d)
103.99	552.39	(c)	178.49	1094.86	(d)	199.30	1289.21	(d)	221.00	1464.12	(d)
110.97	577.36	(c)	178.71	1096.35	(d)	199.97	1292.20	(d)	221.23	1468.61	(d)
112.57	584.97	(c)	178.94	1097.85	(d)	201.09	1296.68	(d)	221.45	1473.09	(d)
119.78	613.56	(c)	179.38	1102.33	(d)	202.21	1299.67	(d)	221.68	1477.58	(d)
121.47	629.71	(c)	179.83	1106.82	(d)	203.33	1302.66	(d)	221.90	1482.06	(d)
123.64	662.50	(c)	180.05	1108.31	(d)	204.44	1305.65	(d)	222.12	1486.55	(d)
125.12	685.03	(c)	180.73	1111.30	(d)	204.89	1307.15	(d)	222.35	1491.03	(d)
125.61	690.30	(c)	181.84	1115.79	(d)	205.12	1308.64	(d)	222.57	1495.52	(d)
130.51	709.99	(c)	182.52	1118.78	(d)	205.34	1311.63	(d)	222.79	1500.00	(d)
131.20	717.38	(c)	182.74	1120.27	(d)	205.56	1316.12	(d)	262.00	1778.00	(e)
132.91	728.04	(c)	182.96	1123.26	(d)	205.79	1320.60	(d)			

Method (a) = core top has present age; (b) = correlation of U1304 susceptibility to that of core KN166-14-JPC-13 (Hodell et al., 2010); (c) = correlation of U1304 benthic $\delta^{18}\text{O}$ to LR04 stack (Lisiecki and Raymo, 2005); (d) = automated matching (Lisiecki and Lisiecki, 2002) of Site U1304 RPI to the PISO-1500 RPI stack (Channell et al., 2009); and (e) top of the Olduvai subchron according to shipboard paleomagnetic data (Expedition 303 Scientists, 2006).

Table 2. Geomagnetic reversals and excursions recorded at Site U1304.

Reversal/Excursions	Top		Bottom		Center		MIS
	Depth (mcd)	Age (ka)	Depth (mcd)	Age (ka)	Depth (mcd)	Age (ka)	
(b) C1n (M/B Boundary)	138.59	767.31	139.36	776.50	138.98	771.96	19
(t) C1r.1n (Jaramillo)	164.43	991.80	164.82	993.43	164.63	992.64	28
(b) C1r.1n (Jaramillo)	175.96	1069.62	176.11	1072.07	176.04	1071.13	31
(t) C1r.2n (Cobb Mountain)	188.14	1192.65	188.2	1193.85	188.17	1193.25	35
(b) C1r.2n (Cobb Mountain)	191.38	1218.06	191.7	1219.49	191.54	1218.78	37
Magnetic Excursion I	149.23	885.84	149.34	889.22	149.29	887.97	22/23
Magnetic Excursion II	182.96	1123.22	183.02	1124.02	182.99	1123.62	34
Magnetic Excursion III	220.75	1459.04	221.18	1467.67	220.97	1463.46	48/49

(b) = base; (t) = top.; M/B = Matuyama/Brunhes; MIS = Marine Isotope Stage. Estimated ages are based on linear interpolation of age model in Table 1.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10