1 Relative paleointensity (RPI) and oxygen isotope stratigraphy at IODP Site U1308:
2 North Atlantic RPI stack for 1.2-2.2 Ma (NARPI-2200) and age of the Olduvai
3 Subchron

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8 Abstract

9 Integrated Ocean Drilling Program (IODP) Site U1308 (49° 53'N, 24° 14'W;
10 water depth 3871 m) provides a record of relative paleointensity (RPI) and benthic
11 stable isotope stratigraphy back to 3.2 Ma. The record since 1.5 Ma was published
12 previously, and here we present the interval from 1.5 Ma to 3.2 Ma (Early
13 Pleistocene-Late Pliocene). The benthic oxygen isotope record in this interval is
14 correlated to Marine Isotope Stage (MIS) 51 to KM2, with an apparent hiatus that
15 removed part of the interval spanning MIS 104-G2 (2.6 to 2.65 Ma), at the Gauss-
16 Matuyama boundary. The mean sedimentation rate for the 1.5-3.2 Ma period is 8.5
17 cm/kyr. The age model was built by correlation of the benthic oxygen isotope
18 record to a global stack (LR04). Apart from the expected polarity reversals, three
19 magnetic excursions are recorded: Punaruu in MIS 31/32 at 1092 ka, Gilsa in MIS
20 54/55 at 1584 ka, and a newly recognized excursion labeled Porcupine (after the

nearby Porcupine Abyssal Plain) in MIS G6/G7 at 2737 ka. The ages of polarity reversals at Site U1308, on the LR04 time scale, are consistent with the current geomagnetic polarity timescale (GPTS) with the exception of the base of the Olduvai Subchron that occurred in MIS 73, corresponding to 1925 ka on the LR04 time scale, 25 kyr younger than in the current GPTS. The RPI record at Site U1308 is calibrated using the oxygen isotope age model, and combined with four other North Atlantic records to obtain a North Atlantic RPI stack for 1.2-2.2 Ma (NARPI-2200) that is compared with published RPI stacks: Epapis, Sint-2000 and PADM2M. For 2.2-3.2 Ma, the Site U1308 RPI record is compared with a RPI record from North Atlantic IODP Site U1314, and with the Pacific Epapis stack. The mean sedimentation rates of the North Atlantic sites in NARPI-2200 are greater (by about an order of magnitude) than most of the records incorporated in other stacks. The comparison of Epapis and NARPI-2200 yields an apparent lag for NARPI-2200 relative to Epapis, attributed the Epapis age model constructed by correlation of magnetic concentration parameters (a proxy for carbonate percentage) to a calibrated oxygen isotope record. The long RPI record from Site U1308 yields a very similar mean value for the Brunhes and Matuyama virtual axial dipole moments (7.05×10^{22} Am²), implying no polarity bias in the strength of the main geomagnetic dipole, in contrast to interpretations from Sint-2000 and PADM2M. The results strengthen the case that RPI can be used to improve global stratigraphic correlation, for sites with mean sedimentation rates up to several decimeters/kyr.

Keywords: Quaternary, Late Pliocene, North Atlantic, stable isotopes, relative paleointensity, age of Olduvai Subchron.

1. Introduction

Integrated Ocean Drilling Program (IODP) Site U1308 (49°53' N, 24° 14' W) was drilled in 2004 in a water depth of 3871 m (Fig. 1). It is a reoccupation of Deep Sea Drilling Project (DSDP) Site 609, drilled in 1983. Construction of a complete composite section at DSDP Site 609 was compromised by intervals of poor recovery (Ruddiman et al., 1987). Six holes were drilled at Site U1308 and a composite splice was constructed shipboard to 248 meters composite depth (mcd) (Expedition 303 Scientists, 2006). The composite splice was then modified from its shipboard version in the upper 103 mcd (Hodell et al., 2008) and, for this paper, we use a mcd-scale amended to follow the modified version. Shipboard paleomagnetic measurements (Expedition 303 Scientists, 2006) demonstrated that the recovered section at Site U1308 reaches to just below the base of the Mammoth Subchronozone at ~264 meters composite depth (mcd), although the composite splice ends at 248 mcd, just below the base of the Kaena Subchronozone. Post-cruise magnetic measurements using u-channel samples, continuous 2x2x150 cm³ samples encased in plastic, collected from each core section within the composite splice, extend to just below the Kaena Subchronozone to 248 mcd, equivalent to 3.15 Ma. Below this depth, the sediments were too indurated for efficient u-channel sampling.

Benthic stable isotope data, and relative paleointensity (RPI) data, from the upper 109 mcd (last 1.5 Ma) at Site U1308 have been published (Hodell et al., 2008; Channell et al., 2008). The Site U1308 RPI/isotope records were used as the target

for the generation of the PISO-1500 paleointensity stack for the last 1500 kyrs that utilized 13 records, mainly but not entirely from the North Atlantic, which have both oxygen isotope data and RPI associated with them (Channell et al., 2009). For PISO-1500, the RPI and oxygen isotope records from the 12 sites were optimally correlated *in tandem* to the RPI and oxygen isotope records of Site U1308, using the *Match* protocol (Lisiecki and Lisiecki, 2002). The PISO stack therefore utilizes the Site U1308 age model. Here, we present Site U1308 RPI/ $\delta^{18}\text{O}$ records for 1.5-3.2 Ma and generate a RPI stack for 1.2-2.2 Ma (NARPI-2200) from five available North Atlantic records, which is then compared with existing RPI stacks.

DSDP Site 609 has been important for studies of benthic $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and CaCO_3 for the Pleistocene and late Pliocene and the interpretation of these records in terms of ice sheet variability and oceanic circulation, and for producing orbitally-tuned timescales integrated with magnetic polarity stratigraphy (Ruddiman et al., 1986, 1989; Raymo et al., 1989). Also at this site, early recognition of detrital (Heinrich-type) layers led to the correlation of sea surface temperature minima to cold stadials in the $\delta^{18}\text{O}$ from the Summit (Greenland) ice core record and to the grouping of Dansgaard-Oeschger (ice core) cycles into Bond cycles where each cycle is terminated by the massive iceberg discharges from Hudson Strait that constitute Heinrich events (Bond et al., 1993, 1999). The purpose of this paper is to integrate the magnetic stratigraphy and oxygen isotope stratigraphy at Site U1308 prior to 1.5 Ma, and produce a high-resolution temporally calibrated RPI record for 1.5-3.2 Ma.

2. Methods

2.1 Magnetic methods

Core sections within the composite splice at Site U1308 were sampled using u-channels to the base of the composite section at 248 mcd. For this paper, the natural remanent magnetization (NRM), of u-channel samples collected from the 110-248 mcd interval, was measured at 1-cm intervals on a u-channel cryogenic magnetometer at the University of Florida, with a 10-cm leader/trailer, as the sample was tracked through the measurement space. Each 1-cm measurement is not independent of adjacent measurements due to the ~4.5-cm width at half-height of the magnetometer response function (see Weeks et al., 1993; Guyodo et al., 2002). After initial NRM measurement, stepwise AF demagnetization was carried out at 5 mT intervals from 10-80 mT, and then 10 mT steps to 100 mT. Component magnetizations were determined for a uniform 20-80 mT demagnetization interval, at each 1-cm measurement step, using the standard procedure (Kirschvink, 1980) and the software of Xuan and Channell (2009). The maximum angular deviation (MAD) values provide a measure of the definition of each determination of the component magnetization direction, with values $<5^\circ$ indicating high quality data. After NRM measurements, the volume magnetic susceptibility (κ) of u-channel samples was measured at 1-cm spacing using a susceptibility track designed for u-channels (Thomas et al., 2003).

The principal behind relative paleointensity (RPI) proxies is that variations in magnetizing field strength at the time of NRM acquisition, can be determined by normalizing the NRM intensity by the intensity of a laboratory-induced

magnetization that activates the same population of grains that carry NRM, thereby compensating for changes in concentration of NRM-carrying grains down-core (see Banerjee and Mellema, 1974; Levi and Banerjee, 1976; Tauxe, 1993). The normalizer may be anhysteretic remanent magnetization (ARM) or isothermal remanent magnetization (IRM), and the method is only applicable if the magnetization is carried by fine-grained (pseudo-single domain or single domain) magnetite or titanomagnetite, with grain sizes $< \sim 10\text{-}20\text{ }\mu\text{m}$.

For this study, ARM was acquired using a $50\text{ }\mu\text{T}$ DC bias field, and IRM was acquired in a 0.3 T and then a 1 T field. The ARM normalizer activates a finer population of magnetite grains than IRM, so the choice of effective normalizer depends on the grain-size distribution of the magnetite. In practice, we use both normalizers (ARM and IRM) and carry out both ARM acquisition (ARMAQ) and ARM/IRM demagnetization (e.g. Channell et al., 2014). For three methods of normalization (NRM/ARM, NRM/ARMAQ and NRM/IRM), we determine slopes in a chosen demagnetization/acquisition peak AF field interval (in this case 20-60 mT), and determine the linear correlation coefficient (r) at 1-cm intervals associated with each of three slopes (see Channell et al., 2002, 2014). Agreement amongst the three normalized remanence ratios may improve confidence in the RPI proxies. Note that the RPI proxies yield *relative* paleointensity estimates, and must be calibrated using absolute paleointensity determinations, to yield virtual axial dipole moments (VADM).

The anhysteretic susceptibility (κ_{ARM}) is the ARM intensity divided by the DC bias field used to generate it (in our case: 50 μT). The ratio $\kappa_{\text{ARM}}/\kappa$ can be used as a proxy for magnetite grain size (King et al., 1983), and uniformity in the ratio is an indicator of uniformity in magnetite grain size that is an important criterion for generating useful RPI data. Using a Princeton Measurements Corp. vibrating sample magnetometer (VSM), we measured small (toothpick) samples in #4 gel-caps (N=71), from each u-channel sample, to determine the hysteresis parameters. These measurements yield a Day et al. (1977) diagram of hysteresis ratios, also indicative of magnetite grain size.

2.2. Stable isotope methods

After completion of magnetic measurements, the u-channel samples were subsampled for stable isotope analyses, with authorization from IODP. Sediment was bagged in 5-cm aliquots from the square 2x2 cm cross-section of the u-channels, resulting in 30 samples from each 150-cm u-channel. These 20 cm^3 samples, from each 5-cm stratigraphic interval of the u-channels, were washed and sieved at 63 μm . Foraminifer specimens were picked from the >212- μm size fraction, and one to five individuals were used for analysis. Stable isotopes were measured on the benthic foraminifer *Cibicidoides wuellerstorfi* and/or *Cibicidoides kullenbergi* at a sample spacing of 5-10 cm for the 100-248 mcd interval of Site U1308. Foraminifer tests were soaked in ~15% H_2O_2 for 30 min to remove organic matter. The tests were then rinsed with methanol and sonically cleaned to remove fine-grained particles. The methanol was removed with a syringe, and samples were dried in an

oven at 50°C for 24 hr. The foraminifer calcite was loaded into individual reaction vessels, and each sample was reacted with three drops of H₃PO₄ (specific gravity = 1.92) using a Finnigan MAT Kiel III carbonate preparation device. Isotope ratios were measured online by a Finnigan MAT 252 mass spectrometer at the University of Florida. Analytical precision was estimated to be +/-0.08 ‰ for $\delta^{18}\text{O}$ by measuring eight standards (NBS-19) with each set of 38 samples. A total of 1708 samples of *Cibicidoides wuellerstorfi* and/or *Cibicidoides kullenbergi* were measured for this study.

3. Results

3.1. Stable isotopes

The oxygen isotope data for Site U1308 in the 100-250 mcd interval (Fig. 2) display glacial-interglacial cycles that are interpreted to represent MIS 44 to KM2. The Site U1308 $\delta^{18}\text{O}$ data from MIS 50 to recent have been published previously (Hodell et al., 2008). Bailey et al. (2010) measured $\delta^{18}\text{O}$ for Site U1308 samples in the vicinity of the Gauss-Matuyama (G/M) boundary (MIS 100-G8, red symbols in the $\delta^{18}\text{O}$ record of Figure 2), and their identification of isotopic stages is entirely consistent with the interpretation given here. Compared to our $\delta^{18}\text{O}$ values, the higher values of $\delta^{18}\text{O}$ from Bailey et al. (2010) are due to the addition of 0.64 ‰ that was applied to adjust their data to equilibrium calcite to compensate for the species used (*Cibicidoides wuellerstorfi*) that secretes calcite out of equilibrium with seawater (Shackleton and Hall, 1984). In Bailey et al. (2010), the G/M boundary was

placed in a 4 m interval at 195.5-199.5 mcd (adjusted to mcd-scale of Hodell et al., 2008) based on shipboard magnetic data (Expedition 303 Scientists, 2006), and MIS 100-102 and MIS G4-G7 are recorded (see Figure 4 of Bailey et al., 2010). We can now revise the placement of the G/M boundary to ~196-197 mcd (Table 2); however, the G/M boundary interval is complicated by low sedimentation rates and syn-sedimentary deformation. It appears, from our data, that part of the MIS 104-G2 interval was removed resulting in a hiatus between ~2.6 to 2.65 Ma (Fig. 2). Support for a hiatus in this interval comes from digital image scans of core sections from 195-198 mcd of the composite splice (sections U1308F-19H-3 to 5) that provide evidence of slumping in the vicinity of 197 mcd (Fig. 3).

U-channel samples collected from the composite splice provide overlap at ties between holes within the composite splice, because, if part of the core section was within the composite splice, we were permitted to sample the entire core section. This allowed us to test three ties within the composite splice that were deemed less secure during shipboard composite section construction (Expedition 303 Scientists, 2006). The ties at 193.65 mcd (adjusted mcd to Hodell et al, 2008) between Holes U1308A and U1308F, at 210.61 mcd between Holes U1308F and U1308C, and at 229.63 mcd between Holes U1308F and U1308C, were tested, and the overlapping isotope data support these hole-to-hole correlations.

3.2. Rock magnetism

The rock magnetism of the upper part of the section at Site U1308 (< 1.5 Ma) indicates that magnetite is the dominant magnetic mineral (Channell et al., 2008). In

Figure 2, we also plot three magnetic concentration parameters, intensity of ARM and IRM (after demagnetization at peak fields of 30 mT) and volume susceptibility, with the ratio of anhysteretic susceptibility to susceptibility ($\kappa_{\text{ARM}}/\kappa$), a grain size proxy for magnetite. In general, glacial isotopic stages are associated with higher magnetic concentration (high values of ARM, IRM and susceptibility) and a tendency for coarser grain sizes (lower values of $\kappa_{\text{ARM}}/\kappa$) (Fig. 2). A long-term decrease in $\kappa_{\text{ARM}}/\kappa$ from 250 mcd to ~ 190 mcd can be associated with increased detrital (magnetite) grain size during the intensification of Northern Hemisphere Glaciation (NHG), although the overall range of the $\kappa_{\text{ARM}}/\kappa$ ratio indicates a restricted range of magnetite grain size. The distribution of points on a κ_{ARM} versus κ plot indicates highly uniform magnetite grain size (Fig. 4a), and can be associated with a particular magnetite grain size (King et al., 1983). The grain size estimates are unrealistically low ($<0.1 \mu\text{m}$) unless the ARM intensity after AF demagnetization is plotted (Fig. 4a). A peak demagnetization field of 30 mT for ARM yields an estimate of $\sim 1 \mu\text{m}$ for the mean magnetite grain size, consistent with the estimate from the Day et al. (1977) hysteresis ratio plot (Fig. 4b) in which samples lie in the pseudo-single domain (PSD) field in a region associated with mean magnetite grain sizes of $\sim 1 \mu\text{m}$. The restricted scatter of hysteresis ratios on the magnetite grain-size mixing line in Figure 4b (mixing line adopted from Dunlop and Carter-Stiglitz, 2006), and the calibration of this mixing line using sized unannealed magnetite grains (Dunlop, 2002), provide estimates of “mean” magnetite grain size.

3.3. Natural Remanent Magnetization (NRM)

NRM magnetization components were determined for a uniform (20-80 mT) demagnetization range, and anchoring at the origin of orthogonal projections is unnecessary due to the progression of data points to the origin of the projection. The use of a uniform demagnetization interval for determination of component magnetization directions allows a straight-forward assessment of data quality through maximum angular deviation (MAD) values that are largely below 5° (Fig. 5), indicating well defined magnetization components. Declinations were corrected by uniform rotation of individual cores so that mean declinations of cores point North or South for positive and negative inclination intervals, respectively. Although the “tensor” core orientation tool was used for many individual cores, it was found to be sporadically unreliable and therefore was not employed here. The polarity pattern (Fig. 5) is compatible with the interpretation of the younger part of the Site U1308 record (<1.5 Ma) that records the Matuyama/Brunhes (MB) boundary, the Jaramillo and Cobb Mt. subchrons, and the Punaruu excursion (Channell et al., 2008). In the 100-270 mcd interval, the Olduvai and Réunion subchronozones, the Gauss/Matuyama (G/M) boundary and the Kaena Subchronozone are recognized. Shipboard paleomagnetic data in the 250-270 mcd interval (Fig. 5), after demagnetization at a single peak field of 10 mT, led to tentative identification of the Mammoth Subchronozone (Expedition 303 Scientists, 2006). Erratic component magnetization directions, particularly apparent in inclination values, at the G/M boundary are attributed to the interval of slumping (Fig. 3) that may also account of the missing isotopic stages in the MIS 104-G2 interval (Fig. 2). Two magnetic excursions are apparently recorded: the Gilsa Excursion at 117 mcd that has been

recognized elsewhere (Channell et al., 2002), and a newly recognized excursion just below the G/M boundary at 204 mcd, named Porcupine excursion after the nearby Porcupine Abyssal Plain. The Porcupine excursion is recorded by u-channel samples from the archive halves of two core sections from different holes (U1308C-20H-5 and U1308F-20H-2), and by discrete samples from the working half of U1308F-20H-2 (Fig. 5).

Orthogonal projections of AF demagnetization data from u-channel measurements of core section U1308F-11H-5 define the Gilsa excursion (Fig. 6a). U-channel samples from archive halves of core sections from different holes (1308C-20H-5 and U1308F-20H-2), and discrete 7-cm³ samples from the working half of core section U1308F-20H-2, define the Porcupine excursion (Fig. 6b-d). The orthogonal projections indicate the fidelity of the NRM directions even in the vicinity of the Gilsa and Porcupine excursions where magnetization directions are influenced by rapidly changing magnetizing fields.

3.4. Relative Paleointensity (RPI)

NRM/ARM, NRM/ARMAQ and NRM/IRM provide relative paleointensity (RPI) proxies determined as slopes calculated for a uniform 20-60 mT demagnetization/acquisition interval, at 1-cm intervals along the composite splice. Linear correlation coefficients (r) were calculated for each slope as a measure of their definition (Fig. 7). The linear correlation coefficients are lower (less than 1) for NRM/IRM slopes than for NRM/ARM or NRM/ARMAQ slopes, indicating less well defined NRM/IRM slopes, particularly in the 125-150 mcd interval. This is also

the interval in which magnetic concentration parameters indicate enhanced magnetic concentration, from high values of ARM and IRM intensity and susceptibility, and coarser magnetic grain size from lower values of $\kappa_{\text{ARM}}/\kappa$ (Fig. 2). In general, the three RPI proxies (NRM/ARM, NRM/ARMAQ and NRM/IRM) are consistent (Fig. 7). Bivariate plots of the NRM/ARM slopes versus ARM intensity and $\kappa_{\text{ARM}}/\kappa$ indicate no statistically significant relationship between the RPI proxy (NRM/ARM) and the normalizer (ARM) and the magnetite grain size proxy (Fig. 8). The NRM/ARM slope is used as the RPI proxy for Site U1308, and incorporated in the North Atlantic RPI stack (NARPI-2200).

4. Site U1308 Age Model

The Site U1308 age model was constructed by correlation of the benthic oxygen isotope data to the LR04 benthic oxygen isotope stack (Lisiecki and Raymo, 2005) using 50 tie points (Table 1) with linear interpolation between adjacent tie points. The age model is consistent with the younger (<1.5 Ma) part of the Site U1308 record (Hodell et al., 2008), and uses an analogous procedure through correlation to LR04. For MIS 51-93 (1500-2400 ka), the age model indicates sedimentation rates in the 3-18 cm/kyr range with a minimum in glacial stage MIS 70 and a maximum in interglacial MIS 93 (Fig. 9). For the MIS 95-KM2 (2400-3150 ka) interval, very low sedimentation rates occur in MIS 96, MIS 100, the MIS 104-G2 interval and MIS G12; with a maximum sedimentation rate in MIS G13 (Fig. 9). Sedimentation rates prior to 3125 ka are derived from shipboard paleomagnetic measurements (Expedition 303 Scientists, 2006) that allow tentative identification of the Mammoth

Subchronozone (Fig. 5). We observe a progressive increase in $\delta^{18}\text{O}$ from ~ 2.6 ‰ in MIS K1 to ~ 3.5 ‰ in MIS 99-100, and thereafter, as global ice volume progressively increased during the intensification of NHG. The step increase in benthic $\delta^{18}\text{O}$ at $\sim$ MIS 100, highlighted by Bailey et al. (2010) from LR04, is less apparent in our data. As mentioned above, the resulting age model includes a hiatus that appears to have removed part of the MIS 104-G2 interval (Fig. 9), at the G/M boundary, attributed to apparent slumping (Fig. 3). The ties from Site U1308 to LR04 (Table 1) were determined without reference to the polarity stratigraphy; therefore, the apparent age of polarity reversals (Table 2) provides a check on the age model.

5. The geomagnetic polarity timescale (GPTS)

Prior to 1990, the consensus GPTS for the last 2.5 Myrs was based on K-Ar radiometric ages from lavas that recorded polarity reversals of the Brunhes and Matuyama chrons (e.g., Mankinen and Dalrymple, 1979). Shackleton et al. (1990) were the first to demonstrate that these ages were 5-7% too old, from astronomical tuning of the ODP Site 677 planktic and benthic $\delta^{18}\text{O}$ records. Polarity reversals were not recorded at ODP Site 677, and the ages of reversals in Shackleton et al. (1990) were determined by correlation of the $\delta^{18}\text{O}$ record of ODP Site 677 to DSDP Site 607, which possessed both an oxygen isotope and polarity reversal stratigraphy (Raymo et al., 1989; Clement and Robinson, 1987). The ages of polarity reversals provided by Shackleton et al. (1990) have stood for 25 years, supported by corroborating ages determined from cyclostratigraphy of Mediterranean sapropels

(Hilgen, 1991a), and are adopted with only minor modifications in current versions of the GPTS (Lourens et al., 2004; Ogg, 2012) (Table 2).

Lisiecki and Raymo (2005), referred to as LR04, estimated the age of reversals back to the G/M boundary by determining the mean ages (Table 2) at eight DSDP/ODP sites (607, 659, 677, 982, 983, 1090, 1146, 1148) incorporated in the LR04 isotope stack, with standard deviations of mean values in the (undifferentiated) 5-22 kyr range. Only three of these sites (Sites 607, 982, 983) carry adequate magnetic stratigraphies. Apart from Site 677, that has no polarity stratigraphy, other sites (Sites 659, 1090, 1146 and 1148) have poorly-resolved polarity stratigraphies mainly based on limited shipboard demagnetization treatment (Tauxe et al., 1989; Shipboard Scientific Party, 1999; Wang et al., 2000).

Uncertainties in ages of reversals in LR04 are due to uncertainties in the position of the reversal relative to the marine isotope stratigraphy, rather than uncertainties in the ice-volume model (Imbrie and Imbrie, 1980) that provides the ages for LR04. Ice-volume model uncertainties are, however, important in assessing uncertainties in the Site U1308 age model that relies on LR04. Errors in ice-volume forcing in LR04, insolation at 65°N from Laskar et al. (1993), can be attributed to uncertainties in the tidal dissipation terms in the insolation calculation, and they are unlikely to be more than a few kyr at 3 Ma (see Lourens et al., 2004). There are uncertainties in the values for nonlinearity (b) and time constant (T_m) in the ice-volume calculation, that in the case of LR04 vary linearly from $b = 0.3$ and $T_m = 5$ kyr at 3.0 Ma, to $b = 0.6$ and $T_m = 15$ kyr at 1.5 Ma, with constant values since that time.

Likely ranges for b and T_m contribute to uncertainties in the ice-volume model of a few kyr at the Matuyama-Brunhes boundary (see Channell et al., 2010).

For the younger part (<1.5 Ma) of the Site U1308 record (Channell et al., 2008), ages of polarity reversals and their correlation to MIS are consistent with Shackleton et al. (1990) for the bounds of the Jaramillo and the Cobb Mountain subchrons (Fig. 10 and Table 2). Shackleton et al. (1990) provided a single age (1190 ka) for the Cobb Mt. Subchron, whereas a 21 kyr duration (1187-1208 ka) is resolved at Site U1308, and the age is ~ 10 -15 kyr older than in the current GPTS (Table 2). A magnetic excursion interpreted to be the Punaruu excursion (Fig. 10, Table 2), has an age of 1092 ka at the MIS31/32 transition at Site U1308 that is ~ 23 kyr younger than the age for this excursion determined at ODP Site 983/984 (1115 ka, Channell et al., 2002), and younger than the age for this excursion determined from $^{40}\text{Ar}/^{39}\text{Ar}$ radiometric dating of lavas (1105 ka, Singer et al., 1999) that record the excursion (Chauvin et al., 1990).

The Site U1308 age model yields ages for polarity reversals that are consistent with the current GPTS, for the end/top of the Olduvai Subchron but not for the onset/base of the Olduvai Subchron (Fig. 11, Table 2). For the age of the top of the Olduvai, Shackleton et al. (1990) correlated MIS 63 at DSDP Site 607, where the top Olduvai was recorded (in the boxed interval of Fig. 11c), to the ODP Site 677 astrochronology. At Site 607, the base of the Olduvai Subchronozone is at 81.68-82.46 mcd (within the boxed interval of Figure 11c), within MIS 73 according to Raymo et al. (1989). Shackleton et al. (1990) chose, however, to assign the base

Olduvai to MIS 71 and 1950 ka, although MIS 71 would now have an age <1900 ka according to LR04 (Lisiecki and Raymo, 2005). The Shackleton et al. (1990) MIS/age designation for the base of the Olduvai does not follow the Site 677/607 $\delta^{18}\text{O}$ correlation procedure used for determining the age of other Quaternary reversals, but was based on the association of the base Olduvai with the extinction of *Discoaster brouweri*, and the association of this extinction event with MIS 71 at DSDP Site 607 (Ruddiman et al., 1986) and at other locations (Backman and Shackleton, 1983). Lisiecki and Raymo (2005) placed the base Olduvai in MIS 74 at 1968 ka based on the averaging procedure mentioned above. At Site U1308, however, the base Olduvai clearly lies in MIS 73 at ~1925 ka on the LR04 time scale (Fig. 11, Table 2), which is 26-44 kyr younger than the ages predicted by Shackleton et al. (1990), Hilgen (1991a), Lisiecki and Raymo (2005), and those incorporated in current versions of the GPTS (Lourens et al., 2004; Ogg, 2012).

Although the base Olduvai was placed in MIS 74 at 1968 ka in LR04 (Lisiecki and Raymo, 2005), there are numerous examples of the base Olduvai being placed in MIS 71 and/or at ~1950 ka, although, as mentioned above, MIS 71 no longer corresponds to 1950 ka and has an age <1900 ka according to LR04. The onset of the Olduvai Subchron was correlated to MIS 71 in Core MD972143 (Hornig et al., 2002), ODP Sites 981 and 984 (Channell et al., 2003; Channell et al., 2002), although in all three cases, MIS 71 appears to have been misidentified, perhaps due to the “reinforcement syndrome” provided by the designation of Shackleton et al. (1990). At ODP Site 659, where the $\delta^{18}\text{O}$ record is well defined (Fig. 11d, Tiedemann et al.,

1994), the base Olduvai was again placed in MIS 71 although the subchron boundary is poorly defined at this site (Tauxe et al., 1989). In Italian land sections, the base of the Olduvai was placed at 1950 ka (Hilgen, 1991a) within MIS 71 (Lourens et al., 1992, 1996a) although the polarity subchron boundary is poorly defined in the relevant land sections (Monte Singa and Vrica) in southern Italy (Zijderveld et al., 1991), due to complex remagnetizations attributed to authigenic iron sulfides (Roberts et al., 2010). At North Pacific ODP Site 882, the age of the base Olduvai was also placed at 1.95 Ma, based on astrochronology utilizing orbital power in the GRAPE density data (Tiedemann and Haug, 1995).

A single Réunion Subchron is recognized at Site U1308 in the MIS 79-81 interval (Fig. 11, Table 2). The identification of a single normal polarity Réunion Subchron, rather than a doubled normal subchron (e.g., Mankinen and Dalrymple, 1979) is consistent with the result from ODP Site 982 (Channell and Guyodo, 2004) and ODP Site 981 (Channell et al., 2003). The duration of the Réunion Subchron at Site U1308 (21 kyr) is, however, about half that estimated at ODP Site 981 (38 kyr). The Gilsa excursion at Site U1308 occurs in MIS 54/55 at ~1584 ka (Fig. 11, Table 2), consistent with its age at ODP Sites 983 and 984 (Channell et al., 2002).

Although the G/M boundary at Site U1308 (Fig. 5) is affected by slumping (Fig. 3), the estimated age for the G/M boundary at Site U1308 is consistent with the age given by Shackleton et al. (1990) and Hilgen (1991a) and incorporated in the current GPTS (Table 2). The age and MIS designation for the Kaena Subchron at Site U1308 (Fig. 12, Table 2) differs from the assignment given by Lisiecki and Raymo

(2005) and is more consistent with Hilgen (1991a,b). The apparent excursion at 2737 ka within MIS G6/G7 (Fig. 12, Table 2), labeled Porcupine Excursion, lies well below the interval affected by slumping (Fig. 3).

6. Atlantic RPI Stack

At Site U1308, the three relative paleointensity (RPI) proxies, slopes of NRM/ARM, NRM/ARMAQ and NRM/IRM are often consistent (Fig. 7), but the categorical test of the Site U1308 RPI record must come from comparison with RPI records from other sites. For the 1.5-3.2 Ma interval, there are few RPI records for comparison. Because the non-axial dipole (NAD) components of the geomagnetic field have centennial-scale repeat times, based on analysis of the recent field (Lhuillier et al., 2011; Valet et al., 2008), RPI data recovered from sediments with mean sedimentation rates $< \sim 50 \text{ cm/kyr}$ are generally thought to represent the main axial dipole and therefore be globally correlative.

From the Pacific, there is the Epapis RPI stack (Yamazaki and Oda, 2005) for 0.8-3.0 Ma. Three of five studied cores in Epapis cover the 2.0-3.0 Ma interval and sedimentation rates are in the 0.8-1.5 cm/kyr range (Table 3). The Sint-2000 RPI stack (Valet et al., 2005) and PADM2M stack (Ziegler et al., 2011) incorporate, for the 1.5-2.0 Ma interval, the records from the west Philippine Sea (Horng et al., 2002, 2003), the Ontong-Java Plateau (Kok and Tauxe, 1999), the equatorial Pacific (Valet and Meynadier, 1993), the California Margin (Hayashida et al., 1999) and the Indian Ocean (Meynadier et al., 1994). The PADM2M stack also incorporates the records used in the Epapis stack (Yamazaki and Oda, 2005). The mean sedimentation of the

majority of the cores incorporated in these stacks are ~ 1 cm/kyr or less, in contrast to the sedimentation rates associated with the Atlantic records (Table 3). The quality of the age models for records within Sint-2000 and PADM2M is variable with the better quality age models from benthic $\delta^{18}\text{O}$ in the west Philippine Sea (Horng et al., 2002, 2003), and the ODP Leg 138 (equatorial Pacific) record (Valet and Meynadier, 1993) associated with the astrochronology of Shackleton et al. (1995). Note that the Sint-2000 stack, in the interval of interest here, involves matching of RPI records to one reference record from the Indian Ocean (Core MD90-0940, Meynadier et al., 1994).

Due to the low mean sedimentation rates of the Pacific/Indian ocean records, relative to North Atlantic records, we first generate a North Atlantic stack from cores that cover portions of the 1.5-2.2 Ma interval, and compare it with the Sint-2000, PADM2M and Epapis stacks. The North Atlantic RPI stack is constructed for the 1.2-2.2 Ma interval because this interval is bound by prominent minima in RPI records associated with the Cobb Mt. Subchron and Réunion Subchron. Apart from Site U1308, four North Atlantic RPI records represent this time interval (Fig. 1): ODP Sites 982 (Channell and Guyodo, 2004), ODP Sites 983 and 984 (Channell et al., 2002) and IODP Site U1307 (Mazaud et al., 2015). Of these records, only ODP Site 982 has a complete oxygen isotope record that extends back to the onset of the RPI record at 2.2 Ma (Venz et al., 1999; Flower et al., 2000; Kleiven et al., 2003; Raymo et al., 2004). The RPI records at ODP Sites 983 and 984 extend to 1.9 Ma and 2.15 Ma, respectively, and the mean sedimentation rates are several times that at Site 982

(Table 3). On the other hand, paucity of foraminifera beyond ~1.5 Ma at Sites 983 and 984 limit the extent of $\delta^{18}\text{O}$ at these sites (Raymo et al., 1998; McIntyre et al., 1999), and the age models beyond 1.5 Ma depend on sporadic $\delta^{18}\text{O}$ combined with correlation of the magnetic susceptibility (carbonate productivity proxy) to a calibrated oxygen isotope template (Channell et al., 2002). At IODP Site U1307, the RPI record extends to 2.2 Ma, with anomalous data in the 1.45-1.55 Ma interval, and the site lacks an oxygen isotope age model. The age model at Site U1307 was constructed by correlation of the RPI record to the PISO-1500 stack back to 1.5 Ma, and polarity stratigraphy (reversals), only, beyond 1.5 Ma (Mazaud et al., 2015).

In Figure 13, the Site U1308 RPI record (in red) is compared with three Atlantic stacks (in shades of blue) for the 1.2-2.2 Ma interval. Stack 1 comprises Sites U1308, 983 and 984 (3 records). Stack 2 comprises these three with IODP Site U1307 added (4 records), and Stack 3 comprises these four with ODP Site 982 added (5 records). Age ranges, and other information for individual records, are given in Table 3. The stacking procedure was as follows. The age models for ODP Sites 982, 983, and 984 and IODP Site U1307 are as published (Channell et al., 2002, Channell and Guyodo, 2004; Mazaud et al., 2015) with adjustment to account for the revised age for the onset of the Olduvai Subchron. This adjustment was carried out by linear interpolation for each record from the top of the Olduvai to the top of the Réunion Subchron, utilizing the revised age for the base of the Olduvai Subchron (1925 ka). For each record, values of RPI were then linearly interpolated to 1-kyr intervals, and each record was normalized to a common mean and a common standard deviation.

The stacks were then calculated as the arithmetic mean at each 1-kyr interval, with a standard error associated with each stack.

Stack 3 (NARPI-2200) from the North Atlantic, comprising IODP Site U1308, U1307 and ODP Sites 982-984, is shown (in blue) in Figure 14, with the error bars indicating 2 times the standard error (2σ) associated with the stack, compared to the PISO stack (in red) that extends to 1.5 Ma (Channell et al., 2009), the Sint-2000 and PADM2M stacks that extend to 2.0 Ma (Valet et al., 2005; Ziegler et al., 2011), and the Epapis stack that extends to 3.0 Ma (Yamazaki and Oda, 2005). Prominent minima in the RPI stacks are associated with polarity reversals, and with magnetic excursions.

For the 2100-2750 ka interval, beyond NARPI-2200, the RPI record from North Atlantic IODP Site U1314 (Fig. 1, Ohno et al., 2012), and can be compared with the record from Site U1308 and with the Epapis Pacific stack (Fig. 15). The age model for the Site U1314 RPI record (Ohno et al., 2012) was constructed using the correlation of susceptibility and natural gamma radiation records to LR04 (Hayashi et al., 2010). The sedimentation rate at Site U1314 in the Matuyama Chron is ~ 10 cm/kyr, and closer to 20 cm/kyr in the Gauss Chron (Ohno et al., 2012). In the 2100-2750 ka interval, the comparison between Site U1314 and Site U1308 indicates common RPI minima at polarity reversals and at the Porcupine excursion but also in intervals between reversals and excursions (Fig. 15).

7. Discussion

The GPTS for the last 3.2 Myr has remained essentially unchanged since the temporal calibration of polarity reversals provided 25 years ago by Shackleton et al. (1990) and Hilgen (1991a,b). As the $\delta^{18}\text{O}$ record at Site U1308 has been interpreted irrespective of the magnetic stratigraphy, the age model provides a check on the GPTS for the last 3.2 Myrs. The ages and MIS assignments of polarity reversals are largely confirmed by this work (Table 2, Figs. 10-12), however, the onset of the Olduvai is at 1925 ka, within MIS 73 (Fig. 11), rather than the traditional assignment at 1950 ka (Shackleton et al., 1990; Hilgen, 1991a) within MIS 71 (Shackleton et al., 1990; Lourens et al., 1992, 1996a) or MIS 74 (Lisiecki and Raymo, 2005). In addition, three magnetic excursions are observed at Site U1308: Punaruu at 1092 ka in MIS 31/32 (Fig. 10), Gilsa at 1584 ka in MIS 54/55 (Fig. 11) and Porcupine at 2737 ka in MIS G6/G7 (Fig. 12). The Porcupine excursion is a newly-recognized excursion, and confirmation of this excursion requires its identification elsewhere.

For the 1200-2200 ka interval, North Atlantic RPI Stacks 1-3, comprising 3, 4 and 5 RPI records, respectively, are compared with the Site U1308 RPI record in Figure 13. All five RPI records used in the stacks (IODP Sites U1308 and U1307, ODP Sites 982, 983 and 984) are placed on their individual published age models with a single adjustment by linear interpolation (from the top of the Olduvai to the top of the Réunion Subchron) to account for the revised age of the onset of the Olduvai Subchron. As dictated by the age ranges of the records themselves, the final RPI Stack 3 (NARPI-2200) comprises all 5 North Atlantic records for 1.2-1.9 Ma (although there is a data gap in the 1.45-1.55 Ma interval at Site U1307), only 4

503 records (excluding Site 983) for 1.9-2.1 Ma, and 3 records (excluding Sites 983 and
504 984) for 2.1-2.2 Ma.

505 The NARPI-2200 stack (thick dark blue line in Figs. 13 and 14, with the 2σ error
506 in grey shading) shows correspondence with alternative stacks :PISO, Sint-2000,
507 PADM2M and Epapis (Fig. 14). Prominent RPI minima are associated with the Cobb
508 Mt. Subchron (~ 1200 ka), the boundaries of the Olduvai Subchron (1780 ka and
509 1925 ka), and the Réunion Subchron (2150 ka) as well as the Gilsa excursion (1584
510 ka), and the Gardar excursion (1470 ka) recorded at ODP Sites 983 and 984
511 (Channell et al., 2002) but not at Site U1308. The correspondence between the
512 NARPI-2200 stack, Epapis, Sint-2000 and PADM2M goes beyond the subchron
513 boundaries, for example, some RPI minima within polarity chrons (e.g., ~ 1360 ka,
514 ~ 1460 ka and ~ 1570 ka) appear to be present all three stacks (Fig. 14). Beyond 2.2
515 Ma, the Site U1308 RPI record can be compared with the IODP Site 1314 RPI record
516 (Ohno et al., 2012) and with the Epapis stack. Prominent features in RPI records can
517 be matched for Sites U1314 and U1308, and to a less extent with Epapis (Fig. 15).

518 In the Sint-2000 and PADM2M stacks, that extend back to 2.0 Ma, there is
519 evidence for lower mean geomagnetic field intensities in the Matuyama Chron than
520 in the Brunhes Chron, by $\sim 23\%$ according to Ziegler et al. (2011). By scaling the Site
521 U1308 RPI record to VADM (Fig. 16), using the same scaling method used for the
522 PISO stack (Channell et al., 2009), we find that the Site U1308 mean VADM for the
523 Matuyama Chron and for the Brunhes Chron (7.05×10^{22} Am²) are indistinguishable,
524 implying that the geomagnetic field strength is not a function of polarity state. The

apparent difference in RPI/VADM between the Brunhes and Matuyama Chrons in Sint-2000 and PADM2M may be attributed to viscous remanence (VRM) acquired in the Brunhes Chron that reduced the magnetization intensity of some Matuyama data included in the stacks.

Comparison of the North Atlantic RPI stack with the Pacific Epapis stack in the 1.2-2.2 Ma interval (Fig. 14) appears to indicate that the North Atlantic stack lags the Pacific stack. This lag was pointed out in comparison of the PISO and Epapis stacks (Channell et al., 2009), and in comparison of the Site U1314 record and Epapis (Ohno et al., 2012), and we attribute it to the method used to generate the age models for cores included in the Epapis stack. To provide age models for cores included in Epapis, magnetic concentration parameters (ARM intensity and magnetic susceptibility) were correlated to the LR04 calibrated $\delta^{18}\text{O}$ template (Yamazaki and Oda, 2005). Glacial isotopic stages were correlated with magnetic concentration minima, implying biogenic carbonate dilution of magnetic concentration during glacial stages (the so-called Pacific carbonate pattern). The correlation may have induced a lag because magnetic concentration minima may be associated with late glacial intervals or Terminations, rather than glacial marine isotopic stages themselves.

8. Conclusions

IODP Site U1308 now has relatively high-resolution $\delta^{18}\text{O}$ and RPI records for the last 3.2 Ma, with 2-10 cm sample spacing for $\delta^{18}\text{O}$, and nominally 1-cm spacing for

RPI although adjacent RPI determinations are not independent due to the finite size of the magnetometer response function (Weeks et al., 1993; Guyodo et al., 2002). The mean sedimentation rate for the entire Site U1308 record is ~8.5 cm/kyr. A magnetic stratigraphy comprising polarity reversals, magnetic excursions (Punaru, Gilsa and Porcupine), and relative paleointensity (RPI) has been calibrated using the Site U1308 $\delta^{18}\text{O}$ record through correlation to LR04. The calibrated templates for RPI and $\delta^{18}\text{O}$ from Site U1308 (central Atlantic), augmented by three magnetic excursions, can be used as a reference template for improvement of Late Pliocene and Quaternary stratigraphies. Magnetic excursions constitute brief, millennial-scale, directional aberrations of the geomagnetic field that are only sporadically recorded at sedimentation rates similar to those at Site U1308, although they are believed to involve the demise and regeneration of the main axial dipole field that therefore are useful, when recorded, for global stratigraphic correlation (Channell, 2014).

The North Atlantic RPI stack for 1.2-2.2 Ma (NARPI-2200), compiled from five records, including Site U1308, all placed on their individual age models, with modification for a revised age for the onset of the Olduvai Subchron (1925 ka), is compared with published stacks (Epais, Sint-2000 and PADM2M) that cover all or part of the same interval. The resolution (mean sedimentation rate), and quality of the age models, of individual records controls the resolution of each stack, and other stacks are clearly smoothed relative to NARPI-2200 reflecting the lower sedimentation rates associated with RPI records from outside the Atlantic Ocean. An

apparent lag of NARPI-2200 relative to the Pacific Epapis by a few tens of kiloyears, in some parts of the record, is associated with the procedure of age model construction in Epapis.

The ages and MIS assignments of Matuyama and Gauss polarity reversals at Site U1308 are largely consistent with the current GPTS (Lourens et al., 2004; Ogg, 2012), apart from the base of the Olduvai Subchronozone that occurs in MIS 73 at 1925 ka, rather than 1950 ka. The ages for Quaternary polarity reversals in the current GPTS originate from correlation, through benthic $\delta^{18}\text{O}$, of the astrochronology from ODP Site 677 to the magnetic stratigraphy of DSDP Site 607 (Shackleton et al., 1990). This methodology, however, was not used for the onset of the Olduvai Subchron, that was dated by association of the polarity reversal with the extinction of *Discoaster brouweri* and with MIS 71 (Backman and Shackleton, 1983; Ruddiman et al., 1986) at 1950 ka (Shackleton et al., 1990). The onset of the Olduvai was recorded in MIS 73 at DSDP Site 607 (Raymo et al., 1989), and this placement is confirmed at IODP Site U1308 leading to an age of 1925 ka based on the Site U1308 age model derived by correlation to the LR04 benthic $\delta^{18}\text{O}$ stack (Lisiecki and Raymo, 2005). The ages of Quaternary polarity reversals from Shackleton et al. (1990) have very largely stood the test of time, and have been corroborated by numerous subsequent studies. In the case of the onset of the Olduvai Subchron, however, the conventional association of the polarity reversal with MIS 71 and 1.95 Ma (Shackleton et al., 1990) should be revised to MIS 73 and 1925 ka. The NARPI-2200 relative paleointensity (RPI) stack and the Site U1308 RPI data interpolated

for each 1-kyr in the 1200-2200 ka interval, both unsmoothed and with 10 point smoothing, are given in Supplemental Table 1.

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599 References

- 600 Backman, J., Shackleton, N.J., 1983. Quantitative biochronology of Pliocene and early
601 Pleistocene nannofossils from the Atlantic, Indian and Pacific oceans. *Marine*
602 *Micropaleontology*, 8, 141-170.
- 603 Bailey, I., Bolton, C.T., DeConto, R.M., Pollard, D., Schiebel, R., Wilson, P.A., 2010. A
604 low threshold for the North Atlantic ice rafting from “low-slung slippery” late
605 Pliocene ice sheets. *Paleoceanography*, 25, PA1212, doi:10.1029/2009PA001736.
- 606 Banerjee, S.K., Mellema, J.P., 1974. A new method for the determination of
607 paleointensity from the ARM properties of rocks. *Earth Planet Sci. Letters*, 23, 177-
608 184.
- 609 Bond, G., Broecker, W., Johnsen, S., McManus, J., Laberyie, L., Jouzel, J., Bonani, G.,
610 1993. Correlations between climate records from North Atlantic sediments and
611 Greenland ice. *Nature*, 365: 143-147.
- 612 Bond, G., Showers, W., Elliot, M., Evans, M., Lotti, R., Hajdas, I., Bonani, G., Johnson, S.,
613 1999. The North Atlantic's 1-2 kyr climate rhythm: relation to Heinrich Events,
614 Dansgaard/Oeschger cycles and the Little ice Age. In: *Mechanisms of millennial-*
615 *scale global climate change*. (Webb et al., ed.) *Geoph. Monograph Series*, 112, 35-58.
- 616 Channell, J.E.T., 2014. The Iceland basin excursion: Age duration, and excursion field
617 geometry. *Geochemistry, Geophysics, Geosystems*, 15, 4920-4935,
618 doi:10.1002/2014GC005564.

- 619 Channell, J.E.T., Guyodo, Y., 2004. The Matuyama Chronozone at ODP Site 982
620 (Rockall Bank): Evidence for decimeter-scale magnetization lock-in depths. In:
621 "Timescales of the Paleomagnetic Field". Channell, J.E.T., Kent, D.V., Lowrie, W.,
622 Meert, J. (Eds), AGU Geophysical Monograph 145, 205-219.
- 623 Channell, J.E.T., Hodell, D.A., 2013. Magnetic signatures of Heinrich-like detrital
624 layers in the Quaternary of the North Atlantic. *Earth Planet. Sci. Letters*, 369-370,
625 260-270.
- 626 Channell, J.E.T., Labs, J., Raymo, M.E., 2003. The Réunion Subchronozone at ODP Site
627 981 (Feni Drift, North Atlantic), *Earth Planet. Sci. Letters*, 215, 1-12.
- 628 Channell, J.E.T., Mazaud, A., Sullivan, P., Turner, S., Raymo, M.E., 2002. Geomagnetic
629 excursions and paleointensities in the 0.9-2.15 Ma interval of the Matuyama Chron
630 at ODP Site 983 and 984 (Iceland Basin), *J. Geophys. Res.*, 107 (B6)
631 10.1029/2001JB000491.
- 632 Channell, J.E.T., Hodell, D.A., Xuan, C., Mazaud, A., Stoner, J.S., 2008. Age calibrated
633 relative paleointensity for the last 1.5 Myr at IODP Site U1308 (North Atlantic).
634 *Earth Planet. Sci. Lett.*, 274, 59-71.
- 635 Channell, J.E.T., Xuan, C., Hodell, D.A., 2009. Stacking paleointensity and oxygen
636 isotope data for the last 1.5 Myrs (PISO-1500). *Earth Planet. Sci. Lett.*, 283, 14-23.
- 637 Channell, J.E.T., Hodell, D.A., Singer, B.S., Xuan, C., 2010. Reconciling
638 astrochronological and $^{40}\text{Ar}/^{39}\text{Ar}$ ages for the Matuyama-Brunhes boundary and late

- 639 Matuyama Chron. *Geochem. Geophys. Geosyst.*, 11, Q0AA12,
640 doi:10.1029/2010GC003203.
- 641 Channell, J.E.T., Wright, J.D., Mazaud, A., Stoner, J.S., 2014. Age through tandem
642 correlation of Quaternary relative paleointensity (RPI) and oxygen isotope data at
643 IODP Site U1306 (Eirik Drift, SW Greenland). *Quat. Sci. Reviews*, 88, 135-146.
- 644 Chauvin, A., Roperch, P., Duncan, R.A., 1990. Records of geomagnetic reversals from
645 volcanic islands of French Polynesia, 2-Paleomagnetic study of a flow sequence (1.2
646 to 0.6 Ma) from the Island of Tahiti and discussion of reversal models. *J. Geophys.*
647 *Res.*, 95, 2727-2752.
- 648 Clement, B., Robinson, F., 1987. The magnetostratigraphy of Leg 94 sediments. *Init.*
649 *Reps. Deep Sea Drilling Project*, 94, 635-650.
- 650 Day, R., Fuller, M., Schmidt, V.A., 1977. Hysteresis properties of titanomagnetites:
651 grain-size and compositional dependence. *Phys. Earth Planet. Int.* 13: 260-267.
- 652 Dunlop, D.J., 2002. Theory and application of the Day plot (M_{rs}/M_s versus H_{cr}/H_c) 1.
653 Theoretical curves and tests using titanomagnetite data. *J. Geophys. Res.*, 107, B3,
654 2056, doi:10.1029/2001JB000486.
- 655 Dunlop, D.J., Carter-Stiglitz, B., 2006. Day plots of mixtures of superparamagnetic,
656 single domain, pseudosingle domain, and multidomain magnetites. *J. Geophys. Res.*,
657 111, B12S09, doi:10.1029/2006JB004499.

- 658 Expedition 303 Scientists, 2006. Site U1308. In: Channell, J.E.T., Kanamatsu, T., Sato,
659 T., Stein, R., Alvarez Zarikian, C.A., Malone, M.J., Expedition 303/306 Scientists, Proc.
660 IODP, 303: College Station TX (Integrated Ocean Drilling Program Management
661 International, Inc.). [doi:10.2204/iodp.proc.303306.108.2006](https://doi.org/10.2204/iodp.proc.303306.108.2006)
- 662 Flower, B.P., Oppo, D.W., McManus, J.F., Venz, K.A., Hodell, D.A., Cullen, J.A., 2000.
663 North Atlantic intermediate to deep water circulation and chemical stratification
664 during the past 1 Myr. *Paleoceanography*, 15, 388-403.
- 665 Guyodo, Y., Channell, J.E.T., Thomas, R., 2002. Deconvolution of u-channel
666 paleomagnetic data near geomagnetic reversals and short events. *Geophys. Res.*
667 *Letters*, 29, 1845, doi:10.1029/2002GL014963.
- 668 Hayashi, T., Ohno, G., Acton, G., Guyodo, Y., Evans, H.F., Kanamatsu, T., Komatsu, F.,
669 Murakami, F., 2010. Millennial-scale iceberg surges after intensification of Northern
670 Hemisphere glaciation, *Geochem. Geophys. Geosyst.*, 11, Q09Z20,
671 doi:10.1029/2010GC003132.
- 672 Hayashida, A., Verosub, K.L., Heider, F., Leonhardt, R., 1999. Magnetostratigraphy
673 and relative paleointensity of late Neogene sediments at ODP Leg 167 Site 1010 off
674 Baja California, *Geophys. J. Int.*, 139, 829-840.
- 675 Hilgen F.J., 1991a. Astronomical calibration of Gauss to Matuyama sapropels in the
676 Mediterranean and implications for the Geomagnetic Polarity Time Scale. *Earth and*
677 *Planetary Science Letters* 104: 226–244.

- 678 Hilgen, F.J., 1991b. Extension of the astronomically calibrated (polarity) timescale to
679 the Miocene-Pliocene boundary. *Earth and Planet. Sci. Letters*, 107, 349-368.
- 680 Hodell, D.A., Channell, J.E.T., Curtis, J. H., Romero, O., Röhl, U., 2008. Onset of Heinrich
681 Events in the Central North Atlantic at the end of the Middle Pleistocene Transition
682 (~650 ka)? *Paleoceanography* 23, PA4218, doi:10.1029/2008PA001591.
- 683 Horng, C.S., Lee, M.Y., Pälike, H., Wei, K-Y., Liang, W.T., Iizuka, Y., Torii, M., 2002.
684 Astronomically calibrated ages for geomagnetic reversals within the Matuyama
685 Chron. *Earth Planets Space*, 54, 679-690.
- 686 Horng, C.-S., Roberts, A.P., Liang, W.-T., 2003. A 2.14-Myr astronomically tuned
687 record of relative geomagnetic paleointensity from the western Philippine Sea.
688 *Journal of Geophysical Research*, 108, 2059, doi:10.1029/2001JB001698.
- 689 Imbrie, J., Imbrie, J.Z., 1980. Modeling the climate response to orbital variations.
690 *Science*, 207, 943-953.
- 691 King, J.W., Banerjee, S.K., Marvin, J., 1983. A new rock-magnetic approach to
692 selecting sediments for geomagnetic paleointensity studies: application to
693 paleointensity for the last 4000 years. *J. Geophys. Res.* 88, 5911-5921.
- 694 King, J.W., Channell, J.E.T., 1991. Sedimentary magnetism, environmental
695 magnetism, and magnetostratigraphy. *Rev. Geophys. Suppl.* 358-370.
- 696 Kirschvink, J.L., 1980. The least squares lines and plane analysis of paleomagnetic
697 data. *Geophys. J.R. Astr. Soc.* 62, 699-718.

- 698 Kleiven, H.F., Jansen, E., Curry, W.B., Hodell, D.A., Venz, K., 2003. Atlantic ocean
699 thermohaline circulation changes on orbital to suborbital timescales during the mid-
700 Pleistocene. *Paleoceanography*, 18, 1008, doi:10.1029/2001PA000629.
- 701 Kok, Y.S., Tauxe, L., 1999. A relative geomagnetic paleointensity stack from Ontong-
702 Java Plateau sediments for the Matuyama. *J. Geophys. Res.*, 104, 25,401-25,413.
- 703 Laskar, J., Joutel, F., Boudin, F., 1993. Orbital, precessional and insolation quantities
704 for the Earth from -20 Myr to +10 Myr. *Astronomy and Astrophysics*, 270, 522-533.
- 705 Levi, S., Banerjee, S.K., 1976. On the possibility of obtaining relative paleointensities
706 from lake sediments. *Earth Planet. Sci. Letters*, 29, 219-226.
- 707 Lhuillier, F., Fournier, A., Hulot, G., Aubert, J., 2011. The geomagnetic secular-
708 variation timescale in observations and numerical dynamo models. *Geophys. Res.*
709 *Letters*, 38, L09306, doi:10.129/2011GL047356.
- 710 Lisiecki, L. E., Lisiecki, P. A., 2002. The application of dynamic programming to the
711 correlation of paleoclimate records, *Paleoceanography*, 17(D4), 1049,
712 doi:10.1029/2001PA000733.
- 713 Lisiecki, L.E., Raymo, M.E., 2005. A Pliocene-Pleistocene stack of 57 globally
714 distributed benthic delta O-18 records. *Paleoceanography*, 20(1).
- 715 Lourens, L.J., Hilgen, F.J., Gudjonsson, L., Zachariasse, W.J., 1992. Late Pliocene to
716 early Pleistocene astronomically forced sea surface productivity and temperature
717 variations in the Mediterranean. *Marine Micropaleontology*, 19, 49-78.

- 718 Lourens, L.J., Antonarakou, A., Hilgen, F.J., Van Hoof A.A.M., Vergnaud-Grazzini, C.,
719 Zachariasse, W.J., 1996a. Evaluation of the Plio-Pleistocene astronomical timescale.
720 *Paleoceanography*, 11 (4), 391-413.
- 721 Lourens, L.J., Hilgen, F.J., Raffi, I., Vergnaud-Grazzini, C., 1996b. Early Pleistocene
722 chronology of the Vrica section (Calabria, Italy). *Paleoceanography*, 11(6), 797-812.
- 723 Lourens, L., Hilgen, F.J., Shackleton, N.J., Laskar, J., Wilson, D., 2004. The Neogene
724 Period. In: Gradstein, F.M., Ogg, J.G., Smith, A.G. (Eds.), *A Geologic Time Scale 2004*.
725 Cambridge University Press, pp. 409-440.
- 726 Mankinen, E.A. Daltrymp, G.B., 1979. Revised geomagnetic polarity timescale for
727 the interval 0-5 m.y.b.p. *J. Geophys. Res.*, 84, 615-626.
- 728 Mazaud, A., Channell, J.E.T., Stoner, J.S., 2015, The paleomagnetic record at Site
729 U1307 back to 2.2 Ma (Eirik Drift, off south Greenland). *Earth Planet. Sci. Letters*,
730 429, 82-89.
- 731 McIntyre, K., Ravelo, A.C., and Delaney, M.L., 1999, North Atlantic intermediate waters
732 in the late Pliocene to early Pleistocene. *Paleoceanography*, 14, 324-335.
- 733 Meynadier, L., Valet, J.-P., Bassinot, F.C., Shackleton, N., Guyodo, Y., 1994.
734 Asymmetrical saw-tooth pattern of geomagnetic field intensity from equatorial
735 sediments in the Pacific and Indian oceans. *Earth Planet. Sci. Letters*, 126, 109-127.
- 736 Naafs, B.D.A., Hefter, J., Stein, R., 2013. Millennial-scale ice rafting events and Hudson
737 Strait Heinrich Events during the late Pliocene and Pleistocene: a review. *Quat. Sci.*

- 738 Reviews, 80, 1-28.
- 739 Ohno, M., Hayashi, T., Komatsu, F., Murakami, F., Zhao, M., Guyodo, Y., Acton, G.,
740 Evans, H.F., and Kanamatsu, T., 2012. A detailed paleomagnetic record between 2.1
741 and 2.75 Ma at IODP Site U1314 in the North Atlantic: Geomagnetic excursions and
742 the Gauss-Matuyama transition. *Geochem. Geophys. Geosyst.*, 13 (1), Q12Z39,
743 doi:10.129/2012GC004080.
- 744 Ogg, J.G., 2012. The geomagnetic polarity timescale. In: Gradstein, F., Ogg, J.G.,
745 Schmitz, M., Ogg, G. (Eds.). *The geologic time scale 2012*. Elsevier, Amsterdam, 85-
746 128.
- 747 Raymo, M.E., Ruddiman, W.F., Backman, J., Clement, B.M., Martinson, D.G., 1989. Late
748 Pliocene variation in Northern Hemisphere ice sheets and North Atlantic deep water
749 circulation. *Paleoceanography*, 4, 413-446.
- 750 Raymo, M.E., Ganley, K., Carter, S., Oppo, D.W., McManus, J., 1998. Millennial-scale
751 climate instability during the early Pleistocene epoch. *Nature* 392, 699-702.
- 752 Raymo, M.E., Oppo, D.W., Flower, B.P., Hodell, D.A., McManus, J.F., Venz, K.A., Kleiven,
753 K.F., McIntyre, K., 2004. Stability of North Atlantic water masses in face of
754 pronounced climate variability during the Pleistocene. *Paleoceanography*, 19,
755 PA2008, doi:10.1029/2003PA000921.
- 756 Roberts, A.P., Florindo, F., Larrasoana, J.C., O'Regan, M.A., Zhao, X., 2010. Complex
757 polarity pattern at the former Plio-Pleistocene global stratotype section at Vrica

- 758 (Italy): remagnetization by magnetic iron sulphides. *Earth Planet. Sci. Letters*, 292,
759 98-111.
- 760 Ruddiman, W., McIntyre, A., Raymo, M.E., 1986. Matuyama 41,000-year cycles: North
761 Atlantic ocean and northern hemisphere ice sheets. *Earth Planet. Sci. Letters*, 80,
762 117-129.
- 763 Ruddiman, W.F., Raymo, M.E., Martinson, D.G., Clement, B.M., Backman, J., 1989.
764 Pleistocene evolution: northern hemisphere ice sheet and North Atlantic Ocean.
765 *Paleoceanography*, 4, 353-412.
- 766 Ruddiman, W.F., D. Cameron, and B.M. Clement, 1987. Sediment disturbance and
767 correlation of offset holes drilled with the hydraulic piston corer: Leg 94. *Init. Repts.*
768 *DSDP*, 94, 615-624: Washington (U.S. Govt. Printing Office).
- 769 Shackleton, N.J., Hall, M.A., 1984. Oxygen and carbon isotope stratigraphy of Deep
770 Sea Drilling Project Hole 552A: Plio-Pleistocene glacial history. *Init. Repts. Deep Sea*
771 *Drilling Project*, 74, 599-604.
- 772 Shackleton N.J., Berger, A., Peltier, W.R., 1990. An alternative astronomical
773 calibration of the lower Pleistocene timescale based on ODP Site 677. *Transactions*
774 *of the Royal Society of Edinburgh: Earth Sciences*, 81, 251-261.
- 775 Shackleton, N.J., Crowhurst, S., Hagelberg, T., Pisias, N.G., Schneider, D.A., 1995. A
776 new Late Neogene time scale: application to Leg 138 Sites, In: Pisias, N.G., Janacek, L.
777 A., Palmer-Julson, A., Van Andel, T.H. (Eds.). *Proc. ODP Sci. Results*, 138: 73-101.

- 778 Shipboard Scientific Party, 1999. Site 1090. In Gersonde, R. Hodell, D.A., Blum, P., et
779 al., Proc. Ocean Drilling Program, Initial Reports, 177: College Station, TX (Ocean
780 Drilling Program), 1-101.
- 781 Singer, B.S., Hoffman, K.A., Chauvin, A., Coe, R.S., Pringle, M.S., 1999. Dating
782 transitionally magnetized lavas of the late Matuyama Chron: toward a new $^{40}\text{Ar}/^{39}\text{Ar}$
783 timescale of reversals and events, J. Geophys. Res., 104, 679-693.
- 784 Tauxe, L., 1993. Sedimentary records of relative paleointensity of the geomagnetic
785 field: theory and practice. Rev. Geophys., 31, 319-354.
- 786 Tauxe, L., Valet, J-P., Bloemendal, J., 1989. Magnetostratigraphy of Leg 108 advanced
787 hydraulic piston cores. In: Ruddiman, W., Sarnhein, M. et al., 1989. Proc. ODP, Sci.
788 Results, 108: College Station, TX (Ocean Drilling Program), 429-439.
- 789 Tiedemann, R., Haug, G.H., 1995. Astronomical calibration of cycle stratigraphy for
790 Site 882 in the Northwest Pacific. Proc. ODP Sci. Res., 145, 283-292.
- 791 Tiedemann, R., Sarnthein, M., Shackleton, N.J., 1994. Astronomic timescale for the
792 Pliocene Atlantic $\delta^{18}\text{O}$ and dust flux records of Ocean Drilling Program Site 659.
793 Paleoceanography, 9, 619-638.
- 794 Thomas, R., Guyodo, Y., Channell, J.E.T., 2003. U-channel track for susceptibility
795 measurements. Geochemistry, Geophysics and Geosystems (G^3), 1050, doi:
796 10.1029/2002GC000454.

- 797 Valet, J.P., Meynadier, L., 1993. Geomagnetic field intensity and reversals during the
798 past four million years. *Nature*, 366, 234-238.
- 799 Valet, J-P, Meynadier, L., Guyodo, Y., 2005. Geomagnetic dipole strength and reversal
800 rate over the past two million years. *Nature* 435, 802-805.
- 801 Valet, J-P., Herrero-Brevera, E., LeMouel, J-L., Plenier, G., 2008. Secular variation of
802 the geomagnetic dipole during the past 2000 years. *Geochemistry Geophysics*
803 *Geosystems*, 9, doi:10.1029/2007GC001728.
- 804 Venz, K.A., Hodell, D.A., Stanton, C., Warnke, D.A., 1999. A 1.0 Myr record of glacial
805 North Atlantic Intermediate Water variability from ODP Site 982 in the northeast
806 Atlantic, *Paleoceanography*, 14, 42-52.
- 807 Wang, P., Prell, W., Blum, P., et al., 2000. Proc. ODP Init. Reports, 184 [CD-ROM].
808 Available from: Ocean Drilling program, Texas A&M University, College Station, TX
809 77845-9547.
- 810 Weeks, R., Laj, C., Endignoux, L., Fuller, M., Roberts, A., Manganne, R., Blanchard, E.,
811 Goree, W., 1993. Improvements in long-core measurement techniques: applications
812 in palaeomagnetism and palaeoceanography. *Geophys. J. Int.*, 114, 651-662.
- 813 Xuan, C., Channell, J.E.T., 2009. UPmag: MATLAB software for viewing and
814 processing u-channel or other pass-through paleomagnetic data. *Geochem. Geophys.*
815 *Geosyst.*, 10, Q10Y07, doi:1029/2009GC002584.

816 Yamazaki, T., Oda, H., 2005. A geomagnetic paleointensity stack between 0.8 and 3.0
817 Ma from equatorial Pacific sediment cores. *Geochemistry, Geophysics and*
818 *Geosystems* (G³), 6 (11), doi: 10.1029/2005GC001001.

819 Ziegler, L.B., Constable, C.G., Johnson, C.L., Tauxe, L. 2011., PADM2M: a penalized
820 maximum likelihood model of the 0-2 Ma paleomagnetic axial dipole moment.
821 *Geophys. J. Int.*, 184, 1069-1089.

822 Zijdeveld, J.D.A., Hilgen, F.J., Langereis, C.G., Verhallen, P.J.J.M., Zachariasse, W.J.,
823 1991. Integrated magnetostratigraphy and biostratigraphy of the upper Pliocene-
824 lower Pleistocene from Monte Singa and Crotone areas in Calabria (Italy). *Earth*
825 *Planet. Sci. Letters*, 107, 697-714.

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Figure Captions

Figure 1. Location of Sites U1308, U1307 and U1314 recovered during IODP Expedition 303/306; and Sites 982, 983 and 984 recovered during ODP Leg 162.

Figure 2. Site U1308: Volume susceptibility, isothermal remanence (IRM), anhysteretic remanence (ARM), ratio of anhysteretic susceptibility to susceptibility ($\kappa_{\text{ARM}}/\kappa$, magnetic grain proxy), and benthic oxygen isotope ($\delta^{18}\text{O}$) record for the 100-250 mcd from this paper (blue) compared with the $\delta^{18}\text{O}$ record of Bailey et al. (2010) (red dots). Labels on the $\delta^{18}\text{O}$ plot indicate the interpretation in terms of marine isotopic stages (MIS) with postulated hiatus at MIS 104-G2. IRM and ARM intensities are shown after demagnetization at peak fields of 30 mT.

Figure 3. Digital image scans of three 150-cm long core sections within the composite section between 194.78 and 199.28 mcd at Site U1308, providing indications of slumping/disruption at ~ 196.88 mcd.

Figure 4. Site U1308: (a) Plot of the anhysteretic susceptibility (κ_{ARM}) versus volume susceptibility (κ) implying magnetic grain sizes of $\sim 1 \mu\text{m}$ according to King et al (1983), consistent with the hysteresis ratio plot (b) for Site U1308 core sections where values ($N=71$) lie close to the magnetite SD to MD grain-size mixing line (blue open squares), consistent with magnetic grain sizes of $\sim 1 \mu\text{m}$ (according to the calibration of Dunlop, 2002). Fields labeled SD: single domain, PSD pseudo-single domain and MD: multi-domain (after Day et al., 1977).

Figure 5. Site U1308: Component declination, inclination and maximum angular deviation (MAD) values computed for a uniform (20-80 mT) peak field demagnetization interval, plotted versus depth (mcd, meters composite depth) for the 100-270 mcd interval. Dark blue symbols are from u-channel samples from the composite section. Red symbols represent data from a single u-channel sample from Core section U1308C-20H-5 that records the Porcupine excursion. Black open squares represent data from discrete (7-cm³) cubic samples from the working half of core section U1308F-20H-2 that also record the Porcupine excursion. Light blue symbols at base of section are shipboard data after demagnetization at peak fields of 10 mT, no component data are available in this interval. Base: Expanded version of the 202-206 mcd interval.

Figure 6. Orthogonal projections of AF demagnetization data from u-channel measurements from core section U1308F-11H-5 that define the Gilsa excursion (a). Orthogonal projections of AF demagnetization data from u-channel samples that record the Porcupine excursion from core section 1308C-20H-5 (b), and core section U1308F-20H-2 (c). Orthogonal projections of AF demagnetization of discrete (7-cm³) samples from the working half of core section U1308F-20H-2 (d) that also records the Porcupine excursion. Red (blue) lines/symbols represent projection on the vertical (horizontal) plane. For u-channel samples (a,b,c) and 7-cm³ discrete samples (d) the displayed demagnetization ranges are 20-90 mT and 10-90 mT, respectively. The centimeter levels of each measurement in the core sections are indicated. Declinations are relative to the split face of the core sections. Magnetization intensity units are mA/m.

Figure 7. Site U1308 relative paleointensity (RPI) proxies computed as slopes of natural remanent magnetization (NRM) versus anhysteretic magnetization (ARM) after stepwise demagnetization (red), NRM demagnetization versus stepwise acquisition of ARM (blue), and NRM versus isothermal remanent magnetization (IRM) during stepwise demagnetization (green). All slopes determined for the 20-60 mT demagnetization/acquisition interval. Linear correlation coefficients (r) are shown for each slope using the same color codes.

Figure 8. Bivariate plots of slopes of NRM/ARM (relative paleointensity proxy) versus ARM intensity and $\kappa_{\text{ARM}}/\kappa$ (magnetite grain size proxy), showing no statistically significant relationship.

Figure 9. Site U1308 (1500-3350 ka) age model: Benthic oxygen isotope data (red) correlated to the LR04 oxygen isotope stack (black) of Lisiecki and Raymo (2005) with constant sedimentation rates between tie points. Certain marine isotope stages are labeled. Sedimentation rates beyond 3130 ka (light blue) derived from shipboard data.

Figure 10. Site U1308 (950-1250 ka): Virtual geomagnetic polar (VGP) latitudes compared to benthic $\delta^{18}\text{O}$ data with certain marine isotope stages labeled (see Channell et al., 2008; Hodell et al., 2008).

Figure 11. Site U1308 (1550- 2200 ka): (a) Virtual geomagnetic polar (VGP) latitudes compared to: (b) benthic $\delta^{18}\text{O}$ data (black) for the 1550 - 2200 ka interval, (c) benthic $\delta^{18}\text{O}$ data at DSDP Site 607 (red:

Raymo et al., 1989; Ruddiman et al., 1989), placed on the ODP Site 677 timescale of Shackleton et al. (1990) (see Raymo et al., 2004), with boxed intervals indicating original designations of marine isotopic stages (MIS) 63 and 73 (Raymo et al., 1989) with box width indicating the uncertainty in the recording of the top and base of the Olduvai Subchronozone, (d) benthic $\delta^{18}O$ data at ODP Site 677 (blue, Shackleton et al., 1990) and ODP Site 659 (green, Tiedemann et al., 1994) and LR04 (dashed black, Lisiecki and Raymo, 2005), all on their published age models. Black numbering refers to MIS labels for LR04 and Site U1308.

Figure 12. Site U1308 (2700-3150 ka): Virtual geomagnetic polar (VGP) latitudes compared to benthic $\delta^{18}O$ data with certain marine isotope stages labeled.

Figure 13. Site U1308 RPI record (red, NRM/ARM slope) compared with three stacks. Stack 1 (purple) comprises Sites U1308, 983 and 984. Stack 2 (light blue) comprises Sites U1308, U1307, 983 and 984. Stack 3 (NARPI-2200: thick dark blue) comprises Sites U1308, U1307, 982, 983 and 984. Gray shading indicates 2σ error associated with NARPI-2200. Yellow shaded bars indicate positions of reversals/excursions at Site U1308 (see Table 2).

Figure 14. The NARPI-2200 stack for the North Atlantic (blue) with gray shading indicating the 2σ error associated with the stack, compared with PISO-1500 (red, Channell et al., 2009), Sint-2000 (dashed in light green: Valet et al., 2005), PADM2M (dashed in dark green: Ziegler et al., 2011) and Epapis stack from the

Pacific (dashed in black: Yamazaki and Oda, 2005). Yellow shaded bars indicate positions of reversals/excursions at Site U1308 (see Table 2) with Gardar excursion from ODP Sites 983/984 (Channell et al., 2002).

Figure 15. Relative paleointensity proxy (NRM/ARM slope) for Site U1308 for 2100-3150 ka (blue) compared with the Site U1314 RPI record (red: Ohno et al., 2012) and the Epapis stack (dashed black: Yamazaki and Oda, 2005). Yellow shaded bars indicate positions of reversals/excursions at Site U1308.

Figure 16. The entire (0-3150 ka) RPI record for IODP Site U1308 (blue) scaled to virtual axial dipole moment (VADM), compared to the PISO-1500 stack (red). Scaling procedure follows the scaling for PISO-1500 (Channell et al., 2009). Site U1308 data for 0-1500 ka from Channell et al. (2008). Yellow shaded bars indicate positions of reversals/excursions at Site U1308 (see Table 2) with Gardar excursion from ODP Sites 983/984 (Channell et al., 2002).

Table 1. Tie-points used to compile the age model at Site U1308 for 100-250 meters composite depth (mcd).

Table 2. Age/depth of polarity reversals and excursions at IODP Site U1308 compared to ages given in other publications.

Table 3. Sites utilized in the North Atlantic stack (NARPI-2200) and the three other published stacks (Epapis, Sint-2000, PADM2M).

932 Supplemental Table 1. The NARPI-2200 relative paleointensity stack with
933 standard error.

Table 1. Age ties at Site U1308

Depth (mcd)	Age (ka)	Depth (mcd)	Age (ka)
100.77	1372.0	170.92	2250.3
102.29	1398.0	174.29	2273.7
103.05	1406.0	175.30	2290.0
103.80	1412.0	176.69	2309.5
104.05	1418.0	178.77	2332.9
104.76	1426.0	180.23	2349.9
105.72	1442.0	181.69	2373.3
106.40	1448.0	184.40	2388.4
106.55	1456.0	186.09	2406.0
108.80	1488.0	189.04	2428.8
109.13	1496.0	189.24	2434.6
110.08	1510.0	189.44	2463.3
112.59	1530.5	189.99	2487.4
113.84	1546.9	190.79	2500.4
115.51	1571.1	191.04	2520.6
117.03	1584.4	191.64	2540.1
119.12	1610.9	193.24	2553.8
119.84	1618.8	194.89	2571.4
120.34	1625.0	196.94	2594.8
120.84	1628.9	197.49	2601.9
122.09	1641.4	197.80	2650.6
124.01	1666.4	199.67	2681.2
128.59	1698.4	200.58	2689.4
129.25	1710.8	201.95	2704.4
130.84	1743.3	203.63	2730.6
131.57	1752.5	210.78	2799.4
135.42	1787.5	210.99	2805.0
136.34	1800.8	211.51	2819.5
138.11	1815.8	213.17	2840.6
139.09	1826.7	213.75	2847.7
140.43	1859.2	213.84	2857.0
140.74	1870.8	216.51	2864.8
145.26	1898.3	220.34	2876.6
146.12	1906.7	221.69	2893.0
146.85	1913.3	224.34	2913.3
148.87	1941.7	225.76	2936.7
150.35	1964.5	227.67	2957.0
152.06	1998.0	229.17	2967.2
153.16	2009.0	231.51	2982.0
157.13	2044.0	232.67	2999.2
157.74	2047.6	234.09	3015.6
159.64	2062.9	235.17	3025.8
160.86	2086.3	237.09	3040.6
162.21	2116.1	237.59	3048.4
162.63	2123.4	238.98	3059.9
164.28	2145.3	240.59	3084.9
165.38	2167.2	241.81	3092.7
167.22	2192.7	247.42	3134.9
168.28	2205.3	255.01	3207.0
169.88	2236.6	263.26	3330.0

Table 2. Age and position of excursions and reversals at IODP Site U1308

Reversal/excursion	MIS [1]	Depth (mcd) [1]	ka [1]	MIS, ka [2] (ka [3])	MIS, ka [4]	GPTS, ka [5] (ka, MIS [6])
Matuyama/Brunhes	19	50.3	774	19, 780	19, 780	781
Top Jaramillo	27	66.6	990	27, 990	28, 991	988
Base Jaramillo	31	72.9	1071	31, 1070	31, 1075	1072
Punaruu Ex.	31/32	76.2	1092			(1115, MIS 34)
Top Cobb Mt.	35	85.2	1187	35, 1190		1173
Base Cobb Mt.	36	86.2	1208	35, 1190		1185
Gilsa Ex.	54/55	116.9	1584			(1571, MIS 54)
Top Olduvai	63	134.9	1780	63, 1770	63, 1781	1778
Base Olduvai	73	147.8	1925	71, 1950	74, 1968	1945
Top Reunion	80	162.2	2116	(2140)		2128
Base Reunion	81	163.6	2137	(2150)		2148
Gauss/Matuyama*	104	196.3	2595	104, 2600	G2, 2608	2581
Porcupine Ex.**	G6/G7	204.0	2737			
Top Kaena	G21	235.8	3029	(3040)	G22, 3045	3032
Base Kaena	K2	244.7	3116	(3110)	KM2, 3127	3116
Top Mammoth		255.4	3220	(3220)	KM6, 3210	3207

[1] Channell et al. (2008) and this paper; [2] Shackleton et al. (1990); [3] Hilgen (1991a,b); [4] Lisiecki and Raymo (2005); [5] Lourens et al. (2004), Ogg (2012); [6] Channell et al. (2002).

*Affected by hiatus at the Gauss/Matuyama boundary (see text).

** Newly recognized excursion

Table 3. Sites used in North Atlantic RPI stack (1-5), and compared Pacific records

Core	Loc.	Lat.(°)	Long.(°)	WD (m)	Leng th (m)	Age Used (ka)	No. Rec.	Sed. Rate (cm/kyr)	Ref.
1) IODP U1308	N.Atl.	49°53'N	24°14'W	3871	77	1200-2200	1	8	1
2) ODP 983	N.Atl.	60°24'N	23°38'W	1983	112	1200-1889	1	16	2
3) ODP 984	N.Atl.	61°26'N	24°05'W	1648	112	1200-2150	1	12	2
4) ODP 982	N.Atl.	57°31'N	15°52'W	1135	20	1200-2200	1	3	4
5) IODP U1307	N.Atl.	58°30'N	46°24'W	2575	57	1200-2200	1	6	5
Sint-2000	Pac/Ind.	Various	Various			1200-2000	5	0.5-5	6
Epapis	Pac	12°S- 4°N	135°- 198°E	2326- 4623	~20	1200-2000	3	0.8-1.5	7
PADM2M	Various	Various	Various			1200-2000	8	0.5-5	8

References: (1) Channell et al., 2008; Hodell et al., 2008; this paper; (2) Channell et al., 2002; (4) Channell and Guyodo, 2004; (5) Mazaud et al., 2015; (6) Valet et al., 2005; (7) Yamazaki and Oda, 2005; (8) Ziegler et al., 2011.

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