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Invited review

Magnetic paleointensity stratigraphy and high-resolution Quaternary geochronology: successes and future challenges

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ABSTRACT

Magnetic paleointensity stratigraphy is used to detect variations in the strength of Earth's ancient magnetic field. Paleointensity studies have demonstrated that a dominantly dipolar geomagnetic signal can be recorded in a globally coherent manner in different types of sediments and in non-sedimentary archives, including ice core records and marine magnetic anomaly profiles. The dominantly dipolar nature of geomagnetic paleointensity variations provides a global geophysical signal that has come to be widely used to date Quaternary sediments. Despite the many successful applications of paleointensity-assisted chronology, the mechanisms by which sediments become magnetized remain poorly understood and there is no satisfactory theoretical foundation for paleointensity estimation. In this paper, we outline past successes of sedimentary paleointensity analysis as well as remaining challenges that need to be addressed to place such work on a more secure theoretical and empirical foundation. We illustrate how common concepts for explaining sedimentary remanence acquisition can give rise to centennial to millennial offsets between paleomagnetic and other signals, which is a key limitation for using paleointensity signals for geochronology. Our approach is intended to help non-specialists to better understand the legitimate uses and limitations of paleointensity stratigraphy, while pointing to outstanding problems that require concerted specialist efforts to resolve.

1. Introduction

Geomagnetic polarity reversals have been used widely in Quaternary geochronology (Fig. 1) because they result from a virtually synchronous global change in sign of the geomagnetic dipole. The geomagnetic polarity timescale (GPTS; e.g. Cande and Kent, 1995) serves as the backbone for the geological timescale for the last 200 Myr, and is based on temporally calibrated records of Earth's polarity history. Polarity reversals are accompanied by dramatic decreases in geomagnetic paleointensity (Fig. 1). Along with higher-frequency paleointensity variations within periods of stable polarity (Fig. 1), these major intensity changes can also provide a timescale that has come to be used widely in geochronology (e.g. Guyodo and Valet, 1996, 1999; Laj et al., 2000; Kiefer et al., 2001; Stoner et al., 2002; Stott et al., 2002; Valet et al., 2005; Yamazaki and Oda, 2005; Channell et al., 2009; Ziegler et al., 2011). The geomagnetic field is generated in Earth's fluid outer core, and is dominated by the dipole component, so that variations in field intensity have a strong global signal that can potentially be used to provide a high-resolution (millennial scale) timescale for chronostratigraphy. This temporal resolution contrasts with geomagnetic reversals (Fig. 1), which occurred only ~4–5 times per million years over the last ~31 Myr (Lowrie and Kent, 2004).

Determining the magnetic polarity of a given geological unit is straightforward, whereas, as argued below, determining the ancient geomagnetic field strength is not so simple. Given the increasing use of paleointensity estimation in Quaternary geochronology, we provide an overview for a general audience of how geomagnetic paleointensities are estimated. We then summarize the strongest lines of evidence for why such estimations appear to be robust, followed by discussion of some of the uses of geomagnetic paleointensity analysis in high-resolution Quaternary geochronology. This treatment is representative of the successes of paleointensity analyses of Quaternary sediments. However, despite these outstanding successes, challenges remain. The remainder of the paper is devoted to summarizing these challenges. Our overall aim is to help Quaternary scientists to understand better how sedimentary paleointensities are estimated, their potential chronostratigraphic value, and their limitations. We also point out problems that require concerted paleomagnetic

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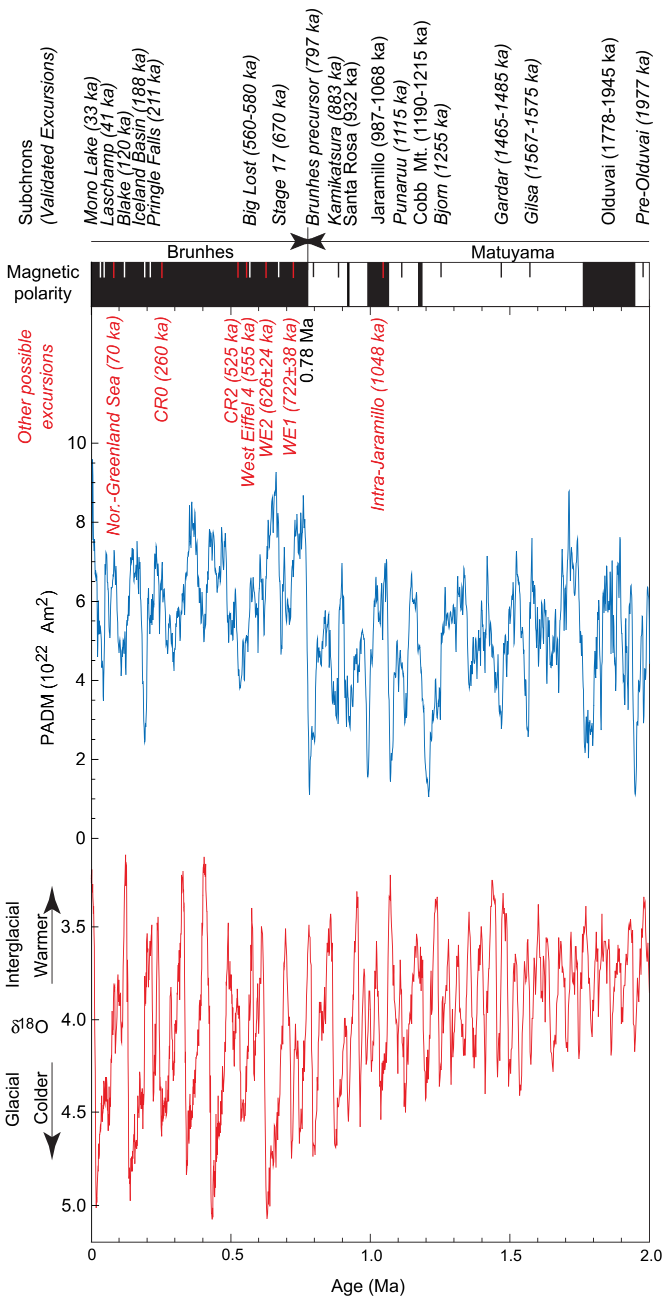

Fig. 1. Geomagnetic polarity timescale for the last 2 million years, with geomagnetic excursions, relative paleointensity variations and benthic $\delta^{18}O$ variations. Geomagnetic polarity is indicated at the top of the figure (black = normal; white = reversed polarity). Validated excursions (white) are indicated along with their respective ages above the polarity log in italics, with "possible" excursions (red) that have yet to be fully validated are indicated in red below the polarity log (Laj and Channell, 2007; Roberts, 2008). Each polarity reversal and excursion coincides with a paleointensity minimum. The paleomagnetic axial dipole moment (PADM) model (Ziegler et al., 2011) is used to represent paleointensity (blue). The climatic context of the geomagnetic variations is illustrated using the global stacked benthic $\delta^{18}O$ record (red) of Lisiecki and Raymo (2005).

effort to resolve. Such efforts will be valuable for improving our understanding of the geodynamo as well as aiding geochronological applications.

2. How does paleointensity determination work?

Reliable contemporary measurement of the intensity of the geomagnetic field has only been possible since the first measurements made by Gauss in 1835. Determination of geomagnetic field intensities for time periods preceding the 19th Century, therefore, requires paleomagnetic analysis of rocks or archaeological artefacts. Estimating the intensity of an ancient magnetic field is based on the assumption that the magnetization of a rock will be related linearly to the geomagnetic field strength. It is, therefore, expected that the natural remanent magnetization (NRM) of a sample will be related to the ancient field intensity (B_{anc}) as follows:

$$\mathrm{NRM} \cong \alpha_{anc} B_{anc},$$

where α_{anc} is a constant of proportionality. For certain igneous rocks, archaeological artefacts or other materials that have cooled from high temperatures, there is a robust physical theory and experimental protocol that enables determination of the absolute paleointensity from the recorded thermal remanent magnetization (TRM) (Thellier, 1938; Néel, 1955; Thellier and Thellier, 1959). Laboratory experiments are aimed at determining the proportionality of the TRM intensity to the geomagnetic field strength (Thellier and Thellier, 1959). After measuring the ancient TRM (TRM_{anc}) at a given temperature, a TRM can be imparted in the laboratory by heating the sample to the same temperature in a known applied laboratory field (B_{lab}). This enables determination of the laboratory constant of proportionality (α_{lab}). Assuming that α_{lab} is identical to α_{anc}, which can be tested with carefully designed experiments, the paleofield intensity can be determined from:

$$B_{anc} = (\mathrm{TRM}_{anc}/\mathrm{TRM}_{lab})B_{lab}.$$

While the laboratory normalization technique provides a theoretically grounded means of determining absolute ancient field intensities, suitable materials with thermal remanences are neither temporally continuous nor are they globally available. Young volcanic rocks are also notoriously difficult to date and only a small fraction of available material yields useful paleointensity data. Sedimentary sequences are, therefore, an attractive target for obtaining continuous records of geomagnetic paleointensity variations. However, identification of a robust procedure for laboratory normalization of a sedimentary NRM that is analogous to that for a TRM has proved elusive. The problem is that there is no simple means of determining α_{anc} to calibrate the relationship between the NRM of a sample and the strength of the magnetizing field. The magnetization of sediments is affected by the strength of the ambient magnetic field, the magnetic mineral that records the paleomagnetic signal, the concentration of this magnetic mineral fraction, its grain size and the mechanism by which the magnetization was acquired. An empirical approach has been developed for estimating paleointensities from sediments in which the NRM is normalized by an artificial laboratory-induced magnetization (Levi and Banerjee, 1976). The goal is to remove the influence of rock magnetic variations with non-geomagnetic origins, and to validate the record by imposing strict rock magnetic selection criteria. These criteria traditionally require magnetite to be the only magnetic mineral present and that it occurs within a narrow grain size and concentration range (King et al., 1983; Tauxe, 1993). While this empirical approach appears to work (see discussion below), its theoretical underpinning is complicated (e.g. Tauxe et al., 2006). The result is that, because we cannot determine absolute paleointensities from sediments, we seek to estimate relative paleointensity variations by minimizing the number of variables that contribute to the magnetization of the sediment under investigation.

Despite the lack of a first-principles theory for how sediments become magnetized, we can outline general principles by which normalized remanence records can be interpreted as having a geomagnetic origin. For relative paleointensity estimation to work, the mechanism by which the sediment is magnetized must be constant throughout a studied sequence. This requirement cannot be tested and many questions remain about how sediments acquire a magnetization (see discussion in Section 5). Furthermore, a key assumption in relative paleointensity studies is that there is a linear relationship between magnetization intensity and geomagnetic field strength. There is incomplete empirical evidence to support this assumption for most sedimentary environments, including marine environments. For details, readers are referred to recent in-depth reviews (Valet, 2003; Tauxe and Yamazaki, 2007). We are, therefore, in the uncomfortable position of having an inadequately grounded approach for relative paleointensity analysis. In Section 3, we discuss the empirical evidence for why normalized remanence records appear to be dominated by geomagnetic signals, despite our inadequate understanding of paleointensity recording by sediments. This is followed by a brief discussion of how paleointensity signals are used in Quaternary geochronology (Section 4). We then discuss sedimentary remanence acquisition in Section 5 and issues related to relative paleointensity normalization in Section 6.

3. Do sedimentary paleointensity estimates have a geomagnetic origin?

Despite uncertainties about the details of how sediments become magnetized, many credible relative paleointensity records have been recovered, which indicate that the geomagnetic field intensity varied in a globally coherent manner on millennial and longer timescales (Fig. 1). Three principal "smoking guns" give confidence that sediments can provide robust relative geomagnetic paleointensity estimates. These include: 1. global reproducibility; 2. cosmogenic radionuclides; and 3. ocean crust magnetization. We outline each of these lines of evidence below.

3.1. Global reproducibility

Collections of published relative paleointensity records have been stacked to produce estimates of global field intensity variations (e.g. Guyodo and Valet, 1996, 1999, 2006; Valet et al., 2005; Channell et al., 2009; Ziegler et al., 2011). Multiple records have been stacked in this way because of the overall global coherency of the recorded signals (Fig. 1) despite differences in the sedimentary environments from which the records were obtained. With development of increasing numbers of relative paleointensity records, such stacks have progressively worked back in time from the present to 200 ka (SINT-200; Guyodo and Valet, 1996) to 800 ka (SINT-800; Guyodo and Valet, 1999) to 2.0 Ma, including SINT-2000 (Valet et al., 2005) and PADM2M (Ziegler et al., 2011), to 3 Ma (EPAPIS-3 Ma; Yamazaki and Oda, 2005). In parallel with these developments, it has been recognized that stacking can affect the amplitude of paleointensity features when records have different chronological precision (e.g. Roberts et al., 1997; Guyodo and Channell, 2002) or where variable sedimentation rates and any smoothing associated with paleomagnetic recording cause attenuation of high-frequency features (Hartl and Tauxe, 1996; Guyodo and Channell, 2002; Roberts and Winklhofer, 2004). Thus, in addition to the above-cited paleointensity stacks, which tend to have coarse age control on glacial/interglacial timescales (but sometimes better), an additional family of records or stacks of records has developed, usually from relatively rapidly deposited sediments with millennial-scale chronological resolution. These include the North Atlantic paleointensity stack for the last 75 ka (NAPIS-75; Laj et al., 2000), the South Atlantic paleointensity stack (SAPIS; Stoner et al., 2002), the global paleointensity stack (GLOPIS; Laj et al., 2004), the North Atlantic ODP Site 983 record of Channell and Kleiven (2000) that spans the 700–1100 ka interval, and the paleointensity and stable isotope stack for the last 1.5 Myr (PISO-1500; Channell et al., 2009). For the NAPIS, SAPIS and GLOPIS stacks, millennial-scale chronology is achieved by correlating sediment physical properties into a tight, internally consistent stratigraphy, and using oxygen isotope stratigraphies to correlate millennial climatic events with those recorded in the Greenland GISP2 ice core. For the PISO-1500 stack, simultaneous correlation of oxygen isotope and paleointensity records reduces the degree of freedom for correlation based on either parameter alone. Regional and global reproducibility of multiple records provides a powerful argument for the robustness of relative geomagnetic paleointensity estimates from sediments.

3.2. Cosmogenic radionuclides

Cosmogenic radionuclides are produced by interaction of cosmic rays with Earth's atmosphere. Production of cosmogenic radioisotopes is modulated by variations in cosmic ray flux, solar activity and shielding by the geomagnetic field. Geomagnetic dipole moment variations are the most important modulator of production rate of cosmogenic radioisotopes, which varies inversely with field strength. Variations in production of cosmogenic radionuclides, including ^{14}C (half-life, $T_{1/2} = 5.73$ kyr), ^{36}Cl ($T_{1/2} = 300$ kyr), and ^{10}Be ($T_{1/2} = 1500$ kyr), from ice cores and sediments provide an independent measure of field intensity variations on a range of timescales (e.g. Frank et al., 1997; Raisbeck et al., 2006). These variations can be presented in terms of predicted relative paleointensity by assuming that all of the cosmogenic radionuclide production stems from variations in the geomagnetic field and by transforming production rate into relative paleointensity. Elsasser et al. (1956) used the simple formula $(Q/Q_0) \propto (M_0^{0.5}/M)$, where Q is the radionuclide production rate at a given dipole strength M, relative to initial values for both (Q_o, M_o). Lal (1988) modified the relationship, particularly for low field strengths; this modified relationship is normally used to convert ^{10}Be variations to predicted relative paleointensity variations (e.g. Frank et al., 1997). Good agreement between variations predicted for the paleomagnetic dipole from relative paleointensity data (Ziegler et al., 2011) and a sedimentary record of ^{10}Be production for the last 200 ka (Frank et al., 1997) (Fig. 2a) and Antarctic ice core ^{10}Be flux data (Raisbeck et al., 2006) across the Matuyama/Brunhes boundary interval (Fig. 2b) provide strong evidence for a common signal.

3.3. Ocean crust magnetization

Ocean crust provides a paleomagnetic record of geomagnetic polarity variations over the last ~160 Ma (e.g. Cande and Kent, 1995; Gee and Kent, 2007). Data from deep-towed magnetometer surveys over fast-spreading ocean crust reveal coherent short-wavelength anomalies (Gee et al., 2000). Inversion of these anomaly profiles enables estimation of crustal magnetization (blue curve in Fig. 3), which compares well with paleomagnetic dipole moments (red curve in Fig. 3) from the global paleointensity stack of Ziegler et al. (2011). Minor offsets between the two records (e.g. at about 400 ka) result from imprecisions in the respective age models, particularly the assumption of linear spreading rate for the magnetic anomaly stack. The overall excellent agreement between these continuous records of ocean crustal magnetization and sedimentary paleointensity provides a strong argument that they register a common geomagnetic signal.

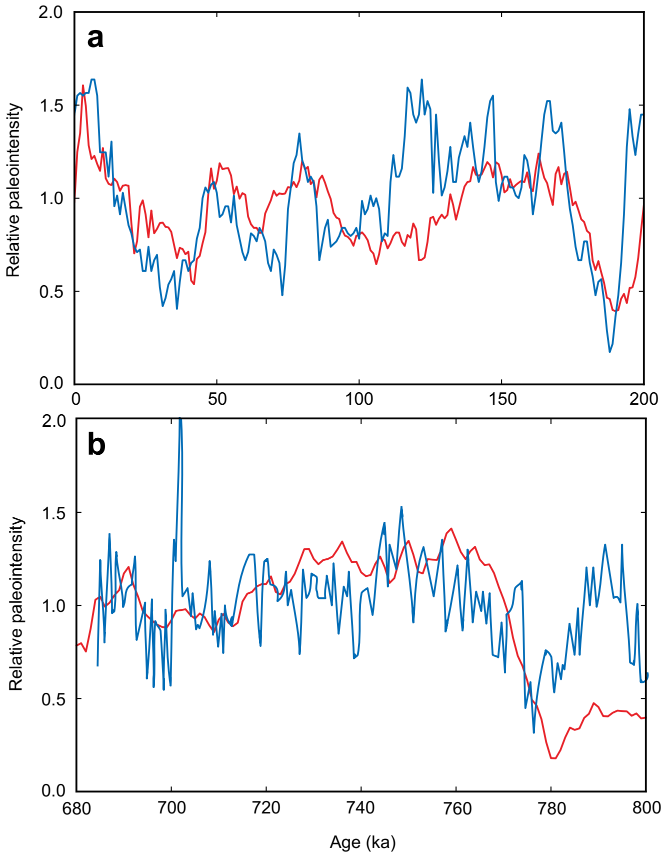

Fig. 2. Comparison of relative paleointensity and cosmogenic radionuclide production records for two time intervals. (a) Relative paleointensity (blue) predicted from normalized ^{10}Be stacks (from Frank et al., 1997) and relative paleointensity (red) from the PADM2M model (Ziegler et al., 2011) for the last 200 ka. (b) Same but with predicted relative paleointensity from ^{10}Be flux in the Antarctic EPICA Dome C ice core (blue) across the Matuyama–Brunhes boundary interval (from Raisbeck et al., 2006), normalized in same way as in (a). The PADM2M model (red) has relatively low resolution compared to the ice core records, but the same large-scale paleointensity features are evident in both the paleointensity and cosmogenic radionuclide production records. The data represented by the blue curves were recalculated so that both curves are directly comparable with the PADM2M dipole moment record (as discussed briefly in the main text).

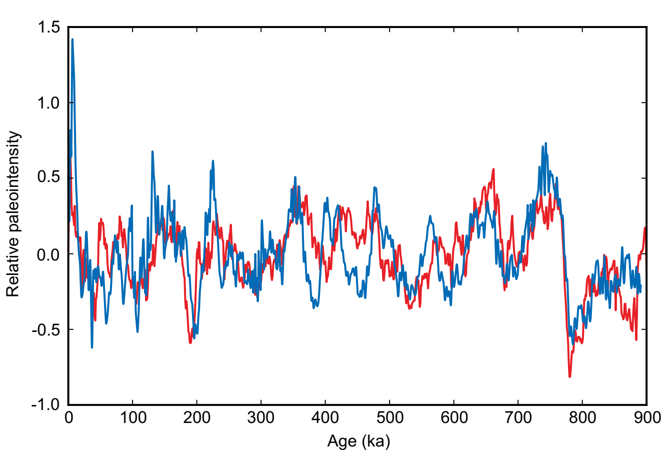

Fig. 3. Comparison of predicted relative paleointensity from the magnetization inverted from a high-resolution marine magnetic anomaly stack (blue; data from Gee et al., 2000) and dipole moments from the PADM2M paleointensity stack (red; Ziegler et al., 2011). Ages for the magnetic anomaly record were rescaled to a common age for the Matuyama–Brunhes boundary (see discussion in the text concerning age offsets of some paleointensity features).

4. Geomagnetic relative paleointensity in high-resolution Quaternary geochronology

Detailed relative paleointensity stacks now span the entire Quaternary (e.g. Valet et al., 2005; Yamazaki and Oda, 2005; Channell et al., 2009; Ziegler et al., 2011). Recognition that precisely dated, detailed paleointensity records have coherency on a global scale has led to the notion that geomagnetic relative paleointensity can be used to constrain the chronology of a sedimentary sequence. This concept is often referred to as paleointensity-assisted chronology (PAC). The key advantage of using paleointensity for high-resolution dating is that dipolar geomagnetic variations are globally synchronous, although it is difficult to test for isochrony at better than millennial scale. In particular, geomagnetic variations are independent of climatic parameters, including seawater chemistry, that are often used to synchronize marine sedimentary chronologies, and are therefore independent of the millennial-scale leads and lags that can affect climatic parameters. Much paleoclimate research is aimed at resolving such phase relationships and independent methods that can constrain such problems are extremely valuable in paleoclimate studies. Paleointensity variations have been used to resolve phasing issues associated with paleoclimate signals through marine isotope stage (MIS) 3, even in records in which there is clear manifestation of millennial-scale climate variability (e.g. Laj et al., 2000; Stoner et al., 2002). However, in other records in which this northern hemisphere climate variability is not evident, Channell et al. (2000) argued that paleointensity can provide a basis for inter-hemispheric correlation of marine sediments at a resolution that is difficult to achieve with δ^{18}O stratigraphy alone (Fig. 1). Paleointensity variations have, therefore, been widely used either as one parameter among others to constrain a chronology (e.g. Channell et al., 2000; Stoner et al., 2000, 2002; Kiefer et al., 2001; Stott et al., 2002; Evans et al., 2007), or, in lake or marine environments that lack suitable material for radiocarbon dating or for δ^{18}O stratigraphy, to develop an independent chronology that is based largely or entirely on paleointensity variations (e.g. Sagnotti et al., 2001; Brachfeld et al., 2003; Macrí et al., 2005, 2006; Willmott et al., 2006; Zimmerman et al., 2006; Lisé-Provonost et al., 2009). The success of the PAC approach is indicated by the fact that the scientific objectives of research cruises of the Integrated Ocean Drilling Program (IODP), which represents the largest international program in Earth and ocean science, have been based on the use of PAC to assess ice sheet–ocean–atmosphere interactions on millennial timescales (e.g. Shipboard Scientific Party, 2005).

As demonstrated above, the geomagnetic field varies in a globally coherent manner on millennial timescales (e.g. Channell et al., 2000; Laj et al., 2000, 2004; Stoner et al., 2002; Channell et al., 2009). Sedimentation rates in most marine environments are not high enough to enable temporal resolution at better than millennial timescales. However, centennial-scale paleointensity variations have been documented in exceptionally high-resolution datasets (e.g. Willmott et al., 2006; Lisé-Provonost et al., 2009; Barletta et al., 2010).

Our rapid summary of the development and use of PAC suggests that it has real promise for high-resolution geochronology. Despite the optimism that might be assumed from our treatment above, many difficulties remain with use of continuous relative geomagnetic paleointensity records. Signal variability in individual records often does not correlate well among records (or with stacks). To illustrate this, we plot for the last 800 ka the SINT-2000 stack (Valet et al., 2005) alongside the PISO-1500 stack of Channell et al. (2009) in Fig. 4a (renormalized to a mean of unity over the interval shown). Although there are similarities between these two stacks, as emphasized above, there are intervals, as shown in the boxes, where agreement is less impressive. These differences far exceed

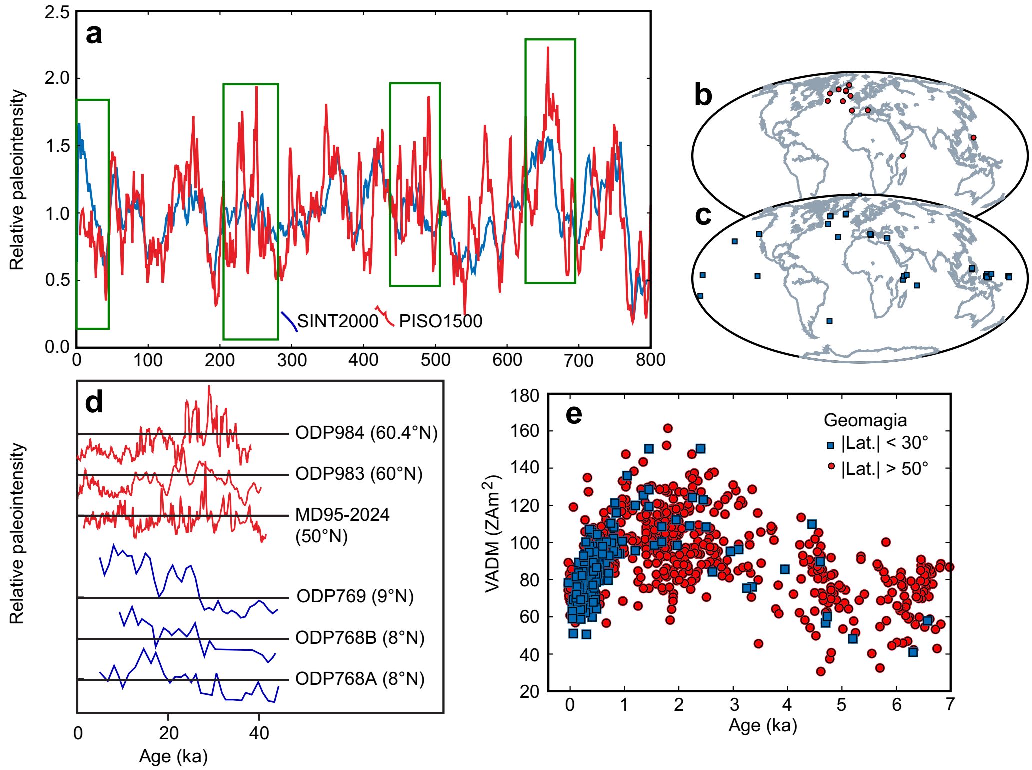

Fig. 4. Comparison of relative paleointensity stacks from different latitudes. (a) Comparison of SINT-2000 (Valet et al., 2005) with PISO-1500 (Channell et al., 2009) for the last 800 ka. Boxes denote periods with poor agreement between the two stacks. (b) Locations of cores in the PISO-1500 and (c) SINT-2000 stacks. (d) Examples of representative individual records from high (top three) and low (bottom three) latitudes (data are as compiled by Tauxe and Yamazaki (2007) and references therein). (e) Subset of the Geomagia50 database (Korhonen et al., 2008) with data from latitudes greater than 50° away from the equator and those from no more than 30° away. Lack of systematic latitudinal differences in the archaeointensity data for the last 50,000 years indicates that the observed differences in the paleointensity stacks are not due to non-dipolar effects.

measurement errors and any errors associated with stacking of multiple records. Why do such discrepancies occur and what do they mean? Do they result from imprecisions in chronological resolution, loss of signal amplitude due to stacking, or could they result from non-dipolar field behaviour (i.e. regional versus global signals, where the PISO-1500 stack is dominated by relatively high latitude records compared to the more global distribution of sites in the SINT-2000 stack (Fig. 4b, c))? One of the biggest differences among these records occurs for the last 40 ka (Fig. 4d). By comparing absolute paleointensities from the Geomagia50 database (Korhonen et al., 2008), we observe no profound difference between high (>±50°) and low (<±30°) latitude data (Fig. 4e). This suggests that a non-dipole explanation for the differences between the stacks is unlikely and that other explanations are necessary. Are they, then, due to variability in remanence acquisition mechanisms within the studied sediments? Do they result from imperfect normalization? Are there grain size variations that are not adequately normalized by standard approaches? What influence does diagenesis have on recording of paleointensity signals? Are there other problems? The temptation to throw away relative paleointensity analysis because of the existence of such problems should be avoided. The case for why relative paleointensity works has been made above. It should be remembered that marine $\delta^{18}O$ records can also be complicated due to variable freshwater inputs (e.g. ice melting, strong evaporation in marginal basins, monsoon variations and ocean stratification, etc.), and that synchronization of paleoclimate records can be blighted by many problems. Tuning and correlation remain a problem with age model construction in paleoclimate studies and standard tuning procedures gloss over attempts to independently determine phase relationships among climate signals of interest. It remains important to resolve the reasons that lie behind the limitations in relative geomagnetic paleointensity determinations in order to better understand the measured signals and to find solutions to the questions posed. The authors of this paper represent part of the spectrum of views in relation to the issues at stake and do not agree on all aspects. In the following treatment, we attempt to represent the spectrum of views in order to articulate the problems that need to be addressed through concerted research effort.

5. Sedimentary remanence acquisition

A major uncertainty concerning relative paleointensity studies relates to sedimentary remanence acquisition. While sedimentary remanence acquisition has been subjected to detailed experimental, theoretical and numerical investigation for over 60 years (e.g. Johnson et al., 1948; King, 1955; Nagata, 1961; Irving and Major, 1964; Kent, 1973; Verosub, 1977; deMenocal et al., 1990; Tauxe et al., 1996, 2006; Carter-Stiglitz et al., 2006; Heslop, 2007a; Tauxe, 2006; Liu et al., 2008; Mitra and Tauxe, 2009; Roberts et al., 2012), much remains unknown. Uncertainties surround the numerous variables that remain difficult to constrain fully. These relate to sedimentology, flocculation and pelletization of sediment particles, salinity, bioturbation, compaction, diagenesis, and differential contributions from biogenic and detrital magnetic minerals. We now describe known issues in relation to these variables and how they influence remanence acquisition and relative paleointensity studies.

5.1. Classical depositional and post-depositional remanent magnetizations

The journey of a magnetic particle begins as it settles through the water column and ends when it is locked into position in the sediment (Fig. 5). Magnetic particles suspended in water are spun about by turbulence, which acts to randomize the magnetic moments. Depending on the chemistry of the water (salinity and/or pH), magnetic particles will adhere to clays and clay particles will coalesce to form larger, less magnetic flocs (Shcherbakov and Shcherbakova, 1983; Lu et al., 1990; Katari and Tauxe, 2000; Tauxe et al., 2006) or fecal pellets. These flocs and pellets are exported to the bottom of the water column. During their descent they are subject to magnetic and hydrodynamic torques. The action of any hydrodynamic torque will dominate the magnetic torque for non-spherical flocs during settling, thereby contributing to an inefficient magnetization of the resultant sediment (Heslop, 2007a). The sediment/water interface is usually not well defined. The lowermost part of the water column is a zone in which the concentration of suspended sediment increases dramatically (Fig. 5). This zone is referred to as the nepheloid layer or the benthic boundary layer, in which sediment remains in suspension due to friction associated with movement of bottom waters over the sediment substrate. At the base of the nepheloid layer, there is a higher density of fluffy floc-rich suspended matter, below which occurs the uppermost part of the sediment column. The sediment is water rich and is actively mixed by organisms (bioturbation) who repeatedly ingest, excrete and re-suspend the material until it finally joins the more consolidated sediment layer below, in which the probability of re-suspension drops dramatically. As the sediment dewaters and compacts, magnetic minerals may undergo further rotation. Finally, magnetic minerals may undergo chemical change(s) during diagenesis, either growing from non-magnetic precursory phases, changing from one magnetic phase to another, or dissolving altogether. Stages of remanence acquisition have traditionally been described as a depositional remanent magnetization (DRM), where magnetic particles rotate freely in aqueous solution, a post-depositional remanent magnetization (pDRM), where magnetic particles respond to magnetic torques in the consolidating zone, and a chemical remanent magnetization (CRM), which is acquired when magnetic particles grow through a critical blocking volume during diagenesis.

When considering DRM and pDRM acquisition, it is important to consider how magnetic particles can realign with the ambient geomagnetic field. The torque on the magnetic moment $\mathbf{m}$ of a particle by the field $\mathbf{B}$ is: $\mathbf{m} \times \mathbf{B}$, or $\mathbf{mB} \sin \alpha$, where α is the angle

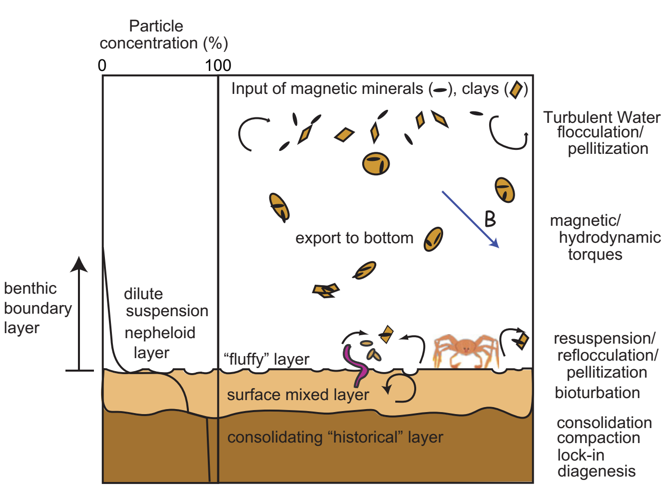

Fig. 5. Schematic illustration of the journey of magnetic particles (black ellipsoids) from the water column to burial with the processes that are likely to contribute to eventual recording of the paleomagnetic field by the particles (modified from Tauxe and Yamazaki (2007)). See discussion in the text for a more detailed explanation of the processes depicted.

between the two vectors. The magnetic moment will rotate to bring $\mathbf{m}$ into alignment with $\mathbf{B}$, but this motion will be opposed by the viscosity of the aqueous fluid. Nagata (1961) solved the equation of motion and demonstrated that the time constant of alignment (τ) of a particle with the ambient field can be approximated by:

$$\tau = \frac{\lambda}{\mathbf{mB}} = \frac{6\eta}{\mathbf{MB}},$$

where λ is the viscosity coefficient opposing the motion of the particle through the fluid (defined as the surface area of the particle times the viscosity of the fluid η), and $\mathbf{M}$ is the magnetization (moment normalized by volume). Choosing reasonable values for magnetic minerals, fields and fluid viscosities, a magnetic particle will align fully and almost instantaneously with the magnetic field. Simple DRM theory, therefore, predicts that for sediments composed of isolated magnetic particles, the particles should have magnetic moments that are fully aligned with the geomagnetic field. As a result, a DRM should be insensitive to changing field strengths. However, as we have demonstrated above, sedimentary remanences record a strong signal due to field strength variations. The magnetic torque on isolated particles can, therefore, only be one part of a complex story.

Other factors are also important for DRM and pDRM acquisition. When a magnetic particle falls through a fluid (air or water), it can roll when it impacts the sedimentary substrate. The net effect is that the recorded paleomagnetic inclination will be systematically shallow (e.g. Johnson et al., 1948; King, 1955; Tauxe and Kent, 2004). Inclination shallowing is, therefore, often associated with a DRM. In contrast, many sediments provide superb recording of the expected time-averaged geomagnetic field for a geocentric axial dipole (e.g. Opdyke and Henry, 1969). Such sediments are often bioturbated, which suggests that bioturbation lessens the density of particle packing, which then allows the geomagnetic field to exert a torque on magnetic particles to realign them with the field after the last mixing event, thereby giving rise to a pDRM (Irving and Major, 1964; Kent, 1973). Subsequent compaction of clay-rich sediments can give rise to an additional type of inclination shallowing (e.g. Anson and Kodama, 1987; Arason and Levi, 1990). Both types of inclination shallowing can be corrected for (Tauxe and Kent, 2004). Regardless, compaction-induced inclination shallowing is not normally observed above depths of 60–85 m below the sediment/water interface (Arason and Levi, 1990) or even at depths of hundreds of metres (Anson and Kodama, 1987). This phenomenon is, therefore, not normally important for Quaternary settings of relevance here. Overall, the poorer recording fidelity expected for a DRM is less desirable than the recording expected for a pDRM. Regardless, other complexities need to be considered before accepting the legitimacy of either the classically defined DRM or pDRM concepts (Fig. 5).

5.2. Salinity and flocculation

The observed field dependence of DRM implies a much longer time constant of alignment than that predicted by Nagata (1961) for settling of isolated magnetic particles. There are several ways to accomplish this from a theoretical point of view (see Tauxe et al. (2006) for a review). The most promising approach, however, is that of Shcherbakov and Shcherbakova (1983) who recognized that in natural waters, particles adhere to each other (which results in "coagulation" or "flocculation"). Qualitatively, particles are drawn together by van der Waals forces. In deionized water, clay particles are surrounded by a double layer of ions that impart an electrical charge to the particles. The charges repulse one another, which keeps the clays apart in a stable colloid. Addition of salt (or other electrolytes) interferes with the double layer, thinning it, and allowing van der Waals forces to become important, thus making the particles more likely to adhere to one another. Flocs of such particles are composites of magnetic and non-magnetic minerals and have a much lower net magnetization than isolated magnetic particles. The lower resulting **M** in the equation of Nagata (1961) increases the theoretical time constant of alignment with the ambient field.

The effect of embedding magnetic particles within flocs is illustrated in Fig. 6a. Numerical simulations predict that small flocs (5 μm) will be essentially fully aligned with the applied field (i.e. the curves are saturated), but the magnetic remanence of larger flocs (25 μm) will be far from saturation and will have the expected quasi-linear dependence of DRM with respect to **B** (Tauxe et al., 2006). This simple floc model, which has a single magnetic moment within each floc (similar to the model of Katari and Bloxham, 2001), needed to be modified to explain the results of laboratory re-deposition experiments (Tauxe et al., 2006). In the model of Tauxe et al. (2006), flocs coalesce to form compound flocs (Fig. 6b inset) so that saturation is never achieved in weak Earth-like fields.

Van Vreumingen (1993) demonstrated that, in addition to a strong influence of salinity on DRM intensity (Fig. 7), the degree of paleomagnetic inclination shallowing also depends strongly on salinity. The flocculation model of Tauxe et al. (2006), with spherical composite flocs, cannot account for this phenomenon. Flocculated particles are hydrodynamically different from isolated particles. They are porous, loose and fragile. Although flocs have irregular shapes and are not rigid, a useful first step is to model them as rigid ellipsoids, which undergo complex motion when settling in a fluid (Belmonte et al., 1998; Galdi and Vaidya, 2001). With this approximation, Heslop (2007a) demonstrated that with increasing floc size, the hydrodynamic torque increases at a much higher rate than the magnetic torque. For prolate ellipsoids with a particular aspect ratio, flocs become dominated by hydrodynamic torques at a critical size.

Mitra and Tauxe (2009) incorporated the approach of Heslop (2007a) into the flocculation model of Tauxe et al. (2006). The new model is conceptually similar to that of King (1955), where particles are treated as collections of "plates and spheres", but differs substantially in the processes involved. Instead of assuming two distinct magnetic grain shape populations, Mitra and Tauxe (2009) divided a continuous distribution of floc sizes into two groups: one small enough to respond mainly to magnetic torques (group **M**) and one large enough to be governed by hydrodynamic torques (group **H**). The net magnetic moment of group **M** is essentially parallel to the applied field while group **H** flocs are more influenced by hydrodynamic torques. The flocs attain hydrodynamic stability while settling (with long axes on average horizontal); the magnetic moments then attempt to align with the field. When the flocs are too large to maintain equilibrium with the field, their magnetization is essentially randomized. Therefore, the net magnetic declination of group **H** flocs tracks the field azimuth, but the net inclination is near zero (Fig. 8). The net magnetic moments of both groups of flocs contribute to the observed DRM (Fig. 8). In reality, a floc is not expected to behave according to the simple scheme of particles aligning through action of a magnetic torque, as envisaged in the model, but it would instead follow a complicated trajectory under the simultaneous influence of magnetic and hydrodynamic torques (Heslop, 2007a). However, the average behaviour of an ensemble of flocs can be approximated by this simple conceptual model. Mitra and Tauxe (2009) used this model to successfully simulate the laboratory results shown in Fig. 7.

A key assumption in sedimentary paleointensity studies is that DRM is linearly related to the applied field, yet laboratory re-deposition in known fields sometimes produces a non-linear DRM-applied field strength relationship (e.g. Fig. 6b; Tauxe et al., 2006). From theory, larger flocs will respond in a more linear fashion with applied field than smaller flocs (Fig. 6a), which reach saturation rapidly (although larger particle sizes are essentially randomly oriented). Given the importance of the linearity assumption to relative paleointensity studies, we note that laboratory re-deposition experiments suggest that these non-linear DRMs, expressed as a fraction of the saturation isothermal remanent magnetization (sIRM; i.e. the strongest remanence that can be acquired by a sample), are much larger than natural sedimentary DRM/sIRM ratios (which are usually a few % at most). Such DRMs will likely have a quasi-linear relationship with applied field.

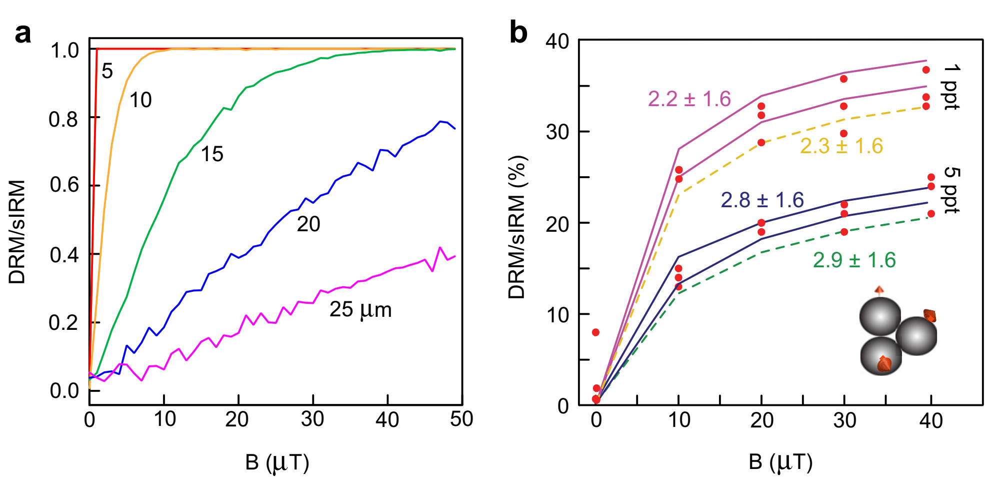

Fig. 6. Calculated results for magnetizations associated with flocs consisting of magnetic and non-magnetic particles. (a) Results of a simple numerical simulation whereby a magnetic moment is embedded in flocs of increasing radius (in microns) and solving for the net DRM after settling through 0.2 m of water. The DRM is expressed as a ratio of the saturation isothermal remanent magnetization (sIRM), which is unity when the magnetic moment of the floc is fully aligned with the ambient field. In this calculation, the floc has a single magnetic moment. (b) Results of laboratory re-deposition experiments as a function of field in a flocculating environment (red dots) for compound flocs (see schematic in inset). Salinities (in parts per thousand, ppt) are 1 and 5 for two different sets of experiments. Solid lines are from numerical simulations using distributions of compound flocs with assumed log-normal size distributions (mean and standard deviations are indicated in microns). The inset is an example of a compound floc with several smaller flocs (magnetic moments are red arrows) that coalesce to form a compound floc whose magnetization is the sum of the three individual moments. In contrast to (a), the more realistic scenario in (b), with compound flocs, never achieves saturation in the weak Earth-like ambient fields used for the experiment (modified from Tauxe et al. (2006)).

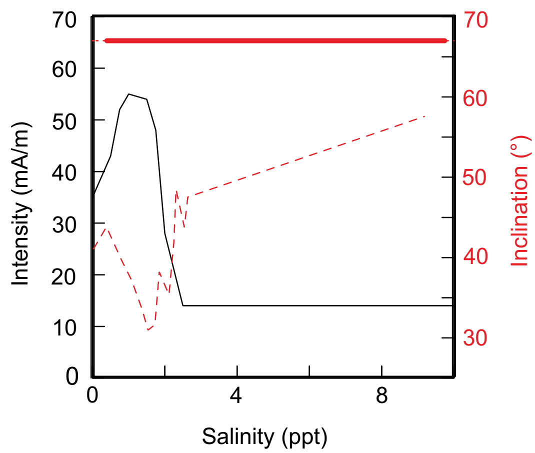

Fig. 7. Illustration of the effect of variations in water salinity on DRM intensity and paleomagnetic inclination. The relationship between DRM intensity and salinity for synthetic sediment composed of a mixture of kaolinite and maghemite is indicated by the solid black line, while inclination versus salinity is indicated by the dashed red line. The heavy red line represents the inclination of the applied field (67°) during the re-deposition experiment (data are from Van Vreumingen (1993)).

5.3. Critical assessment of the classical pDRM concept

In Section 5.1, we briefly described classical concepts concerning DRM and pDRM acquisition in sediments, and we introduced the important complications that arise from particle flocculation in Section 5.2. We now provide a more detailed, and critical, assessment of the evidence for pDRM acquisition in sediments. The pDRM concept is based on laboratory experiments in which wet sediments are re-deposited in a controlled ambient field, and progressively dried with or without stirring to simulate post-depositional sedimentary remanence acquisition processes (e.g. Irving and Major, 1964; Kent, 1973). The resultant magnetizations record no inclination error, which is consistent with paleomagnetic observations from bioturbated sediments. Additionally, it is argued that only small decreases in water content are necessary to lock-in the pDRM, which suggests that shallow lock-in with minimal delays in paleomagnetic recording might be expected to be common (Kent, 1973). The observed linear relationship between pDRM and applied field in such experiments (Kent, 1973; Verosub et al., 1979; Barton et al., 1980; Tucker, 1980) would also be expected to make such sediments suitable for relative paleointensity studies. Nevertheless, as indicated above, the pDRM/sIRM ratio in natural sediments is always much lower than in laboratory re-deposition experiments, which means that these simple and elegant experiments do not capture the full complexity of sedimentary magnetizations. Katari et al. (2000) argued that, because these re-deposition experiments were carried out with deionized water or with added anti-coagulants, which inhibited flocculation, they do not provide good analogues for understanding remanence acquisition in marine sediments.

Despite the fact that a pDRM is considered to be an important remanence acquisition mechanism, verification of the reality of pDRM and quantification of lock-in depth have proved elusive. Studies that have been widely cited as providing evidence for pDRM acquisition (e.g. Lund and Keigwin, 1994; Kent and Schneider, 1995) were critically assessed by Tauxe et al. (2006). In the case of Lund and Keigwin (1994), Tauxe et al. (2006) argued that paleomagnetic records from marine sediments at Bermuda Rise and lake sediments from Minnesota could be correlated without invoking smoothing due to pDRM acquisition. In the case of Kent and Schneider (1995), removal of one data set with a poor chronology removed the basis for invoking paleomagnetic smoothing associated with pDRM acquisition.

Several attempts have been made to quantify the pDRM lock-in depth for sediments. The pDRM lock-in depth is usually estimated by plotting the depth difference between two stratigraphic markers of known age versus sedimentation rate for a number of locations. deMenocal et al. (1990) compiled the depth differences between the Matuyama–Brunhes boundary and MIS 19 ($n = 9$) and concluded that the lock-in depth in marine sediments is, on average, about 16 cm. Tauxe et al. (1996) re-explored this

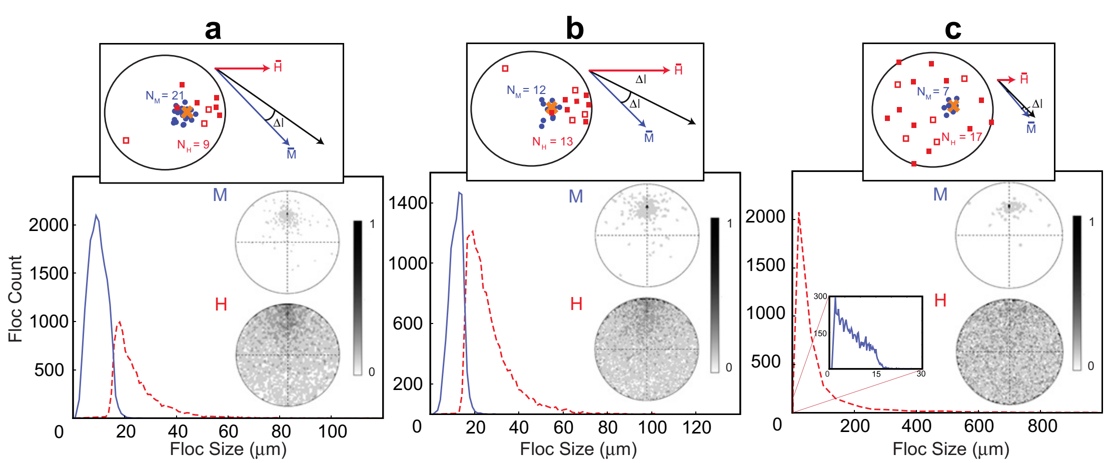

Fig. 8. Numerical simulation of how paleomagnetic intensity and inclination in sediments are affected by floc size. Lower: the solid blue and red dashed lines represent floc distributions in simulated **M** and **H** groups, respectively, where **M** represents flocs that respond to magnetic torques and **H** represents flocs that are dominated by hydrodynamic torques. A northward-directed applied field with 45° inclination and 45 μT intensity was used for the simulations. In equal area projections of floc moments for the **M** (upper) and **H** (lower) groups, no distinction is made between hemispheres; the lower plots are normalized by the maximum concentration of flocs. High and low concentrations are indicated by darker and lighter shading, respectively, and homogeneity (or lack of alignment) of moment direction is indicated by mid tone shading over the entire stereographic projection (as in c). For the **M** group, small dark areas indicate that the majority of moments are well aligned with the applied field. Upper insets: schematic representations of three cases. Left: blue dots and red squares in stereographic projections represent individual floc directions from the **M** and **H** groups, respectively. The orange cross is the applied field direction. Right: blue and red arrows represent the recorded paleomagnetic inclinations for the **M** and **H** groups. The black arrow is the resultant, and ΔI is the inclination flattening. (a) For small floc sizes, most flocs are in group **M**; few flocs are in group **H**, and ΔI is small. (b) For larger floc sizes, more flocs are in group **H** and ΔI increases. (c) For the largest floc sizes, most flocs are in group **H** and are so large that they are oriented randomly with respect to the field. The small number of group **M** flocs is sufficient for ΔI to become small. The net magnetic moment decreases from (a) to (c) because of the less efficient alignment of increasingly larger flocs (modified from Mitra and Tauxe (2009)).

relationship with a larger data set ($n = 19$) and concluded that the lock-in depth in marine sediments is negligible: 2.2 cm on average for all data, and only 1.0 cm if the most problematical data are removed from the analysis. Liu et al. (2008) subsequently demonstrated that such analyses depend crucially on construction of precise millennial chronologies. The foraminiferal $\delta^{18}O$ records used for such analyses depend on the oceanic water masses in which the foraminifera dwelt. The potential presence of different water masses at different water depths, in different ocean basins, and water-mass dependent differences in $\delta^{18}O$ signals in benthic and planktic foraminifera mean that such analyses can be affected by age offsets of the order of several kyr (e.g. Skinner and Shackleton, 2005). Liu et al. (2008) argued that such problems can be overcome by restricting inter-core comparisons to planktic–planktic or benthic–benthic $\delta^{18}O$ correlations for sediments deposited beneath the same water masses. This substantially decreases the available global data set for such comparisons. No suitably large global data set has been subsequently constructed. Any meaningful assessment of the pDRM lock-in depth requires analysis of paired local datasets. Liu et al. (2008) analysed one such pair of sediment cores with modern high-quality paleomagnetic and benthic–benthic $\delta^{18}O$ results from the North Atlantic Ocean (Venz et al., 1999; Channell and Kleiven, 2000), and concluded that the recorded pDRM has a lock-in depth of 23 ± 6 cm. Liu et al. (2008) concluded that most of the depth offset associated with the pDRM is associated with the depth of the surface mixed layer (due to bioturbation), which in this region is estimated at 10–20 cm (Thomson et al., 2000) and that the additional depth offset due to pDRM acquisition was only about 5 cm. This conclusion is consistent with the earlier analysis of Channell et al. (2004) and Channell and Guyodo (2004) from the same sequences, but based on systematic offsets in recording of the Matuyama–Brunhes boundary and of multiple paleomagnetic reversals through the Matuyama Chron. Tauxe et al. (2006) critiqued the approach of Channell et al. (2004) by arguing that the millennial-scale age offsets documented in the studied North Atlantic drift deposits are due to reworking of foraminifera. However, as argued by Liu et al. (2008), the $\delta^{18}O$ signals in question are carried by benthic foraminifera, which live within the sediment and are unlikely to have been reworked. We, therefore, consider the evidence from these North Atlantic sites to be indicative of pDRM acquisition.

Other evidence for pDRM acquisition is outlined as follows. Sagnotti et al. (2005) analysed two sediment cores from the Gulf of Salerno, and correlated the cores using high-resolution environmental magnetic measurements. Within this tight stratigraphic framework, they observed that well-defined paleomagnetic directional features are variably offset with respect to correlative susceptibility features between the two cores. No clear pattern was observed, with paleomagnetic recording in one core leading the signal recorded in the other core in some stratigraphic intervals, and lagging it in others. The maximum amplitude of these offsets was between -12 cm and $+15$ cm. Sagnotti et al. (2005) concluded that such stratigraphically variable lock-in depths cause centennial-scale uncertainty in age models for such rapidly deposited sediments. The possibility of such age offsets needs to be taken into account when using paleomagnetic records for dating sedimentary sequences. Using more slowly deposited sediments, Suganuma et al. (2010, 2011) observed offsets between the relative paleointensity minimum associated with the Matuyama–Brunhes boundary and its precursor (e.g. Hartl and Tauxe, 1996) and a maximum in ^{10}Be flux in marine sediment cores. They reported a pDRM lock-in depth of 17 cm. Suganuma et al. (2010) argued that 15–17 cm offsets associated with pDRM lock-in can cause age differences of more than 10 kyr in these slowly deposited (1–2 cm/kyr) carbonate sediments. Suganuma et al. (2011) carried out forward numerical simulations and inverse parameter estimations and determined that the best-fit pDRM filter function, surprisingly, has a Gaussian form. Conventionally, pDRM lock-in has been assumed to develop with progressive sediment compaction and dewatering, which is expected to proceed in an exponential manner during early burial (Lambe and Whitman, 1969). The Gaussian filter function estimated by Suganuma et al. (2011) is the only empirically determined filter function reported to date, and suggests that compaction and dewatering may not be the most important early processes associated with pDRM acquisition, although they must be important in the lower part of the lock-in zone. This result also demonstrates that we know too little about pDRM acquisition.

If a pDRM is an important remanence acquisition mechanism, it would be surprising if the lock-in depth was to have a fixed value throughout a given sedimentary sequence. Sagnotti et al. (2005) provide a rare glimpse of this likelihood. The maximum depth of bioturbation in the surface mixed layer is expected to be a key determinant of the lock-in depth. Burrowers focus their activity on surface sediments where there is an ample supply of high-grade organic matter (Gage and Tyler, 1991). Trauth et al. (1997) demonstrated that it is the carbon flux that determines the bioturbational mixing depth, and that time-variable mixing due to changes in productivity and organic carbon flux to the seafloor are important for obtaining high-resolution chronologies from sediments. This is likely to be particularly important for high-resolution paleomagnetic studies of sediments, and remains poorly constrained. The potential influence of time-variable lock-in, as well as the influence of sedimentation rate variations (e.g. Roberts and Winklhofer, 2004), ought to be borne in mind when interpreting apparently continuous paleomagnetic records from sediments.

5.4. *Effects of pDRM acquisition on paleointensity signals: model results*

Based on the above evidence, which is not as extensive as one would expect given that a pDRM has been postulated as a realistic remanence acquisition mechanism for 50 years, we now provide results of numerical simulations to illustrate the potential effects of pDRM acquisition and lock-in depth variations. Any such treatment involves significant assumptions because much remains unknown about pDRM acquisition. Most pDRM models assume that remanence lock-in can only begin once substantial surface mixing has ceased. Bioturbation in the surface mixed layer will cause a delay between sediment deposition and remanence recording. The extent of the delay will depend on the sedimentation rate and the thickness of the surface mixed layer, which has been argued to have an average of $\sim$10 cm for marine sediments, with a standard deviation of 4.5 cm and minimum and maximum values of 2 and 30 cm, respectively (e.g. Boudreau, 1994, 1998). Various lock-in functions have been used in pDRM acquisition models (e.g. Løvlie, 1976; Hamano, 1980; Otofuji and Sasajima, 1981; Kent and Schneider, 1995; Meynadier and Valet, 1996; Bleil and von Dobeneck, 1999; Spassov et al., 2003; Channell and Guyodo, 2004; Roberts and Winklhofer, 2004; Suganuma et al., 2011). Remanence lock-in is assumed to result from progressive compaction and dewatering, with expulsion of interstitial water increasing inter-particle friction to overcome the realigning geomagnetic torque on a magnetic particle and fixing the magnetization. An exponential lock-in function (Fig. 9) has been used most commonly for pDRM modelling (e.g. Løvlie, 1974; Hamano, 1980; Otofuji and Sasajima, 1981; Kent and Schneider, 1995; Meynadier and Valet, 1996; Roberts and Winklhofer, 2004) because sediment consolidation is expected to proceed in an exponential manner in early stages (Lambe and Whitman, 1969). However, the

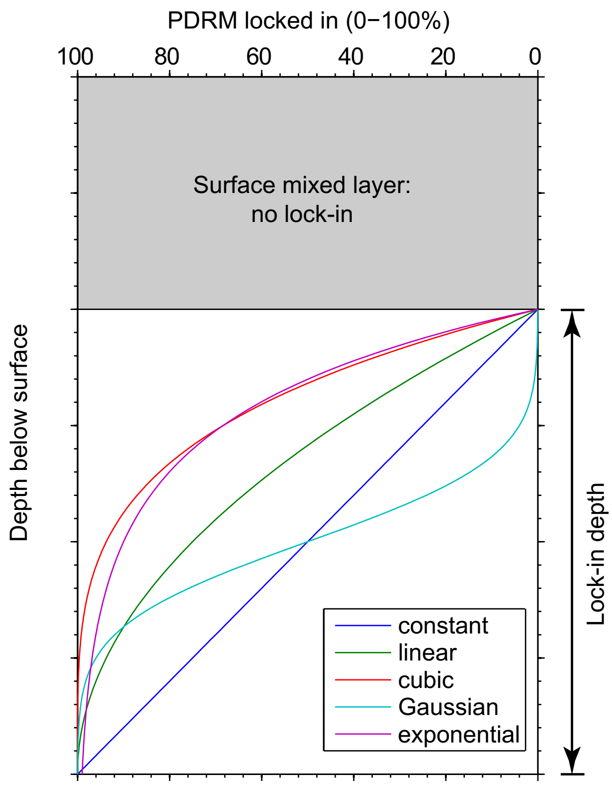

Fig. 9. Illustration of lock-in filter functions used for modelling pDRM acquisition in sediments. No pDRM lock-in is assumed within the surface mixed layer. The curves in the lock-in zone illustrate the cumulative lock-in for constant, linear, cubic, Gaussian and exponential functions. The pDRM is completely locked in at the base of the lock-in zone, except for the Gaussian and exponential functions, which never reach full lock-in (they are, therefore, truncated at +3.5 standard deviations and 99.9% at the base of the lock-in zone, respectively). Numerical pDRM calculations are made by dividing the lock-in zone into discrete depth slices, convolving the geomagnetic input signal with the lock-in function and summing the magnetizations at each depth interval (for details, see Roberts and Winklhofer (2004)).

only empirically determined lock-in function is the Gaussian (Fig. 9) function of Suganuma et al. (2011). For the purposes of illustration, we present model results for Gaussian and cubic lock-in functions in Fig. 10. There is no empirical or theoretical support for a cubic function. It has been used (e.g. Roberts and Winklhofer, 2004) simply because the pDRM locks in more rapidly and at shallower depths than with other commonly used functions (Fig. 9). It, therefore, constrains best-case recording scenarios, which can be used to understand and illustrate the likely minimum effects of signal filtering associated with pDRM acquisition.

We use the same approach as Roberts and Winklhofer (2004) to model pDRM acquisition and do not repeat the details of the procedure here. We use the high-frequency relative paleointensity signal from ODP Site 983 (Channell, 1999) as an input geomagnetic signal. The paleomagnetic signal at Site 983 is interpreted to be a pDRM (e.g. Channell and Guyodo, 2004; Channell et al., 2004). If so, it will be a filtered representation of the geomagnetic field. It is the highest resolution record available, so we use it as an input paleointensity signal to simulate scenarios with lower sedimentation rates than those at Site 983. In Fig. 10, we illustrate how the input paleointensity signal is recorded after filtering by Gaussian (Fig. 10a) and cubic (Fig. 10b) lock-in functions at different sedimentation rates for lock-in zones with 10 and 20 cm thicknesses, respectively. The Gaussian function is symmetric with 50% of the remanence recorded at half the lock-in depth (Fig. 9). In contrast, 50% of the remanence is recorded at ~16% of the lock-in depth for the cubic function and ~90% is recorded at half the lock-in depth. This is a much more efficient function; rapid lock-in produces much less temporal delay and less signal distortion or smoothing. The cubic function also produces complete lock in within the lock-in zone, whereas an exponential function approaches saturation asymptotically, but never reaches it, which is not physically realistic.

Increasing temporal offsets for progressively slower sedimentation rates and deeper lock-in depths (Fig. 10) illustrate the contrasting recording efficiency of Gaussian and cubic lock-in functions. These model results also illustrate the time offset that is introduced into paleomagnetic records by pDRM acquisition. When sedimentary paleomagnetic records are only used to construct magnetic polarity records, such offsets are usually unimportant. However, in the modern era in which continuous high-resolution paleointensity records are prized for providing chronological constraints to assess leads and lags between millennial scale climatic variations, such offsets become a crucial limitation and must be reckoned with as part of the age uncertainty of such studies. Our ignorance of details of the remanence acquisition mechanism, and its possible variability with time (e.g. Sagnotti et al., 2005), is a key limiting factor for paleointensity-assisted chronology. This ignorance illustrates the crucial importance of making concerted efforts to understand better the processes by which sediments become magnetized. Improved understanding of pDRM lock-in could then lead to the use of modelling to provide probabilistic assessment of uncertainties in high-resolution sediment age models. These uncertainties are reduced in environments with the highest sedimentation rates, but it is precisely such environments that are targetted to resolve details of climate forcing and response. Probabilistic assessment of uncertainties is applied routinely when constructing high-resolution radiocarbon chronologies (e.g. Ramsey, 2008), and could be readily adapted to bring greater rigour to uncertainty assessment in paleointensity-based age models.

5.5. The effects of sediment type on paleointensity signal recording

Sedimentological factors are potentially important for paleointensity signal recording. For example, do clay-rich sediments have different recording characteristics than carbonate-rich sediments? Few studies have directly addressed this question (e.g. Carter-Stiglitz et al., 2006). In re-deposition experiments in known magnetic fields with variable water salinities, with and without deflocculants, Spassov and Valet (2012) tested for differences in magnetic recording quality for carbonate and clay-rich sediments. They found that flocculation effects are dominant in clay-rich sediments, while carbonates are less affected by flocculation and have more efficient magnetizations with a linear relationship between magnetizing field and magnetization. Carbonates, therefore, should be ideal paleointensity recorders. Nevertheless, generally slow sedimentation rates mean that carbonate paleointensity records are normally more smoothed and have lower resolution (e.g. Guyodo and Channell, 2002; and references therein). Spassov and Valet (2012) confirmed the importance of flocculation, and suggested that lithological factors are important for paleointensity signal recording. The likely importance of lithological effects means that further experimental work is needed to understand better any differences between paleomagnetic recording in carbonates and clay-rich sediments and the importance of flocculation in clay-rich sediments.

5.6. Diagenesis

Burial of organic matter can have a controlling impact on the magnetic properties of sediments. Assessing the influence and

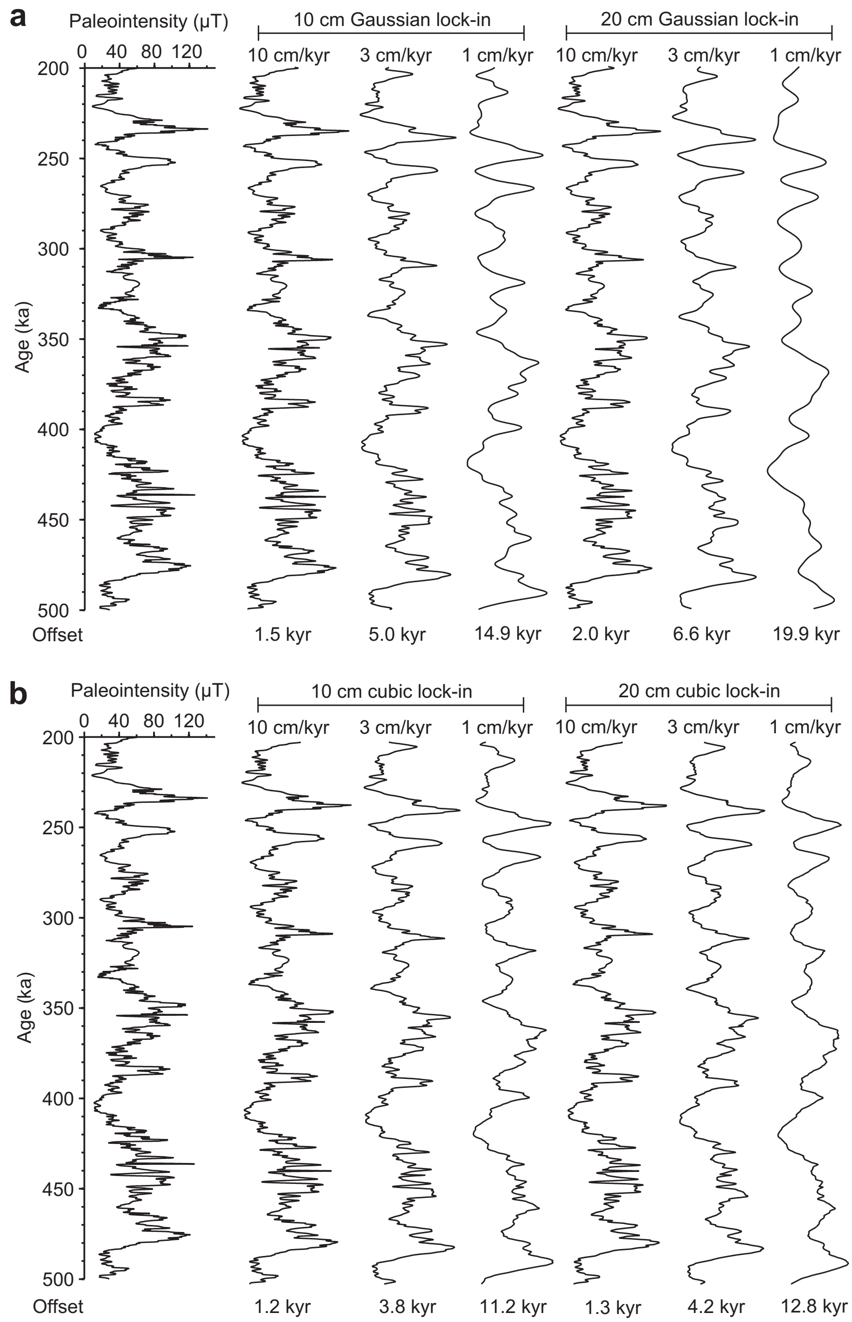

Fig. 10. Illustration of the delays and smoothing of paleointensity signals associated with pDRM recording. In each case, the high-resolution paleointensity record from ODP Site 983 (Channell, 1999) is used as the geomagnetic input signal. This signal is attenuated with delayed recording, as illustrated for a range of sedimentation rates for: (a) a Gaussian lock-in function (see Suganuma et al., 2011), and (b) a cubic lock-in function (see Roberts and Winklhofer, 2004). In each case, results are illustrated for lock-in depths (see Fig. 9) of 10 and 20 cm, with sedimentation rates of 1 cm/kyr (typical of pelagic carbonates), 3 cm/kyr, and 10 cm/kyr. See text for discussion.

extent of diagenesis is, therefore, a routine part of sedimentary paleomagnetic studies. When organic carbon flux is low, oxygen can diffuse into the sediment and organisms, including microbes, will consume or oxidize the organic matter so that organic matter diagenesis does not progress far. In such settings, oxygen diffusion into the sediment can partially oxidize detrital magnetite to form maghemite (e.g. Cui et al., 1994) or titanomaghemite (e.g. Xuan and Channell, 2010) shells on the surface of (titano)magnetite grains. The paleomagnetic complexities that result from oxidation can make oxic sediments unsuitable for paleointensity studies (e.g. Xuan and Channell, 2010). Regardless, generally low sedimentation rates in oxic sedimentary environments make them an infrequent target for paleointensity investigations. In settings with moderate organic carbon fluxes, molecular oxygen is progressively consumed through microbial degradation of organic matter until virtually no oxygen remains (i.e. the suboxic diagenetic zone). Organic carbon degradation then proceeds though microbial use and eventual depletion or consumption of a sequence of the most efficient remaining oxidants. Nitrate and labile manganese oxides are progressively reduced, followed by iron in reactive iron oxides (these reactions all occur under suboxic conditions; see Froelich et al., 1979), followed by sulphate reduction and methanogenesis (under anoxic conditions). The depths at which these reactions occur depend principally on organic carbon flux and sedimentation rate, which vary from setting to setting.

Iron-bearing minerals are strongly affected by organic matter diagenesis (Berner, 1981). Magnetite and other iron oxides can undergo dissolution (e.g. Karlin and Levi, 1983, 1985; Canfield and Berner, 1987), while iron sulphides, including pyrite and greigite (e.g. Berner, 1984; Roberts and Turner, 1993), can grow authigenically at their expense. Dissolution of highly reactive iron-bearing minerals (i.e. ferric hydrous oxide, lepidocrocite and ferrihydrite) occurs in the iron reduction zone (Poulton et al., 2004), whereas magnetite dissolution is ubiquitous in the sulphate reduction zone (Canfield and Berner, 1987) and can lead to near-total dissolution of the paleomagnetically useful detrital magnetic mineral assemblage. Paleointensity analyses usually target sediments deposited at rates above 1 cm/kyr, and preferably tens of cm/kyr. At these sedimentation rates, diagenetic iron reduction and sulphate reduction can become paleomagnetically important. Many attempted paleointensity studies have been compromised by magnetite dissolution in sulphate reducing environments. It is impossible to obtain paleointensity results when no detrital magnetite population remains, so sulphate reducing diagenetic environments can be ruled out as useful targets for such studies. Of greater interest, therefore, is the effect of iron reduction on paleointensity signals.

Few studies have explicitly addressed the effects of iron reduction on paleointensity records. Tarduno et al. (1998) made detailed rock magnetic analyses across the iron reduction front in carbonate sediments from the Ontong-Java Plateau. They reported a shift in the distribution of hysteresis properties and a slight shift in the peak coercivity below the Fe redox front. They attributed this change to the occurrence of magnetite-producing magnetotactic bacteria within the sediment at this depth. If true, this observation will have fundamentally important implications for the timing of acquisition of the paleomagnetic signal, which Tarduno et al. (1998) referred to as a biogeochemical remanent magnetization. The small change in anhysteretic remanent magnetization (ARM) coercivity that they reported across this boundary, however, is unlikely to significantly affect relative paleointensity normalizations. Yamazaki and Solheid (2011) reported a similar shift in hysteresis properties to that reported by Tarduno et al. (1998) across the Fe redox front, which they attributed to dissolution of a surficial maghemite skin on magnetite particles. By comparing the paleointensity record from the studied core with other paleointensity records, Yamazaki and Solheid (2011) concluded that maghemite reduction at the Fe redox boundary does not significantly affect their relative paleointensity estimation. They also interpreted the lack of discordance between their paleointensity record and others as evidence for lack of a significant biogeochemical remanence as suggested by Tarduno et al. (1998). The possible influence of a biogeochemical remanent magnetization (Tarduno et al., 1998; Abrajevitch and Kodama, 2009; Roberts et al., 2011, 2012) is best gauged by direct observations, or by comparison with other paleointensity records. The proposed lack of influence of iron reduction on relative paleointensity estimation (Yamazaki and Solheid, 2011) needs to be tested at other locations because a coercivity shift for a magnetic mineral assemblage might normally be expected to influence relative paleointensity normalizations.

Overall, although diagenesis will affect all sedimentary paleomagnetic records to some extent, it can be argued that successful paleointensity studies are the least likely to be strongly affected by diagenesis. The requirement for excellent paleomagnetic stability and the need to pass strict rock magnetic selection criteria mean that environments that have been strongly affected by diagenetic magnetic mineral alteration will not normally yield robust paleointensity results. This probably also explains why relative paleointensity studies discuss diagenetic complications much less often than other sedimentary paleomagnetic studies.

5.7. Summary

Given the uncertainties described above, it might be surprising that relative paleointensity signals have so much apparently robust structure that can be globally or regionally correlated (see Sections 3 and 4). The inescapable conclusion from Section 5 is that much remains to be learned about how paleomagnetic, including paleointensity, signals are recorded by sediments. We urgently need to develop a better theoretical, numerical, and experimental understanding of the physics of sedimentary remanence acquisition to underpin high-resolution paleomagnetic studies of sediments.

6. Relative paleointensity normalization

As discussed in Section 2, relative paleointensities are estimated by normalizing the NRM of a sediment by an artificial laboratory-induced magnetization to cancel the influence of non-geomagnetic mineralogical variations on the NRM. Readers are referred to Levi and Banerjee (1976), King et al. (1983), Tauxe (1993), and Tauxe and Yamazaki (2007) for details of the rationale behind relative paleointensity normalization. It is well recognized that paleointensity normalization is a "brute force" method that will have complexities associated with various factors. In this section, we provide an up-to-date perspective on some of these complicating factors.

6.1. Flocculation and linear versus non-linear magnetic recording regimes

A fundamental assumption in relative paleointensity studies is that the recorded magnetization (DRM or pDRM) is related linearly to the strength of the magnetizing field. As stated in Section 2, we can only assume this and cannot test whether the assumed linear relationship holds. In Section 5.2, we described the likely importance of sediment flocculation for sedimentary remanence acquisition and presented model results (Fig. 6) for how flocculated composite aggregations of clay and magnetite particles respond to Earth-like magnetic fields. Tauxe et al. (2006) argued that such flocs respond in a significantly non-linear manner to the magnetizing field. They also suggested that current methods for normalizing sedimentary remanence records do not take into account changes in floc size and can, therefore, only be partially effective in isolating the geomagnetic contribution to the NRM. Tauxe et al. (2006) attributed the unexplained scatter seen in relative paleointensity records (e.g. Fig. 4) to unaccounted for changes in floc size within the analysed sediments. This conclusion indicates that we urgently require methods to assess the presence and size of flocs, particularly in clay-rich sediments, and a stronger empirical and numerical understanding of the effects of flocs on sedimentary paleointensity records. Such knowledge is needed to assess whether a particular floc size distribution will place a paleomagnetic record within the non-linear recording regime. Any such work is likely to require a level of sedimentological knowledge that is currently unusual in paleointensity studies and a fundamental shift in the way that paleointensity investigations are undertaken.

6.2. Effects of mixed biogenic and detrital magnetite populations

The rock magnetic selection criteria that have been developed for relative paleointensity studies (see Section 2) require that magnetite is the only magnetic mineral present and that it must occur within a narrow grain size and concentration range (King et al., 1983; Tauxe, 1993). Most rock magnetic methods that have been used to assess grain size variations in relative paleointensity studies are indicative of bulk magnetic properties rather than providing a detailed view of the presence and grain size distributions of different magnetic mineral components within the sediment. Stratigraphic variation in the magnetic contribution of different components can potentially bias paleointensity normalization. For example, with the advent and routine application of magnetic techniques that enable discrimination of different magnetic mineral components within a sediment (Roberts et al., 2000; Egli, 2004; Weiss et al., 2004; Kopp et al., 2006; Egli et al., 2010), magnetite magnetofossils (i.e. the inorganic remains of magnetite-producing magnetotactic bacteria) have been routinely identified in addition to a detrital magnetite fraction. Such magnetic mineral assemblages are particularly common in pelagic carbonate sediments (e.g. Abrajevitch and Kodama, 2009; Roberts et al., 2011, 2012; Larrasoaña et al., 2012). As pointed out by Roberts et al. (2012), this could have a significant impact on relative paleointensity estimations for several reasons. First, the potential presence of a stratigraphically variable biogeochemical remanent magnetization could cause variable recording fidelity with respect to any DRM or pDRM associated with the detrital magnetic mineral component. Second, the empirical framework for relative paleointensity determination applies to magnetite in the 1–15 μm size range (King et al., 1983; Tauxe, 1993). Magnetite magnetofossils occur in the 30–100 nm size range, for which there is no empirical evidence for linearity of NRM with respect to magnetizing field or with respect to the NRM fraction recorded by magnetite particles in the 1–15 μm size range. The fact that magnetofossils occur in the magnetically ideal single domain (SD) size range means that they will contribute more strongly than larger pseudo-single-domain (PSD) particles to the ARM, which is a widely used laboratory-induced magnetization for paleointensity normalization. If there are two different grain size fractions (detrital and biogenic) in a sediment, different, and potentially stratigraphically variable, concentrations of these components means that standard methods of paleointensity normalization will be complicated in ways for which there are currently no correction methods. Roberts et al. (2012) suggested that this might explain some difficulties encountered in paleointensity studies, particularly in pelagic carbonates.

6.3. Contamination of normalized remanence records by climatic variability

With increased publication of relative paleointensity records over the last 20 years, orbital (or near-orbital) periodicities have sometimes been documented within paleointensity signals (e.g. Channell et al., 1998; Yamazaki, 1999; Yamazaki and Oda, 2002), while others have shown that near-orbital periodicities are not coherent with climatic records (e.g. Tauxe and Shackleton, 1994). This has led to revival of the debate about whether the geomagnetic field is powered not only by an internal dynamo mechanism, due to thermal and compositional convection in Earth's electrically conducting liquid outer core, but also by an external orbital component (e.g. Malkus, 1968). Critical analysis of such claims has led to identification of orbital contamination of relative paleointensity signals due to failure of normalization procedures to remove subtle expressions of climatically forced lithological variations from the paleointensity signal (e.g. Guyodo et al., 2000). In other cases, claims of orbital control on the geodynamo have been based on spectral analyses (e.g. Yamazaki and Oda, 2002) that are not statistically significant and that have highly variable phase compared to the consistent orbital modulation expected for a genuinely orbitally forced signal (Roberts et al., 2003; Heslop, 2007b; Xuan and Channell, 2008). There is yet to be a convincing demonstration that orbital periodicities recorded in normalized remanence records reflect orbital energization of the geomagnetic field. Nevertheless, the fact that orbital periodicities have contaminated paleointensity signals indicates that imperfections in normalization methods need to be better understood (see detailed discussion by Xuan and Channell (2008)).

6.4. Magnetostatic interactions

A key empirical criterion for selecting sediments for relative paleointensity investigations requires that the magnetite concentration in a sediment should not vary by more than a factor of 10 (e.g. King et al., 1983; Tauxe, 1993). The purpose of this criterion is to avoid the effects of magnetostatic interactions among particles, which increase with larger magnetite concentrations, and can significantly affect paleomagnetic recording fidelity (e.g. Sugiura, 1979; Muxworthy et al., 2003; Heslop et al., 2006). By using first-order reversal curve (FORC) diagrams (Pike et al., 1999; Roberts et al., 2000), Yamazaki (2008) argued that non-interacting biogenic magnetite occurs commonly in pelagic carbonates, along with a more strongly interacting detrital magnetite component. Observed glacial–interglacial variations in the concentration of these components will have a stratigraphically variable effect on ARM acquisition, which could then affect ARM normalizations used for relative paleointensity estimations. This could also give rise to coherence between the normalized remanence and the normalizing parameter, which is known to contaminate some paleointensity records (see Section 6.3). While it is useful to assess whether magnetostatic interactions influence relative paleointensity normalizations, we argue that the approach of Yamazaki (2008) is probably not valid. The vertical spread in FORC diagrams that is used to assess the importance of magnetostatic interactions (Pike et al., 1999; Roberts et al., 2000) has a different origin in SD compared to multi-domain (MD) magnetic materials. Magnetostatic interactions among SD particles result from the close proximity of particles whose magnetic moments physically interact with each other. In contrast, PSD and MD materials give rise to an inherent vertical spread in FORC diagrams due to processes internal to a particle, such as vortex structures, domain wall interactions, and nucleation, annihilation and pinning of domain walls (e.g. Pike et al., 2001), rather than the particle–particle interactions assumed by Yamazaki (2008). We, therefore, argue that Yamazaki (2008) did not document different magnetostatic interaction regimes, but rather the inherent characteristics of two magnetite grain size distributions (biogenic and detrital). The issues that arise in relation to paleointensity signal recording then revert to the type discussed in Section 6.2. If our reasoning is correct, the effects of magnetostatic interactions remain largely un-assessed in relation to their potential importance in relative paleointensity investigations.

6.5. Summary

As outlined above, several phenomena can complicate relative paleointensity estimation. Paleointensity practitioners need improved information of several types to isolate these effects to avoid, or correct for, these effects. For example, routine sedimentological analysis of clay-rich sediments is needed to assess the presence and size distribution of flocs. Improved rock magnetic methods are needed to detect and screen for non-linear recording regimes associated with large flocs. Magnetic methods that enable determination of the grain size distribution of magnetic mineral components in a sediment need to be routinely used rather than bulk magnetic methods that cannot discriminate between different magnetite populations or their stratigraphic variations. Placing paleointensity analysis on a sound theoretical and experimental footing requires development of methods that can identify and correct for effects that cause non-ideal normalization.

7. Conclusions and future directions

Developments in relative paleointensity studies over the last two decades provide much new information about dynamic geomagnetic field behaviour and provide a reference signal that has become widely used for dating sediments, often at millennial or higher resolution. The continuity of data provided by sedimentary relative paleointensity studies provides a level of detail and temporal continuity that is unavailable from absolute paleointensity records. This detail demonstrates the global coherence and dynamism of field variability, with the field often collapsing to low values; these paleointensity minima are often accompanied by geomagnetic excursions (e.g. Laj and Channell, 2007; Roberts, 2008), which supports the view that excursions are much more frequent features of geomagnetic variability than was once thought (Fig. 1). Our knowledge of geomagnetic field behaviour would be much poorer without detailed relative paleointensity records.

Ongoing work will inevitably lead to development of global paleointensity stacks further back in time, with greater global coverage to enable assessment of dipolar versus non-dipolar signals, with greater resolution to understand the effects of smoothing on the paleomagnetic record and to reconstruct better the frequency spectrum of ancient geomagnetic variations. This task will require special cases with a combination of desirable factors, such as high deposition rates without significant reductive diagenesis, moderate concentrations of ideal magnetite particles, and superb chronologies, much like the outstanding records obtained from the North Atlantic Ocean (Channell and Kleiven, 2000; Channell et al., 2000, 2009; Laj et al., 2000).

Despite the successes of sedimentary paleointensity studies, we remain remarkably ignorant of the physical processes by which sediments record paleomagnetic signals. We urgently need new knowledge to enable us to understand how sediments record information about the intensity of the geomagnetic field. We need a better physical understanding of how sediments acquire a remanent magnetization (with theoretical, numerical and experimental approaches). We need robust rock magnetic methods to enable screening to help identify whether sediments under investigation can be expected to have a linear relationship between magnetizing field and the recorded magnetization. We need to find ways to assess and take into account the presence and size distribution of flocs in order to understand their effects on paleointensity normalizations. Much work remains to be done to place relative paleointensity analyses on a secure theoretical and empirical foundation.

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