CHAPTER THREE

Magnetic Stratigraphy in Paleoceanography: Reversals, Excursions, Paleointensity, and Secular Variation

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Contents

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1. Introduction

The magnetic field of the Earth is thought to result from a self-exciting dynamo in the Earth's outer core (e.g., Bullard, 1949): a product of electrical currents generated through the fluid motion of an iron alloy conductor. Paleomagnetism is the study of this magnetic field as preserved in geologic material. Originally recognized in China a few thousand years ago, the study and use of Earth's magnetic field has a long history. In Europe, detailed magnetic studies were ongoing during the Middle Ages (for more details on the historical development of geomagnetism, see Jonkers, 2003). The recognition that the Earth's magnetic field was of internal (not celestial) origin and generally dipolar (like a bar magnet) goes back to at least 1600 when William Gilbert published *De Magnete*. By that time, the use of the compass for navigation was becoming increasingly established (Jonkers, 2003). The development by Carl Friedrich Gauss (1838) of the spherical harmonics expansion to characterize the geomagnetic field provided a mathematical means by which the position of the poles could be predicted and the intensity of the dipole calculated. This gave the geomagnetic field tangible and predictable elements, which paved the way for the future science of paleomagnetism.

The realization that the Earth's magnetic field was at times substantially different, even reversed, occurred at the turn of the century (Brunhes, 1906). Yet, it took more than half a century of debate before observations of magnetic reversal were universally accepted and proposals for its use as a stratigraphic synchronization tool explored (Hospers, 1955; Khramov, 1955, 1957). However, it was not until the observation of the same reversal sequence in dispersed radiometrically dated volcanic outcrops (Cox, Doell, & Dalrymple, 1963), magnetic anomalies (Vine & Wilson, 1965; Pittman & Heirtzler, 1966), and marine sediments (Opdyke, Glass, Hays, & Foster, 1966) that field reversals were fully accepted. This confirmed continental drift and began a stratigraphic revolution with the development of the geomagnetic polarity time scale (GPTS) (Doell & Dalrymple, 1966) that continues to be refined and serves as the backbone of Cenozoic stratigraphy to the present day (see Opdyke & Channell, 1996; Gradstein, Ogg, & Smith, 2004).

Since the recognition that magnetic reversals could be recorded in marine sediments (Harrison & Funnell, 1964; Opdyke et al., 1966), the GPTS has been an instrumental tool for paleoceanographic research. For example, inter-comparisons of high southern latitude radiolaria with the paleomagnetic record (Opdyke et al., 1966; Hays & Opdyke, 1967) were seminal contributions, placing late Neogene to Quaternary radiolarian events and biozones in a robust magnetostratigraphic framework. Since the 1960s, biostratigraphic researchers have been acutely aware of the importance of correlating their observations to magnetostratigraphies. The importance of this rests on the fact that magnetic reversals are globally synchronous on geological timescales, and that they are environmentally independent events that can be recorded in both deposited and thermally cooled materials alike. Much of the focus on understanding time in the geologic past has centered upon the dating of reversals and the inter-calibration of the reversal record

with other chronological and stratigraphic tools (see Opdyke & Channell, 1996; Gradstein et al., 2004). For modern paleoceanography, the pioneering study of Shackleton & Opdyke (1973) presented the first example of modern marine stratigraphy with the inter-calibration of oxygen isotopes with the magnetic reversal (Matuyama/Brunhes) record.

Magnetic stratigraphy rests on the idea that the recorded magnetization of a rock reflects the behavior of the geomagnetic field. The fact that sediment deposited in water can record the geomagnetic field has been known for more than 50 yr (Johnson, Murphy, & Torreson, 1948). In the simplest case, the natural remanent magnetization (NRM) of sediment is aligned with the (geo)magnetic field and is a function of its intensity and direction at the time of deposition. In practice, many factors may work to modify the original geomagnetic input signal. Under favorable circumstances and with detailed diligence, some of these effects can be separated and others avoided so that an accurate paleomagnetic record is recovered.

Over the last decade, a significant amount of paleoceanographic research has focused on timescales much shorter than the typical interval between magnetic reversals. This has reduced the impact of magnetic polarity stratigraphy on paleoceanography, though not its fundamental importance as a stratigraphic tool (see Opdyke & Channell, 1996). Paleomagnetism has been working to keep pace and a significant new understanding of geomagnetic field behavior during times of constant polarity has begun to emerge. Essentially, it has been found that high amplitude, high frequency variations in the Earth's magnetic field occur over large spatial scales even during times of constant polarity. These changes can occur over a millennium or even less, with coherence on a regional and sometimes even global scale. New magnetostratigraphic opportunities over a range of temporal and spatial scales are emerging. Much of this contribution will outline the practical aspects of reconstructing the paleomagnetic record from marine sediments. In the latter part of this chapter, we will briefly discuss some of the recent observations on the Quaternary geomagnetic record that are being made and their uses as a stratigraphic tool for paleoceanographic research.

2. Background

2.1. Geomagnetism and Orientation

On a time-averaged basis, the geomagnetic field during times of stable polarity approximates a geocentric axial dipole (GAD). Evidence for the GAD hypothesis has been accumulated from the inclination distribution of deep-sea sediments (e.g., Opdyke & Henry, 1969; Schneider & Kent, 1990) and statistical studies of randomly distributed igneous rocks (e.g., Merrill & McElhinny, 1977; Mejia, Opdyke, Vilas, Singer, & Stoner, 2004). The dip of the field lines from horizontal in the vertical plane is known as inclination (I). Magnetic deviations from true geographic north are known as declination. Using the GAD hypothesis, inclination can be predicted as a function of latitude (λ) $\tan I = 2 \tan \lambda$ and declination would be zero. Yet, at any point in time the geomagnetic field is not a GAD. Deviations

from a GAD, ~25% of the present field, are generally not considered to be stationary (this is an active topic of research) and how these change with time are known as secular variations.

The field at any point on the Earth's surface is a vector (F), which possesses a horizontal component (H), which makes an angle (D) between the geographical North and the magnetic meridian (Figure 1). Declination (D) is the angle from the geographical North measured eastward and ranging from 0° to 360°. The inclination (I) is the angle made by the magnetic vector with the horizontal plane. By convention, it is positive if the north-seeking vector points below the horizontal plane (present Northern Hemisphere) or negative if it points above (present Southern Hemisphere). The vertical component Z is positive down. The horizontal component (H) has two components, one to the North (X) and one to East (Y) (Figure 1).

Spherical harmonic analyses show that the geomagnetic field is almost entirely of internal origin. Approximately 90% of the present field can be explained by a dipole inclined to the Earth's axis of rotation by ~11.5°. The present magnitude of the dipole is 7.8×10^{22} Am^2. Based on the current world magnetic model (WMM), the geomagnetic North Pole in 2005 was 79.74°N and 71.78°W and

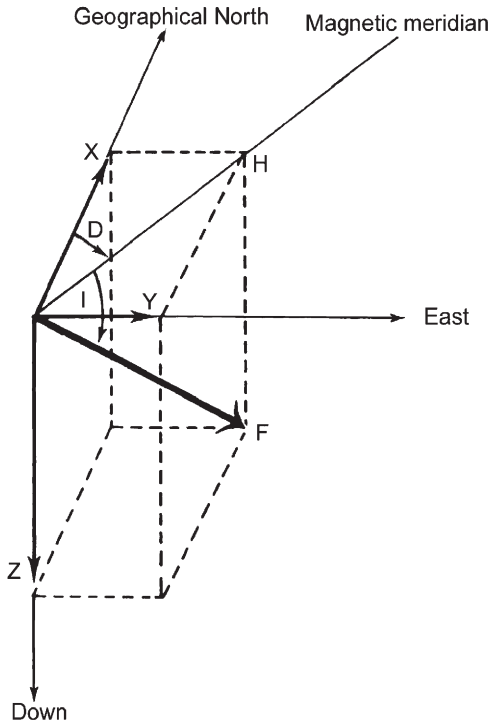


Figure 1 Earth's magnetic field. The total magnetic field is illustrated by the vector F. I = inclination; D = declination. The vertical component of the field is illustrated by Z. The horizontal component (H) of the field has two components, one to the North (X) and one to East (Y). Redrawn with permission from Thompson and Oldfield (1986).

the geomagnetic South Pole was 79.74°S and 108.22°E. These are based on the best fitting geocentric dipole. The great circle midway between the geomagnetic poles is called the geomagnetic equator. The actual magnetic poles (where inclination = 90°) and magnetic equator (where inclination = 0°) can deviate significantly. In 2005, based on the WMM model, the North Magnetic Pole was at 83.21°N and 118.32°W (Arctic Ocean, NW of Ellesmere Island, Canada) and South Magnetic Pole 64.53°S and 137.86°E (Southern Ocean, North of Antarctica, South of Australia).

2.2. Magnetism, Magnetic Units, and Conversions

Throughout this chapter, three types of magnetic units will be presented: one for the magnetic field, one for magnetization, and one for magnetic susceptibility. The motion of an electric charge will generate a magnetic field in the space around it. In accordance with Moskowitz (1991), as an analogy to an electron orbiting an atom, considering a loop of radius r and current i, the magnetic field is defined by $H = i/2r$ (A/m) at the center of the loop. The current loop has a magnetic moment, $m = i \times$ area (Am^2). The intensity of the magnetization (M or J), defined as a magnetic moment per unit of volume, $M = m/v$ (A/m), is the response of a material to a magnetic field passing through (note that M and H have the same units). The relationship between M in the material and the external field (H) allows the determination of the magnetic susceptibility, $k = M/H$ (dimensionless). Susceptibility is a measure of how magnetizable a material is in the presence of a magnetic field and can be used in a general way to describe the various classes of magnetic substances. The response of any material in that space is the magnetic induction (B). The unit of B is called the Tesla (T) and the total B field is the sum of the H field and the magnetization M. In the SI (Système International) system, $B = m_0(H+M)$, where the permeability of free space $m_0 = 4\pi\ 10^7\ \mathrm{Hm}^{-1}$ (H: Henry). Table 1 summarizes the different units in SI and cgs. SI is the standard, but because both types of units are often reported in the literature or directly given by the instruments, the relationships between the units of both systems are provided.

3. Soft Sediment Paleomagnetic Methods

3.1. Sampling

Sediment paleomagnetism is based on the requirement that the primary vector magnetization that is locked in the sediment at or shortly after the time of deposition is preserved in its original orientation, from the seafloor and through the laboratory analyses. Maintaining sediment orientation without deformation (physical or magnetic) is a non-trivial task. Assumptions are sometimes made. We will attempt to indicate, when, where, and what types of assumptions are commonly used for soft sediment paleomagnetism. For paleoceanographic uses, the first basic assumption is the original horizontality and lateral continuity of

Table 1 Magnetic Units and Relationship Between SI and CGS (centimeter–gram–second) Units.

Quantity	Symbol	SI	cgs	Relationship
Magnetic moment	m	Am^2	emu	$1\,Am^2 = 10^3$ emu
Magnetization	M	Am^{-1}	emu cm^{-3}	$1\,Am^{-1} = 10^{-3}$ emu cm^{-3}
Magnetic field	H	Am^{-1}	Oersted (oe)	$1\,Am^{-1} = 4\pi \times 10^{-3}$ oe
Magnetic induction	B	T	Gauss (G)	$1\,T = 10^4$ G
Permeability of free space	μ_0	Hm^{-1}	1	$4\pi\ 10^7\,Hm^{-1} = 1$
Magnetic susceptibility				
Volumetric	κ	Dimensionless	emu cm^{-3} oe	1 SI $= 1/4\pi$ emu cm^{-3} oe^{-1}
Mass	χ	$m^3\ kg^{-1}$	emu g^{-1} oe	$1\,m^3\ kg^{-1} = 10^3/4\pi$ emu g^{-1} oe^{-1}

Source: With permission from Tauxe (1998).
Note: Other relationships are as follows — $1\,H = kg\ m^2A^{-2}s^{-2}$, $1\,emu = 1\,G\ cm^3$, $1\,T = kg\ A^{-1}s^{-2}$.

the targeted sediment relative to the cored sediment water interface. The next assumption is that the coring process allows the acquisition of sediments that have not been stretched, compressed, or deformed in any way. Additionally, the corer should enter the sediment vertically and not rotate during penetration. Obtaining an undisturbed sample through coring or drilling is among the most difficult, yet critical components for retrieving a reliable paleomagnetic record.

Orientation control during coring is rare and/or often unreliable. Declinations are therefore most often relative. If enough time is involved, and 10,000 yr is often considered long enough (e.g., Merrill & McFadden, 2003), the assumption that the mean declination is zero is considered reasonable. One of the most critical (and easy), though sometimes neglected, measures is to split the whole core along a common plane. This can be easily done by drawing a consistent line along the core liner(s) and if there is more than one section, aligning each section together prior to insertion into the core barrel. Without this simple measure, declination must be (arbitrarily and undesirably) corrected for an unknown offset between each section, possibly degrading a critical part of the record. Nevertheless, even with these measures, the declination records are sometimes adjusted to compensate for slight rotations during coring, offsets caused by core splitting or for values close to 0° or 360°. Much of the subjectivity of these corrections, except for the latter, which is just a simple scale adjustment, can be removed by measuring duplicate or overlapping material.

There are only a handful of laboratories that have magnetometers capable of running whole or half-round core sections, therefore sub-sampling is almost always required and often desired. Half or whole-round magnetometers are generally used as survey instruments with the most notable and widely used example being the pass-through magnetometer on-board the RV JOIDES Resolution, the research vessel used by the Ocean Drilling Program (ODP) and more recently

by the Integrated Ocean Drilling Program (IODP). Having such a survey instrument is critical when dealing with large volumes of sediment (e.g., up to 8 km from a single ODP/IODP Leg). Initial shipboard observations provide an invaluable resource that allows later focused study of the sediments. In addition, shipboard observations are sometimes the only paleomagnetic analysis done on a sample, providing critical magnetic properties and geomagnetic polarity information that would otherwise be lost. On the other hand, sub-sampling is valued because these samples are taken from the pristine center part of the core, thus avoiding the disturbed outside part of the core section which can degrade whole or half round measurements (e.g., Acton, Okada, Clement, Lund, & Williams, 2002). In addition, measurements on sub-sampled material can be made at much higher resolution than half/whole round measurements. The width of the response function of a pass-through magnetometer is governed by the diameter of the sample measurement space. Each analysis from half/whole round magnetometer will, therefore, integrate over ~10 cm or more of stratigraphic length. More detailed studies of the NRM or laboratory magnetizations can be accomplished on sub-sampled materials at resolutions as tightly spaced as 1 cm.

3.2. Discrete Samples

Soft sediments are generally sampled using either discrete or u-channel samples (Tauxe, LaBrecque, Dodson, & Fuller, 1983). Discrete sampling of soft sediments is usually done using non-magnetic plastic cubes that are nominally 7–8 cm^3. There are several varieties of cubes with slight variations between them. If possible, truly cubic samples are more desirable. Mini-cubes of 1 cm^3 are sometimes used as they allow higher resolution sampling. However, the increased surface area to volume can result in more significant sample disturbance relative to cubes with larger volumes. Additionally, the significant reduction in material of a mini cube reduces the magnetic moment that may render weakly magnetized pelagic sediments below magnetometer sensitivity. When dealing with weakly magnetized materials, it can be advisable to demagnetize the cubes with an alternating field (AF) and measure their remanence prior to sampling so that the background can be subtracted.

3.3. U-channel Method

U-channel samples are collected by pushing rigid u-shaped plastic liners (2 × 2 cm cross-section) that are up to 1.5 m in length into the split halves of core sections (Figure 2). The u-channel is cut free using fishing line, removed from the core and capped with a matching plastic lid. End caps or waterproof polyethylene tape are often used to seal the ends. U-channels have significant advantages over discrete samples. First, u-channel sampling is much faster and introduces considerably less sediment deformation than back-to-back discrete sampling with standard 7 or 8 cm^3 plastic cubes. Second, the speed of data acquisition using u-channel samples makes it feasible to take continuous measurements on long or replicate sediment sequences that would be impractical, or even impossible, with discrete samples. Third, u-channel samples are optimal for a range of other high-resolution

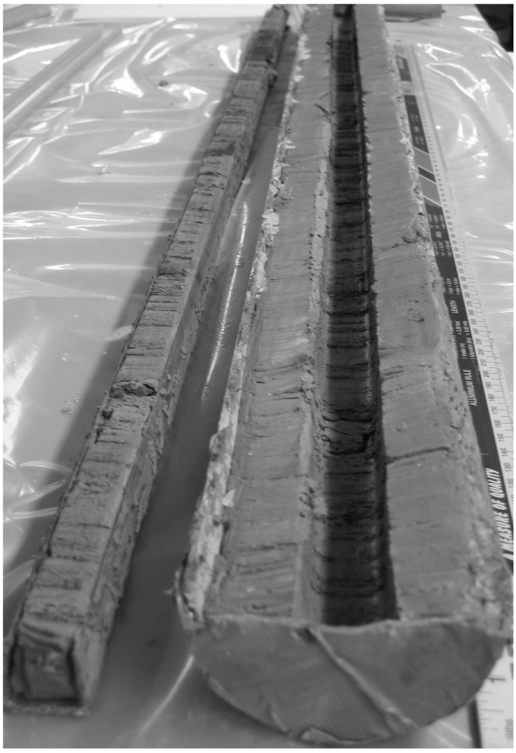


Figure 2 Example of u-channel sampling. U-channels are used for continuous sampling of sediment cores by pushing the open u-channel into the split surface of the core. A lid closes the u-channel and the ends are sealed with tape.

measurements, including computerized axial tomography (CAT-scan), X-ray density scan, gamma ray attenuation density scan, and X-ray microfluorescence scan (see St-Onge, Mulder, & Farncus, this book). These not only provide useful and complementary information for downcore interpretation of the paleomagnetic data, but also allow monitoring of sediment structures and integrity. Fourth, the u-channel sample can function as a permanent archive that takes significantly less space and uses less sediment than a traditional archive half section. Because the sample is completely enclosed in plastic, dehydration is minimized and with a little additional polyethylene tape essentially eliminated. The major disadvantage of u-channel samples results from their use for continuous measurements. Though samples are measured at 1 cm intervals, each measurement is not independent and is smoothed over ~4.5 cm. This results in edge effects that compromise the measurements on the top and bottom 5 cm of each section. Gaps result in a similar effect. Sediment heterogeneities are also not easily dealt with and anomalous sediment inputs, such as tephra layers for example, will affect the measurements as much as 6 cm on either side. Careful application of deconvolution procedures can minimize these effects (Guyodo, Channell, & Thomas, 2002) and are discussed below.

4. Magnetometers

As in many fields in Earth Sciences, paleomagnetic advances have been paced by technological innovations. The rock magnetometer is the required instrument for paleomagnetic research and its development has a long history. A critical advance was made with the development of the astatic magnetometer (Blackett, 1952) which allowed weakly magnetized rocks to be accurately measured. However, the sensitivity of this instrument came with the tradeoff of requiring an extremely quite magnetic and vibration free environment. The development of spinner rock magnetometers overcame some of these limitations by using the spin to amplify the magnetic signal. The high rate of spin required was, however, disastrous for soft marine sediments. The development of the slow spin fluxgate magnetometer (Foster, 1966) was a major breakthrough for marine paleomagnetism, as it allowed soft sediment to be measured without destruction. This initiated the merger of paleomagnetism and paleoceanography that was orchestrated by Neil Opdyke at Lamont Doherty Geologic Observatory in the late 1960s and 1970s (e.g., Opdyke, 1972), essentially pioneering all marine sediment paleomagnetic work that has followed. Spinner magnetometers are still widely used, although superconducting rock magnetometers are generally the preferred instrument for marine sediments because of their great sensitivity, speed, and measurement flexibility. For more details on magnetic instrumentation the reader is referred to Collinson (1983).

4.1. Superconducting Rock Magnetometers

The development of the superconducting rock magnetometer using a Superconducting Quantum Interference Device (SQUID) designed with a radio frequencies (RF) driven weak link sensor (Goree & Fuller, 1976) revolutionized paleomagnetism since weakly magnetized samples could be rapidly measured. These systems are cooled to cryogenic temperatures using liquid helium. Precise measurements are made by reading the flux changes generated by the insertion of the sample into the pick-up coil array. This resulted in substantially increased sensitivity and dynamic range. Sediments no longer had to be rotated at high speeds or vibrated so long-cores or fragile samples could be measured. The newer Direct Current (DC) SQUIDs developed in the late 1990 s improved the sensitivity almost two orders of magnitude. The DC SQUID sensor has a total magnetic moment sensitivity of $10^{-12}\,\mathrm{Am}^2$ which, when expressed in terms of magnetization, translates to $10^{-7}\,\mathrm{A/m}$ for a $10\,\mathrm{cm}^3$ sample. Further advantages of the DC SQUIDs are the improved dynamic range and faster measurement acquisition time. The most recent improvement to superconducting rock magnetometers is the elimination of liquid helium by using a pulse tube cryocooler. This allows the system to be warmed to room temperature or cooled to cryogenic temperatures with minimal effort or consequence.

4.2. Large Axis Pass-through Magnetometers

For marine sediment studies pass-through or long-core magnetometers are most practical because of their large throughput. We will therefore focus our discussion

on those types of instruments. The most widely used large axis pass-through magnetometer that can perform remanence measurements and AF demagnetization is the long-core cryogenic magnetometer (2G Enterprises model 760-R) on board the JOIDES Resolution. This instrument is equipped with a DC SQUID and has an inline AF demagnetizer capable of reaching a peak AF of 80 mT. The spatial resolution measured by the width at half-height of the pickup coils response is ~10 cm for all three axes, although they can sense a magnetization that extends up to 30 cm of core length (see Explanatory Notes Section of Channell, Sato, & Malone, 2005). The cryogenic magnetometer is sensitive to a magnetic moment of $\sim 10^{-9}$ emu or 10^{-7} A/m for a 10 cm^3 rock volume. However, the practical noise level is affected by the magnetization of the core liner ($\sim 8 \times 10^{-6}$ A/m) and the background magnetization of the measurement tray ($\sim 1 \times 10^{-5}$ A/m).

4.3. U-channel Magnetometer

The modern workhorse of marine paleomagnetism is the automated small access pass-through u-channel superconducting rock magnetometer (Figure 3). Since its development just over a decade ago (see Weeks et al., 1993; Nagy & Valet, 1993; Roberts, Stoner, & Richter, 1996; Verosub, 1998; Sagnotti et al., 2003; Brachfeld, Kissel, Laj, & Mazaud, 2004), our knowledge of the spatial and temporal variability of the Quaternary geomagnetic field has improved dramatically. The speed of data

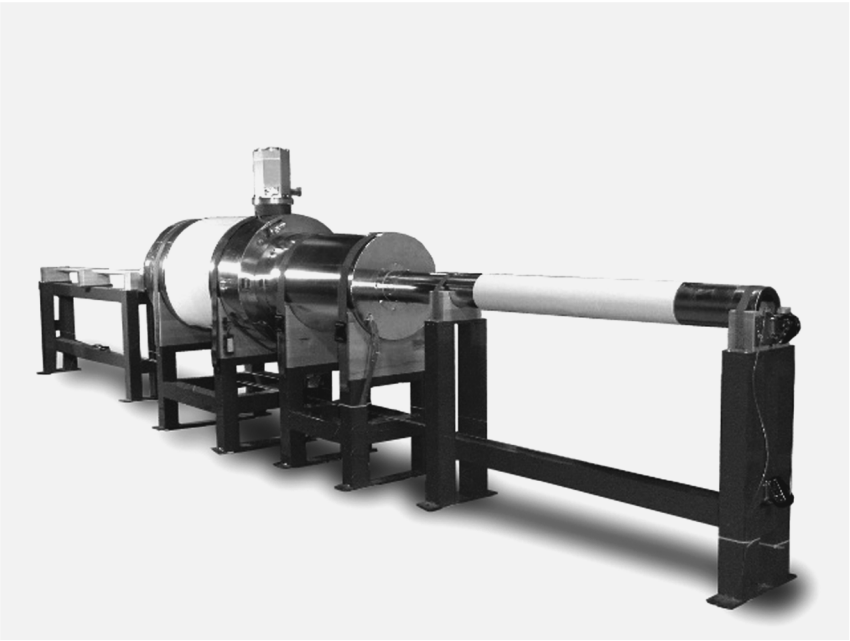


Figure 3 2G Enterprises automated small access pass-through u-channel cryogenic magnetometer. The system shown was recently installed at the University of Alberta, Canada. Picture courtesy of Bill Goree (2G Enterprises).

acquisition allows observations that could not practically be made with discrete samples. Continuous measurements at 1 cm intervals on long and/or replicate sediment sequences are becoming routine, as one measurement (e.g., one AF demagnetization step; see Section 5.1 for the use of stepwise demagnetization) of a 1.5 m long u-channel takes about 15 min.

Recent u-channel superconducting rock magnetometers fitted with DC SQUID are designed with a software controlled tracking system, in-line AF demagnetization coils capable of reaching a peak AF of 300 mT for the axial (*Z*) coil and 180 mT for the transverse (*X*, *Y*) coils (Nagy & Valet, 1993), in-line DC coil for acquisition of an anhysteretic remanent magnetization (ARM; see Table 2), and with a separate (or in-line) pulse magnetizer for acquisition of an isothermal remanent magnetization (IRM; see Table 2). The pick-up coils in the 2G Enterprises high-resolution cryogenic magnetometer are designed to have a narrow response function translating to a higher resolution measurement (see Weeks et al., 1993).

An important development of the long-core magnetometer is the in-line demagnetization process. A series of three AF coils are positioned in a mutually perpendicular fashion so that the sample can be demagnetized along its *X*, *Y*, and *Z* axes. The u-channel sample is demagnetized as it passes through the AF at a constant velocity. The peak AF is reached when the sample passes through or perpendicular to the coil and cycles down as it moves away, thus mimicking the cycling down of a standard discrete sample AF unit (Collinson, 1983). The entire process takes place in a magnetically shielded environment allowing coercivities lower than the peak AF to be randomized. Because of space restrictions, these coils cannot be operated simultaneously without interactions. Therefore, three passes (one for each perpendicular axis) are required at each demagnetization step. An in-line DC coil allows the ARM to be acquired using the axial (*z*) coil. It has been observed that translation speed can inversely affect the efficiency of ARM acquisition. Slow track speeds of 1 cm/s or less are most effective when acquiring an ARM on u-channel samples. Therefore, care should be taken when comparing ARM data generated with u-channel and discrete sample systems (Sagnotti et al., 2003), highlighting the need for standardization and calibration of paleomagnetic procedures.

5. Measurements and Magnetizations

Several types of magnetization are now routinely imparted and measured on both discrete and u-channel samples. These include ARM, IRM, and volumetric magnetic susceptibility. Short explanations and rule of thumb downcore interpretations of these magnetizations, as well as other magnetic parameters are given in Table 2. Below, we discuss some of the requirements for acquiring high quality paleomagnetic records.

5.1. Natural Remanent Magnetization-AF Demagnetization

Even though some sediments preserve a stable magnetization after little or no demagnetization, modern standards in paleomagnetic research dictate that

Table 2 Generalized Table of Downcore Magnetic Parameters and Their Interpretations.

Parameter	Interpretation
Bulk magnetic measurements	
Natural remanent magnetization (NRM): The fossil (remanent) magnetization preserved within the sediment. NRM recorded as declination, inclination and intensity.	Dependent on mineralogy, concentration, and grain size of the magnetic material as well as mode of acquisition of remanence, and intensity and direction of the geomagnetic field.
Volumetric magnetic susceptibility (k): A measure of the concentration of magnetizable material. Defined as the ratio of induced magnetization intensity (M) per volume to the strength of the applied weak field (H): $k = M/H$.	k is a first order measure of the amount of ferrimagnetic material (e.g., magnetite). k is particularly enhanced by superparamagnetic (SP) magnetite ($<0.03\,\mu$m) and by large magnetite grains ($>10\,\mu$m). When the concentration of ferrimagnetic material is low, k responds to antiferromagnetic (e.g., hematite), paramagnetic (e.g., Fe, Mg silicates), and diamagnetic material (e.g. calcium carbonate, silica) that may complicate the interpretation.
Isothermal remanent magnetization (IRM): Magnetic remanence acquired under the influence of a strong DC field. Commonly expressed as a saturation IRM or SIRM when a field greater than 1 T is used. A back-field IRM (BIRM) is that acquired in a reversed DC field after SIRM acquisition.	SIRM primarily depends upon the concentration of magnetic, principally ferrimagnetic, material. It is grain-size dependent being particularly sensitive to magnetite grains smaller than a few tens of microns.
Anhysteretic remanent magnetization (ARM): Magnetization acquired in a biasing DC field within a decreasing alternating field. Commonly expressed as anhysteretic susceptibility (k_{ARM}) when normalized by the biasing field used.	k_{ARM} is primarily a measure of the concentration of ferrimagnetic material, however it is also strongly grain-size dependent. k_{ARM} preferentially responds to smaller magnetite grain sizes ($<10\,\mu$m), and is useful in the development of grain-size dependent ratios.
Constructed magnetic parameters	
The "hard" IRM (HIRM): This is derived by imparting a back-field of typically 0.1 or 0.3 T on a sample previously given an SIRM. The	HIRM is a measure of the concentration of magnetic material with higher coercivity than the back-field. This commonly gives information on the

Table 2. (*Continued*)

Parameter	Interpretation
resulting BIRM, which has a negative sign, is used to derive the HIRM by the formula: HIRM = (SIRM+BIRM)/2.	concentration of the antiferromagnetic (e.g. hematite) or very fine-grained ferrimagnetic (e.g. magnetite) grains depending on the back-field used.
S-ratios: These are derived by imparting a back-field of typically 0.1 or 0.3 T on a sample previously given an SIRM. The resulting BIRM, which has a negative sign, is normalized by the SIRM; S = BIRM/SIRM. This provides a measure of the proportion of saturation at the back-field applied.	The S-ratios can be used to estimate the magnetic mineralogy (e.g., magnetite or hematite). Downcore variations may be associated with changing mineralogy. Values close to −1 indicate lower coercivity and a ferrimagnetic mineralogy (e.g., magnetite); values closer to zero indicate a higher coercivity, possibly an antiferromagnetic (e.g., hematite) mineralogy. S-ratio with a back-field of 0.1 T may be sensitive to mineralogic and grain-size changes whereas the S-ratio with the 0.3 T back-field is more sensitive to mineralogical changes (e.g. proportion of magnetite to hematite).
Pseudo S-ratios: These are derived by imparting an IRM typically of 0.3 T, followed by an SIRM (e.g., 1 T). The pseudo S-ratio is determined by dividing the IRM by the SIRM.	Similarly to the classical S-ratios, the pseudo S-ratios will be used to estimate the magnetic mineralogy, with values close to 1 indicating lower coercivity and a ferrimagnetic mineralogy (e.g., magnetite) and lower values indicating a higher coercivity, possibly an antiferromagnetic (e.g., hematite) mineralogy.
k_{ARM}/k: Indicates changes in magnetic grain size, if the magnetic mineralogy is dominantly magnetite.	If the magnetic mineralogy is dominantly magnetite k_{ARM}/k varies inversely with magnetic grain size, particularly in the 1–10 μm grain-size range. However, the interpretation of this ratio may be complicated by significant amounts of SP or paramagnetic material.
SIRM/k: Indicates changes in magnetic grain size, if the magnetic mineralogy is dominantly magnetite.	If the magnetic mineralogy is dominantly magnetite, SIRM/k varies inversely with magnetic particle size. SIRM/k

Table 2. (*Continued*)

Parameter	Interpretation
	is more sensitive than k_{ARM}/k to changes in the proportion of large (>10 μm) grains. SIRM/k may also be compromised by SP or paramagnetic material.
SIRM/k_{ARM}: Indicates changes in magnetic grain size, if the magnetic mineralogy is dominantly magnetite.	SIRM/k_{ARM} increases with increasing magnetic grain size, but is less sensitive and can be more difficult to interpret than the two ratios above. A major advantage of SIRM/k_{ARM} is that it only responds to remanence carrying magnetic material and is therefore not affected by SP or paramagnetic material.
Frequency dependent magnetic susceptibility (k_f): The ratio of low-frequency k (0.47kHz) to high-frequency k_{hf} (4.7 kHz) calculated by $k_f = 100^*(k–k_{hf})/k$	k_f is used to indicate the presence of SP material. SP material in high concentrations can compromise the grain-size interpretation made using k_{ARM}/k and SIRM/k.
Coercivity ratios: Ratios such as ARM_{20mT}/ARM_{0mT} or NRM_{30mT}/NRM_{0mT}.	These ratios provide information on the mean coercivity state of the sample, which is a reflection of its grain size and mineralogy. For example, for a magnetite-dominated mineralogy, ARM_{20mT}/ARM_{0mT} mostly reflects changes in magnetic grain size.
Median destructive field (MDF): Determined from the AF demagnetization procedure. It is the AF value needed to reduce the initial remanence by one half.	Similarly, the MDF provides information on the mean coercivity state of the sample, which is a reflection of its grain size and mineralogy. Higher MDFs indicate higher coercivity mineralogy. For a uniform mineralogy, finer grains require higher MDFs and coarser grains lower MDFs.
Hysteresis measurements[1]	
Saturation magnetization (M_s): M_s is the magnetization within a saturating field. Saturation remanence (M_{rs}): M_{rs} is the remanence remaining after removal of the saturating field.	M_{rs}/M_s decreases with increasing magnetite grain size in the submicron to few tens of microns grain-size range.

Table 2. (*Continued*)

Parameter	Interpretation
Coercive force (H_c): The back-field required to rotate saturation magnetization to zero within an applied field.	For magnetite, the ratio H_{cr}/H_c increases with increasing grain size (in the submicron to several hundred microns grain-size range) due to the strong grain-size dependence of both parameters, particularly H_c. H_{cr} is a useful guide to magnetic mineralogy.
Coercivity of remanence (H_{cr}): The back-field required to rotate saturation magnetization to zero remanence.	

Source: Modified with permission from Stoner, Channell, and Hillaire-Marcel (1996).
[1]Hysteresis parameters provide a means of monitoring grain-size variations in magnetite. Sediments should be homogeneous because of the small sample size (<0.05 g) typically used for hysteresis measurements. Mixed magnetic mineral assemblages greatly complicate the interpretations. The generalized interpretations listed below are based on a ferrimagnetic (e.g. magnetite) mineral assemblage.

the natural remanent magnetization (NRM) be stepwise demagnetized. This allows the component magnetization and its quality to be assessed. For u-channel samples, this is studied by progressive AF demagnetization. Because the u-channel samples are contained in plastic, thermal demagnetization techniques cannot be used. Though this limits some avenues of study, AF demagnetization does not physically alter the sediments. The sediments thus remain undisturbed for further investigation after the magnetic measurements have been completed. Typically, AF demagnetization is done using 5, 10, or even 20 mT steps between 0 and 80 mT and up to 140 mT. Discrete samples in some systems can be demagnetized at even higher levels. The magnetic mineralogy dictates the number of AF demagnetization steps. It is therefore wise to perform a pilot study prior to picking the best routine for a particular site. Generally, the routine used represents a balance between speed of data acquisition (each demagnetization and measurement step takes ~15 min) and the desire to precisely define the component magnetization of the remanence carrying grains.

5.2. Resolution: The Response Function and Deconvolution

As mentioned above, magnetic measurements on u-channel samples can be made at 1 cm intervals. However, the ~4.5 cm width (at half height) (Weeks et al., 1993) of the response function of the magnetometer pick-up coils is such that adjacent measurements at 1 cm spacing are not independent, and therefore the data are smoothed. This smoothing results in edge and other effects for a u-channel sample. Deconvolution protocols can be employed to reduce the smoothing and edge effects introduced by the response function of the magnetometer. Guyodo et al. (2002) successfully adapted the Oda and Shibuya (1996) deconvolution scheme to u-channel data from ODP sediments. Their study suggested that deconvolved

resolution is comparable to that derived from 1 cm discrete samples (Figure 4). Further work supports this interpretation (Channell, 2006) and is potentially a major breakthrough for paleomagnetic research. Moreover, recent work on understanding magnetometer sensitivity as a function of position within the sensing region (Parker, 2000; Parker & Gee, 2002) may enable magnetometer calibration and future implementation of even more sophisticated corrections and deconvolution techniques.

6. Data Analysis

The amount of data generated with a u-channel cryogenic magnetometer can be substantial. Sediment cores tens or even hundreds of meters in length are

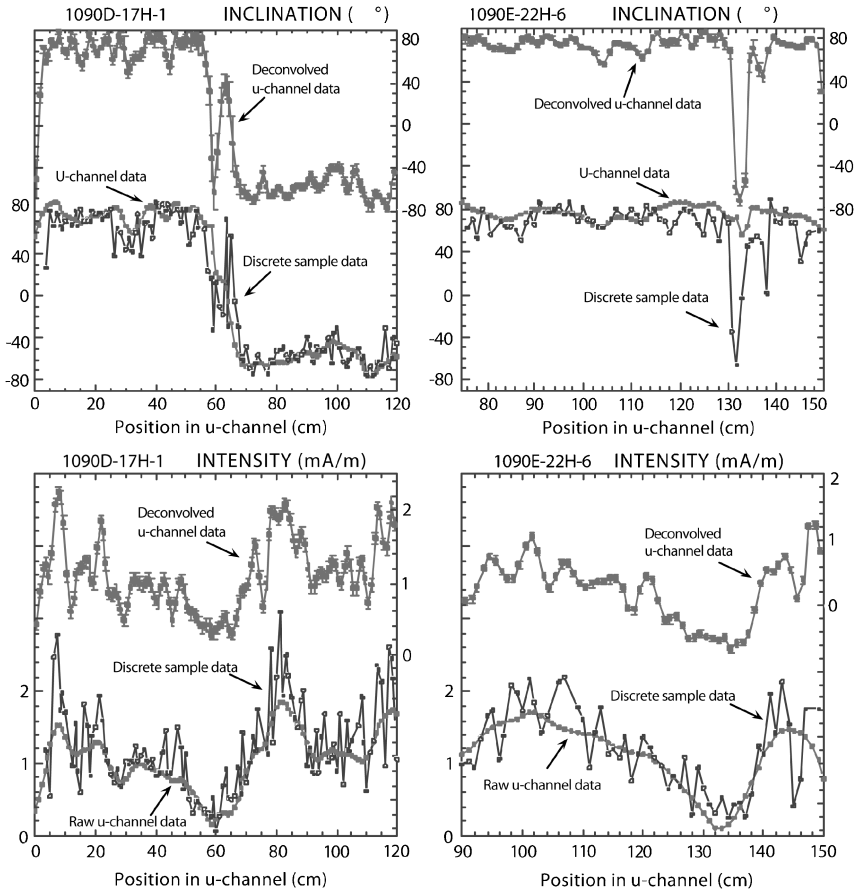


Figure 4 Comparison of component inclination (upper two graphs) and intensity (lower two graphs) from deconvolved u-channel data, raw u-channel data, and 1 cm discrete samples across a transition for low resolution (2 cm/kyr) South Atlantic ODP Site 1090. With permission from Guyodo et al. (2002).

now routinely measured with u-channel systems. Data are often collected at 1 cm intervals with each cm passing through the magnetometer 15 to more than 50 times. Careful evaluation and data reduction procedures are therefore required. In-house or commercially available programs transform the raw x, y, z magnetic moments into inclination and declination in degrees and intensity in A/m. Additional routines are often required to organize the data into the desired format for further analysis. Recently developed simple Excel spreadsheet (Mazaud, 2005) or freeware (e.g., Tauxe 1998; Jones, 2002; Cogné, 2003) can now be used to calculate and/or visualize the magnetic data in order to assess its downcore behavior. Such visualizations include the widely used Zijderveld plot (Zijderveld, 1967) for orthogonal projections of demagnetization data, normalized intensity diagrams, and stereographic projections (Figure 5). Many homegrown programs are also available and the Tauxe (1998) free software package is of particular interest for statistical analyses. The Excel spreadsheet developed by Mazaud (2005) was specifically developed for pass-through magnetometer data and is therefore particularly pertinent for sediment work. This macro allows easy calculation of component magnetizations and maximum angular deviation (MAD) values using the standard principal component analysis (PCA) of Kirschvink (1980). Additionally, the median destructive field at each interval is also provided. In this spreadsheet, the maximum number of demagnetization steps is 20, whereas the maximum number of lines is the Excel limit of 65,536 lines.

6.1. Orthogonal Projections and MAD Values

Orthogonal projections, often referred to as Zijderveld diagrams (Zijderveld, 1967), are the most commonly used approach to analyze changes in intensity and direction during demagnetization (Figure 5). The vector magnetization end points at successive demagnetization steps are plotted on both the horizontal and vertical planes. Straight-line segments indicate that the magnetic vector removed has a constant direction and the characteristic remanence is likely directed towards the origin. The component magnetization is calculated by PCA (Kirschvink, 1980), which provides a best-fit line to the demagnetization data. The MAD value provides a quantitative measurement of the precision with which the best-fit line is determined. MAD values $\geqslant 15^\circ$ are often considered ill defined and of questionable significance (Butler, 1992; Opdyke & Channell, 1996). In the context of relative paleointensity (RPI) and paleomagnetic secular variation (PSV) of Quaternary marine sediments, much better defined components should be expected with MAD values $\geqslant 5$ to be considered suspect during times of stable polarity. High MAD values generally reflect a complex magnetization where different coercivities record a magnetization lock-in (see below) at different times and, therefore, with different characteristics. However, high MAD values are also often associated with reversals or excursions, resulting from a rapidly changing geomagnetic field relative to the time interval over which the magnetization is completely locked-in.

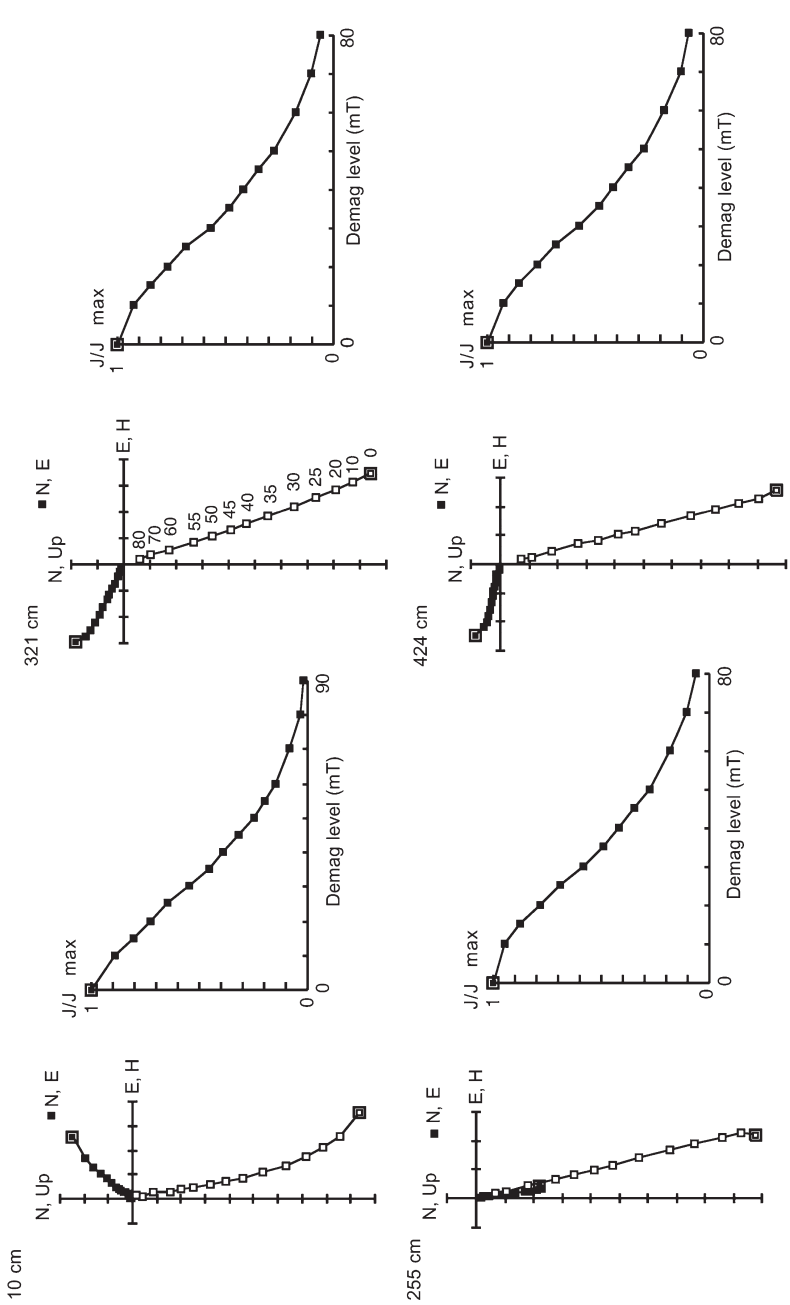


Figure 5 Typical vector endpoint diagrams (or Zijderveld plots; Zijderveld, 1967) and decay of the normalized intensity during AF demagnetization for pilot samples of core 2004–804–803, sampled in the Beaufort Sea (70°37.976′N/135°52.815′W). Closed squares represent north and east components, whereas open squares represent vertical and east components. As an example, the different steps, in mT, are illustrated in the sample at 321 cm.

7. Sediment Magnetism

The fact that sediments can record geomagnetic field variations has been known for more than 50 yr (Johnson et al., 1948). In the simplest case, the natural remanent magnetization (NRM) of sediment is aligned with the (geo)magnetic field and is a function of its intensity and direction at the time of deposition. In practice, many factors may work to convolve the geomagnetic input signal. These include the concentration, composition, and size of the remanence-carrying magnetic grains. Non-magnetic factors also have a substantial effect including sediment composition and grain sizes, rates of deposition, and bioturbation. Preservation is affected by both biogenic and diagenetic processes, which can add, subtract, or sometimes even completely erase the NRM. In addition, magnetization can be altered during sediment retrieval, as the sample must remain both physically and magnetically undisturbed. Nevertheless, under favorable circumstances, some of these effects can be separated and others avoided so that an accurate paleomagnetic record can be recovered.

7.1. The NRM Recording Process

Our understanding of the processes involved in the magnetization of sediments remains incompletely understood even after many years of study. Apart from the redeposition experiment of Katari, Tauxe, and King (2000), which argues that the mechanical reorientation of magnetic grains after deposition is highly unlikely, it is generally assumed that sediments acquire their magnetization through a mechanism termed post-depositional remanent magnetization (PDRM). In this mechanism, the magnetization is locked-in at a specific depth below the sediment/water interface (the lock-in depth), where dewatering and compaction constrain the remobilization of the magnetic grains below that depth (e.g., Verosub, Ensley, & Ulrick, 1979). The magnetic grains thus record the intensity and direction of the ambient field at the moment they pass-through the lock-in depth or zone. This lock-in depth is dependent on the thickness of the mixed layer, where no lock-in occurs because of the mixing produced by bioturbation. Boudreau (1994, 1998) estimated a worldwide and environmentally invariant (i.e., independent of water depth or sedimentation rate) mean thickness of the mixed layer to be 9.8 ± 4.5 cm, but with variable minimum and maximum values, respectively 2 and 30 cm. On the other hand, the work of Trauth, Sarnthein, and Arnold (1997) or Smith and Rabouille (2002) suggest that particulate organic carbon fluxes to the seafloor is the primary factor controlling the depth of the mixed layer, with higher fluxes resulting in thicker mixed layers. Below the mixed layer, the magnetic grains are believed to follow a lock-in function of uncertain nature. The paleomagnetic signal measured from sediment cores thus results from the integration of geomagnetic field fluctuations, the depth of the mixed layer, and a lock-in process function. This function was previously modeled as linear (Bleil & von Dobeneck, 1999; Roberts & Winklhofer, 2004), exponential (Lovlie, 1976; Hamano, 1980; Otofuji & Sasajima, 1981; Denham & Chave, 1982; Kent & Schneider, 1995; Meynadier & Valet, 1996; Guyodo & Channell, 2002; Roberts & Winklhofer, 2004), cubic

(Roberts & Winklhofer, 2004), or sigmoidal (Channell & Guyodo, 2004) and introduces smoothing of the paleomagnetic record in the form of a low-pass filter (Teanby & Gubbins, 2000; Guyodo & Channell, 2002; Channell & Guyodo, 2004; Roberts & Winklhofer, 2004). In the recent study of Channell and Guyodo (2004), the authors showed that the magnetization was acquired in only a few centimeters when the magnetic grains passed below the mixed layer, highlighting the influence of the mixed layer thickness rather than the lock-in function itself to explain the apparent age offset between reversals in the Matuyama Chron. Moreover, according to modeling results (e.g., Vlag, Thouveny, & Rochette, 1997; Roberts & Winklhofer, 2004), the smoothing introduced by the lock-in process could remove excursions or other high frequency variations if the sedimentation rate is in the order of only a few cm/kyr. Finally, the recent study of Sagnotti, Budillon, Dinarès-Turell, Iorio, and Macrì (2005) suggests that the lock-in depth may vary through time, likely as the result of lithological factors (Bleil & von Dobeneck, 1999). Clearly, additional work is needed to fully understand the NRM recording process and its effect on the paleomagnetic record.

7.2. Magnetic Mineralogy

Among the more important factors controlling the quality of the paleomagnetic record is the magnetic mineralogy, its concentration, and preservation (e.g., Clement, Kent, & Opdyke, 1996). Magnetic minerals on the sea floor are generally derived from terrigenous or *in-situ* bacterial production, and biogenic magnetite is generally thought to be a poor paleomagnetic recorder (e.g., Schwartz, Lund, Hammond, Schwartz, & Wong, 1997). Sediments with high terrigenous content may therefore provide high quality paleomagnetic records and glaciated or previously glaciated terrains generally provide an excellent source of fine-grained magnetic minerals for paleomagnetic records. In contrast, highly organic sites may be poor targets for paleomagnetic records, though there are many complicating factors controlling the stability of magnetic minerals (e.g., Channell & Stoner, 2002; Sagnotti et al., 2003). In general, magnetite and titano-magnetite are the primary remanence carriers of high quality paleomagnetic records, especially if they have a fine and homogeneous grain size. Preservation is generally controlled by the degree of reductive diagenesis, with dissolution of magnetic oxides potentially the most important factor controlling the fidelity of a paleomagnetic record (Clement et al., 1996).

8. Development of Paleomagnetic Records

8.1. Directional Records

Inclinations and declinations can be read directly off the magnetometer or after some minor transformation of the x, y, and z magnetic moments. Yet, establishing reliability is often another story. For magnetostratigraphic data primarily applied to polarity stratigraphy, Opdyke and Channell (1996) established a set of reliability

criteria following those advocated for paleomagnetic studies by Van der Voo (1990). Here, we discuss those criteria that most directly apply to marine sediment records, while emphasizing and expanding on those we believe to be the most important for Quaternary paleoceanographic studies.

For directional data, each discrete sample or individual measurement of a u-channel sample should undergo stepwise AF (thermal if appropriate) demagnetization. Orthogonal projections should be used for PCA analysis (e.g., Kirschvink, 1980) to calculate the component magnetization and the coercivity (unblocking) spectra should be determined. Blanket AF demagnetization, still relatively common for Holocene secular variation and shipboard magnetostratigraphic studies, should be used primarily for pilot studies and discouraged as definitive studies, unless no other reasonable options are available. MAD values should be used to assess the quality of the magnetization. Initially, these should be determined in a non-subjective manner, using a consistent demagnetization range and including all demagnetization steps (unless affected by a measurement error) for a particular sediment sequence. Selective use of demagnetization steps to assess the "best" component magnetization should only be done as a second step. Where possible, component inclination, declination, intensity at all demagnetization levels, and MAD values should be presented versus depth. For polarity stratigraphy, virtual geomagnetic pole (VGP) latitude should also be shown. Appropriate statistics should be shown when applicable. Evidence of magnetic mineralogy should be presented. Reversals should be antipodal. Multiple sections or replicate cores, where possible, should be studied. For high-resolution Holocene studies, attempts should be made to capture the upper sediments so that comparison and calibration with historical data (e.g., Jackson, Jonkers, & Walker, 2000) can be attempted.

8.2. Relative Paleointensity Determinations

Deriving proxy records of relative geomagnetic paleointensity from sediments involves normalizing the NRM by a magnetic parameter that compensates for change in the concentration of NRM-carrying grains. King, Banerjee, and Marvin (1983) put the use of ARM as a normalizer on a firm theoretical and empirical basis. According to these authors, the NRM/ARM ratio can be used as a paleointensity proxy if the NRM is a detrital remanent magnetization (DRM) and is carried by magnetite in the 1–15 μm single domain/pseudo-single domain (SD/PSD) grain-size range. The restricted grain-size range is due to the grain-size dependence of both ARM acquisition and NRM retention. Sub-micron grains and coarse multidomain (MD) magnetite grains can be efficient carriers of ARM but relatively inefficient carriers of stable DRM. According to the same authors, concentrations of magnetite should not vary downcore by more than a factor of 20–30 because of the effect of particle interactions on the efficiency of ARM acquisition. Demagnetization of both NRM and ARM in the NRM/ARM ratio serves to restrict the grain-size range contributing to both remanences.

Besides ARM, IRM and low field susceptibility (k) have been used as normalizers to derive paleointensity proxies from sediments, although the theoretical basis for the use of IRM or susceptibility (k) has not been adequately documented.

Empirically, IRM has often proved to be a better match to the NRM's coercivity than ARM and therefore a good choice as a normalizer (e.g., Stoner, Channell, Hillaire-Marcel, & Kissel, 2000). The RPI proxies can be calculated as means of NRM/IRM or NRM/ARM over a specific demagnetization range, as slopes on the plots of NRM versus ARM or IRM, or as subtracted vectors (e.g., $NRM_{25mT-50mT}/IRM_{25mT-50mT}$ or $NRM_{25mT-50mT}/ARM_{25mT-50mT}$). More complex methods have been advocated, like the pseudo-Thellier method of Tauxe, Pick, and Kok (1995). However, little difference has been found among the various methods when the sediments are high quality magnetic recorders (Valet & Meynadier, 1998), which can be assessed through studying the demagnetization behavior and is generally indicated by low MAD values. In practice, the choice of the normalizer depends on the response of the sediments to the acquisition and demagnetization of NRM, ARM, and IRM. The success of these techniques is illustrated by the proliferation of RPI studies over the last decade. This progress has been greatly accelerated by the development of the u-channel magnetometer (Weeks et al., 1993), which allows high-resolution studies as a practical undertaking and the extensive data acquisition required for RPI studies.

Determining whether normalized intensity actually reflects RPI changes is not always straightforward. Separation of geomagnetic and environmental factors may be difficult. Therefore, magnetic homogeneity has been considered a requirement for the development of trustworthy RPI records. Careful evaluation of the sediments, both in terms of their magnetic properties and the integrity of the recorded magnetization, are required. Criteria have been established that outline the types of sediments likely to provide reliable RPI records (e.g., Levi & Banerjee, 1976; King et al., 1983; Tauxe, 1993; Valet, 2003). However, it should be remembered, as with the NRM recording process, that our understanding is incomplete and that following these criteria will not guarantee a quality record, nor does it mean that only sediments that strictly follow these criteria can provide RPI estimates.

Following on the directional criteria listed above and those previously established for RPI studies (King et al., 1983; Tauxe, 1993; Valet, 2003), we will expand and add some recommendations for the construction of reliable RPI estimates. The basic idea is that the sediments should be good geomagnetic field recorders and should be geologically homogenous. Specific criteria that should be applied are: (1) the NRM of all samples should be studied by progressive AF demagnetization, permitting the removal of any viscous (low coercivity and unstable) magnetization and the determination of the component magnetization as well as the coercivity spectrum of the NRM. The stability of the magnetization and its characteristic remanence should be determined by orthogonal projections (Zijderveld plots; Figure 5) and PCA analysis. MAD values of less than 5 are preferable. Component magnetization with higher MAD values should be treated as suspect. (2) The magnetization should be carried by stable magnetite of PSD range. The mineralogy and grain size of the magnetic faction should be established (see Tauxe, 1993; Dunlop & Özdemir, 1997). (3) The sediments must be free of inclination error. This can notably be assessed by comparing the inclination record with the expected inclination for the latitude of the sampling site according to a GAD ($\tan I = 2 \tan \lambda$).

(4) Changes in magnetic concentration should vary by less than one order of magnitude, with rapid or abrupt changes treated with extra caution. These can be evaluated by looking at concentration dependent parameters such as NRM, ARM, IRM, and *k*. (5) The normalizer should activate the same magnetic assemblage that is responsible for the NRM acquisition. This can be assessed by demagnetization of the normalizer using the same steps as the NRM. Similar values at successive ratios provide a test of the coercivity match. The use of magnetic susceptibility as a normalizer should be carefully evaluated as it activates large MD magnetite grains and small (superparamagnetic) grains, both of which are not stable carriers of the NRM. (6) The RPI proxy should not be coherent with its normalizer and the RPI proxy should not be coherent with bulk rock magnetic parameters. This is often determined by cross-spectral analysis (Tauxe & Wu, 1990), or more recently by cross-wavelet analysis (Guyodo, Gaillot, & Channell, 2000). (7) Comparison between individual RPI records with other regional records provides replication of observations. Additional comparisons with regional stacks such as the North Atlantic paleointensity stack (NAPIS) (Laj, Kissel, Mazaud, Channell, & Beer, 2000), the South Atlantic paleointensity stack (SAPIS) (Stoner, Laj, Channell, & Kissel, 2002), or global stacks such as the high resolution global paleointensity stack (GLOPIS-75) (Laj, Kissel, & Beer, 2004), and the low resolution Sint-200 (Guyodo & Valet, 1996), Sint-800 (Guyodo & Valet, 1999), or Sint-2000 (Valet, Meynadier, & Guyodo, 2005) paleointensity stacks. Additionally, comparison with inverted cosmogenic isotope records can help assess the validity of the reconstructed paleointensity proxy. Finally, non-magnetic changes can also substantially affect the ability to derive a reliable RPI record. Recent improvements in non-destructive physical methods (e.g., St-Onge et al., this book) allow a rapid and precise visualisation of the cores or u-channels in order to evaluate the potential influence of cracks, coring deformation, turbidities, sand layers, or other disturbances of the paleomagnetic signal.

9. The Paleomagnetic Record as a Stratigraphic Tool

9.1. Geomagnetic Polarity Time Scale (GPTS)

The GPTS (Cande & Kent, 1995; Ogg & Smith, 2004) and magnetic polarity stratigraphy represent the stratigraphic backbone on which the geologic timescale for the last 150 Myr is based. For a thorough review, see Opdyke and Channell (1996). Magnetic reversals are applicable to many types of geologic materials and their global synchronicity and environmental independence make them a supreme stratigraphic tool. Magnetic reversals are a non-periodic, possibly even stochastic process, resulting from an incompletely understood mechanism within the Earth's core. Reversals separate times of constant polarity with durations from 20 kyr to 50 Myr. Considering that reversals take about 5,000 yr on average (see Clement, 2004), they provide among the most precise methods of global correlation. Patterns of normal and reversed strata can be classified as polarity zones that can provide, in some circumstances, a distinct

fingerprint match to the GPTS. These magnetostratigraphic polarity zones can consist of strata with a single polarity, alternating normal and reversed units or dominantly normal or reversed units with minor amounts of the opposite polarity. Correlation to the GPTS is only provided at the recorded reversal boundaries and, therefore, temporal estimates must be interpolated between reversal boundaries. In addition, reversals are non-unique, coming in either one of two forms, normal to reversed or reversed to normal. Cross-calibration is thus extremely important to uniquely identify a reversal, most often occurring with isotopic (e.g., Shackleton & Opdyke, 1973; Channell & Kleiven, 2000; Channell, Mazaud, Sullivan, Turner, & Raymo, 2002; Channell, Curtis, & Flower, 2004) or biostratigraphic data (e.g., Hays & Opdyke, 1967; Berggren, Kent, Swisher, & Aubry, 1995).

The naming of the first four magnetic polarity Chrons after prominent geomagnetists, Brunhes (normal), Matuyama (reversed), Gauss (normal), and Gilbert (reversed), spanning the past ~6 Myr, was put forward by Cox et al. (1963). These Chrons were identified based on dispersed radiometrically dated volcanic rocks, so no type localities were available. Shorter Subchrons were named after type localities, such as Jaramillo Creek in New Mexico or Olduvai Gorge in Tanzania. Beyond the Plio-Pliestocene, polarity Chrons are designated by numbers correlated to marine magnetic anomalies. The anomaly sequence of the South Atlantic was taken as a marine standard for the Late Cretaceous through the Cenozoic. The anomalies of the Cenozoic or "C-sequence" were numbered from 1 to 34 (oldest). The polarity Chron nomenclature has evolved progressively to accommodate revisions (LaBrecque, Kent, & Cande, 1977; Harland et al., 1982; Cande & Kent, 1992). The corresponding polarity Chrons (time) and polarity zones (stratigraphy) are prefaced by the letter C, with a suffix "*n*" denoting the younger normal polarity interval, or "*r*" denoting the older reversed polarity interval (e.g., Cande & Kent, 1992). When a major numbered polarity Chron is further subdivided, Subchrons are denoted by a suffix of a corresponding number polarity Chron. For the Plio-Pleistocene, the traditional names are used (Brunhes = C1*n*, Matuyama = C1*r*, Jaramillo Subchron = C1*r*.1*n*).

Starting with the pioneer work of Heirtzler, Dickson, Herron, Pitman, and Pichon (1968), the timescale for the GPTS has been derived based on assumptions that seafloor spreading at specific locations was constant or smoothly varying over long time intervals. Age calibration was achieved by fitting a smooth curve to nine-calibration levels for South Atlantic spreading history (Cande & Kent, 1992). The rest of the timescale is dated by interpolation between these tie-levels. The absolute ages of the tie-levels are much less exactly known than the relative length of the polarity intervals. The timescale has evolved with improvements in the resolution of magnetic anomalies, definition of oceanic block models, magnetostratigraphic correlation, and dating of calibration points. Recent updates, like that of Berggren et al. (1995) which was used by Cande and Kent (1995) and Ogg and Smith (2004), have incorporated improved ages for the calibration levels with many of these based on astronomical tuning (e.g., Shackleton, Berger, & Peltier, 1990; Hilgen, 1991; Lourens, Hilgen, Shackleton, Laskar, & Wilson, 2004).

9.2. Relative Paleointensity Stratigraphy

One of the most important developments in paleomagnetism over the last decade has been the demonstration that sediment normalized intensity records show globally coherent variations during times of constant polarity (e.g., Meynadier, Valet, Weeks, Shackleton, & Hegee, 1992; Tric et al., 1992; Stoner, Channell, & Hillaire-Marcel, 1995; Stoner et al., 2000; Guyodo & Valet, 1996; Guyodo & Valet, 1999; Laj et al., 2004; Valet et al., 2005; Yamazaki & Oda, 2005). The importance of these observations is that they suggest that the geomagnetic field can provide a stratigraphic tool between magnetic reversals, overcoming a major limitation of the GPTS, where correlation is only possible at reversal boundaries. In fact, high-resolution studies suggest that RPI records maybe globally coherent on timescales as short as a few thousand years (e.g., Stoner et al., 2000; Laj et al., 2004).

The principal challenge in the development of RPI stratigraphy is to define the "true" character of the record. Paleointensity cannot be predicted by theory or from numerical simulation, since the mechanisms involved in the geodynamo are not sufficiently constrained. Thus, our understanding of the record is based on continual observations and cross correlation with other dating and stratigraphic techniques. Comparison between RPI records from sediments and absolute paleointensity from thermally cooled materials (e.g., volcanic rocks and ceramic artifacts) can be used to calibrate the sediment record (e.g., Guyodo & Valet, 1999; Valet, 2003; Laj et al., 2004). In practice, however, the volcanic/sedimentary correlation is often hampered by the discontinuous nature of the volcanic record and the imprecision of available radiometric dating techniques. Comparison of sedimentary paleointensity records from different depositional environments and a detailed investigation of magnetic properties allow separation of geomagnetic and environmental signals. Distributed records from different parts of the globe are necessary to determine the characteristics of the global (as opposed to the local or regional) geomagnetic field. Stacking many individual records provides one method for determining the "true" character of the signal. Spurious features in individual records are averaged out by the stacking process. However, stacking also acts as a low pass filter, degrading the highest temporal resolution records in favor of those of lower resolution. Correlation imprecision tends to reinforce this. The use of low sedimentation rate records such as the global Sint-200 (Guyodo & Valet, 1996), Sint-800 (Guyodo & Valet, 1999), Sint-2000 (Valet et al., 2005), and regional EPAPIS-3000 (Yamazaki & Oda, 2005) paleointensity stacks have averaged out much of the high frequency variability. These records, however, do an excellent job at capturing the 10^4 yr variability of the geomagnetic field (Figure 6a), while providing excellent targets for stratigraphic correlation. Globally dispersed records from sediments that have accumulated at one order of magnitude higher (>10 cm/kyr) give a picture of the geomagnetic field with high amplitude features varying globally and generally coherently on a millennial scale (Stoner et al., 2000; Figure 7). The availability of an increased number of high resolution records and a more sophisticated stacking approach have recently been used to reduce the low pass filter effects (Laj et al., 2004) and provide GLOPIS-75 at a significantly higher resolution (Figure 6b).

One of the unique aspects of paleointensity is its relationship to cosmogenic isotopes. Assuming a constant flux of galactic cosmic rays, to a first order, the production rate of cosmogenic isotopes (e.g., ^{10}Be, ^{14}C, ^{36}Cl) reflects variations in the strength of the Earth's and Sun's magnetic fields, with stronger (weaker)

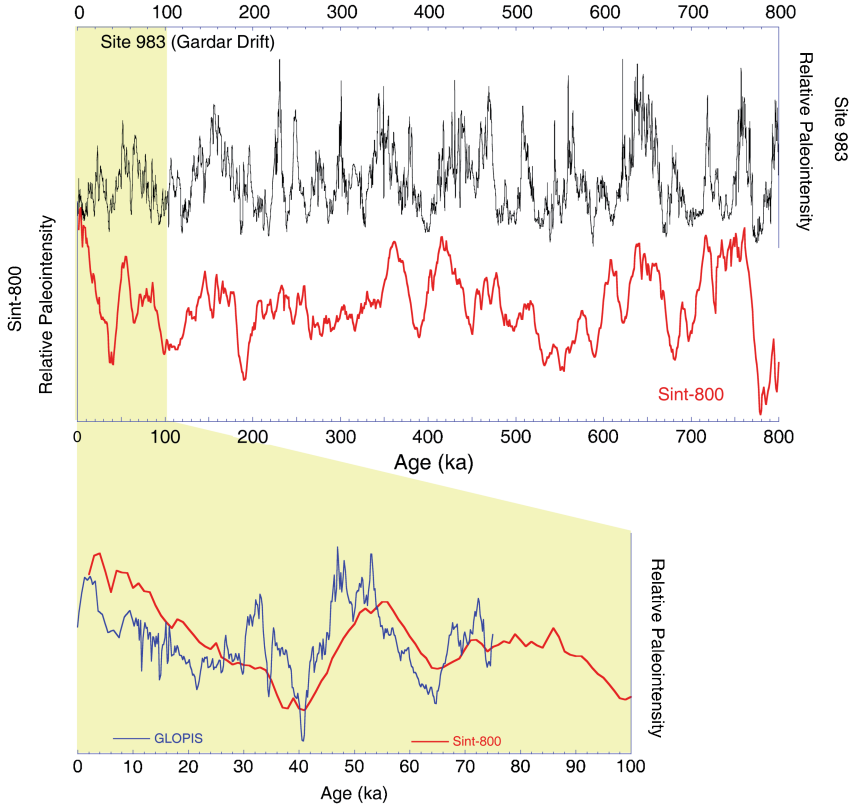
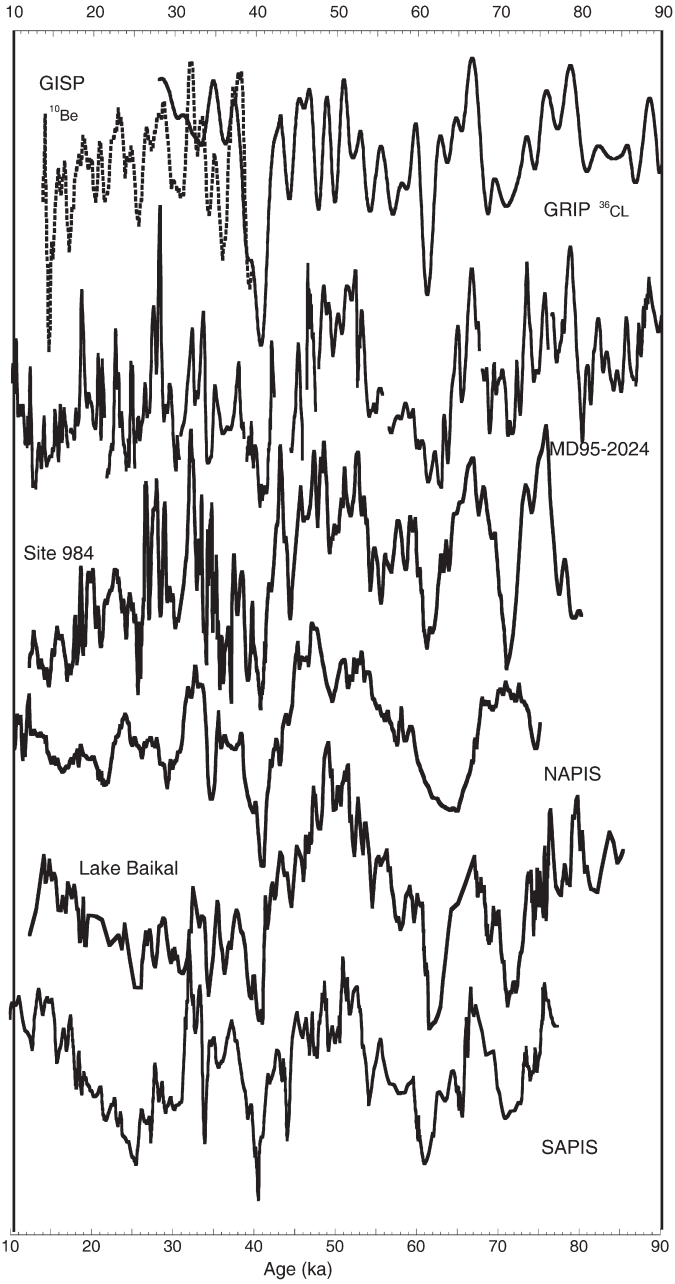


Figure 6 (A) Comparison of North Atlantic ODP site 983 high-resolution relative paleointensity record (Channell et al., 1997; Channell, Hodell, McManus, & Lehman, 1998; Channell & Kleiven, 2000) with the low-resolution SINT-800 stack (Guyodo & Valet, 1999). (B) Comparison between the high-resolution global relative paleointensity stack (GLOPIS; Laj et al., 2004) and the low-resolution SINT-800 stack. The Laschamp excursion is associated with the intensity low at ~40 ka.

Figure 7 Geomagnetic relative paleointensity records and cosmogenic isotope records. From the top to the bottom: the ^{10}Be flux from the GISP2 ice core (72°36′N/38°30′W; Finkel & Nishiizumi, 1997) and ^{36}Cl flux from the GRIP ice core (72°34′N/37°37′W; Baumgartner et al., 1998) placed on the GISP official chronology. Core MD95-2024 (50°12.26′N/45°41.14′W; Stoner et al., 2000) paleointensity record. North Atlantic ODP Site 984 (60.4°N/23.6°W; Channell 1999) paleointensity record, the North Atlantic Paleointensity Stack (NAPIS; 33–67°N/45°W–4°E; Laj et al., 2000), the Lake Baikal paleointensity record (Siberia, Russia; Peck, King, Colman, & Kravchinsky, 1996), and the South Atlantic Paleointensity Stack (SAPIS; 41–47°S/6–10°E; Stoner, Laj, Channell, & Kissel, 2002). Chronology for all records derived by tuning to the GISP2 official chronology. The offset in the NAPIS record results from an additional step of tuning to SPECMAP at ~60 kyr.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SAPIS
Age (ka)

magnetic fields resulting in reduced (increased) production. In general, geomagnetic and solar shielding of galactic cosmic rays is well known and reasonably well understood (e.g., Lal, 1988; Masarik & Beer, 1999; Beer, 2000). What is not so well understood are the timescales of the two components. Recent comparisons between proxies of cosmogenic isotope production rates and RPI (Mazaud, Laj, & Bender, 1994; Robinson, Raisbeck, Yiou, Lehman, & Laj, 1995; Baumgartner et al., 1998; Stoner et al., 2000; Wagner et al., 2000; Beer et al., 2002; Carcaillet, Thouveny, & Bourles, 2003; Christl, Strobl, & Mangini, 2003; Hughen et al., 2004; Thouveny, Carcaillet, Moreno, Leduc, & Nérini, 2004; Muscheler, Beer, Kubik, & Synal, 2005) show substantial agreement. Beer et al. (2002) suggests that variations longer than 2,000 yr in proxy records of cosmogenic isotope production rates can be attributed to variations in the intensity of the geomagnetic field, with higher frequency variations attributed to solar variability. Therefore, RPI provides a viable technique for correlating sediment to cosmogenic isotope records from ice cores (Mazaud et al., 1994; Stoner et al., 2000) and sediment archives (Frank et al., 1997; Hughen et al., 2004) (Figure 7). New observations from RPI records from the St. Lawrence Estuary (St-Onge, Stoner, & Hillaire-Marcel, 2003) and Scandinavia (Snowball & Sandgren, 2002, 2004) suggest that millennial and even centennial scale correlations between cosmogenic isotopes and paleointensity could be used at these timescales for stratigraphic purposes.

9.3. Excursions as a Stratigraphic Tool

Barbetti and McElhinny (1972) defined an excursion as a VGP displacement of more than 40° from the geographic pole, whereas a reversal excursion (Merrill & McFadden, 1994) reflects a VGP in the opposite hemisphere, the latter often being a more useful paleomagnetic definition. A decade ago, the existence of geomagnetic excursions within the Brunhes Chron was treated as suspect. Only two "cryptochrons" are recognized by Cande and Kent (1992, 1995) from the Brunhes and Matuyama Chrons marine magnetic anomaly data, with estimated ages of 500 ka and 1.2 Ma (Cobb Mt. Subchron). Recent observations from sediment records suggest that they may be much more common (Langereis, Dekkers, de lange, Paterne, & van Santvoort, 1997; Lund et al., 1998; Lund, Acton, Clement, Okada, & Williams, 2001a, 2001b), with 12–17 now advocated in the Brunhes and at least another six in the Matuyama (e.g., Channell et al., 2002). The volcanic record also appears to bear this out (Singer et al., 2002), yet a strong stratigraphic understanding of most of these is presently missing. As a correlation tool, excursions have unique potential to provide short duration (<2,000 yr) "golden spikes" in the geologic record. However, only two excursions (Laschamp and Iceland Basin Events) are widely considered as globally synchronous phenomena (Channell, Stoner, Hodell, & Charles, 2000; Mazaud et al., 2002; Stoner, Channell, Hodell, & Charles, 2003), with many of the details still unknown. Many pitfalls therefore remain for their common and uncontrolled use as a stratigraphic tool without other significant stratigraphic calibration. Similarly to a reversal, the lack of any distinct fingerprint to uniquely identify an excursion is a significant stratigraphic drawback. This is

important as excursions may occur in bundles closely spaced in time (Lund, Stoner, Acton, & Channell, 2006; Blanchet, Thouveny, & de Garidel-Thoron, 2006). And, if they result, as is generally thought, from non-dipole components of the geomagnetic field, then their directional signature would be distinctly different for different parts of the globe (e.g., Merrill & McFadden, 2005). However, a recent study suggests that they may be much simpler (Laj, Kissel, & Roberts, 2006), with a potentially global signature.

A more fruitful stratigraphic approach should result by combining excursions with RPI. Valet and Meynadier (1993) pointed out that lows in RPI during the Brunhes Chron from equatorial Pacific (ODP Leg 138) sediments appear synchronized with directional excursions detected elsewhere. The sedimentation rates in Leg 138 sediments and those used to construct the global stack Sint-800 (Guyodo & Valet, 1999) are, in general, too low to record geomagnetic excursions. Direct correlation of excursions and paleointensity is thus incomplete and is only convincingly achieved for a few excursions at a few locations (e.g., Lehman et al., 1996; Channell, Hodell, & Lehman, 1997; Roberts, Lehman, Weeks, Verosub, & Laj, 1997; Channell, 1999; Laj et al., 2000; Channell et al., 2000; Stoner et al., 2003; Thouveny et al., 2004; Lund, Schwartz, Keigwin, & Johnson, 2005). Therefore, our expanding knowledge of the paleointensity record provides an improving template with which to search and a guide that should be used when employing excursions as a stratigraphic tool.

An added difficulty with excursions results from the fact that they occur during times of low field intensity. The weak magnetizing field increases the likelihood that the excursional interval will be overprinted from the stronger post-excursional field (e.g., Coe & Liddicoat, 1994). In addition, it has been suggested that these short-lived phenomena are easily smoothed out of the record by the NRM recording process in low accumulation rate sediments, as discussed above and reviewed by Roberts and Winklhofer (2004). The presence of detailed oxygen isotope records has also been instrumental in improving the age estimates for excursion records in deep-sea sediments, but it will likely take a considerable international and multidisciplinary effort before a complete record of excursions within the Brunhes is successfully mapped out.

As mentioned, one excursion that can be used as a stratigraphic golden spike is the Laschamp event. Bonhommet and Babkine (1967) originally recognized excursional directions in Quaternary lava flows from the Chaine des Puys, France. The Laschamp excursion has now been recognized in many locations and represents one of the most prominent events to have affected Earth over the last 100 kyr. The Laschamp excursion has been found in sediments of the North Atlantic, South Atlantic, Indian, and Pacific oceans. It is generally associated with a broad paleointensity low. The best age estimates come from correlation of sediments to the Greenland Summit ice cores that place the event's occurrence during interstadial (IS) 10 with a GISP2 age of $\sim$40.5 kyr. The type section lavas are independently dated at 40.4 ± 4 kyr (Guillou et al., 2004) and support the ice core derived dates. Further supporting this correlation are the observations of a major increase in the flux of cosmogenic isotopes associated with IS10 in Greenland Ice cores (e.g., Yiou et al., 1997; Wagner et al., 2000) and marine sediments

(e.g., Robinson et al., 1995; Carcaillet, Bourles, & Thouveny, 2004). This event has been recently used to support correlation of paleoclimatic records to the GISP2 chronology (Mazaud et al., 2002; Lamy et al., 2004; Hill, Flower, Quinn, Hollander, & Guilderson, 2006). The fact that this event is within the radiocarbon limit has improved its stratigraphy, though its incompletely understood effect on the production of ^{14}C significantly complicates radiocarbon dating during this time interval.

9.4. Paleomagnetic Secular Variation

Observations of the historical (~the last 400 yr) geomagnetic field document continuous directional secular variations with periods on a centennial scale (Jackson et al., 2000). Paleomagnetic observations primarily from Holocene lake sediments document additional ~1,000–3,000 yr long geomagnetic directional shifts termed paleomagnetic secular variation (PSV) (Mackereth, 1971; Creer, Thompson, Moyneaux, & Mackereth, 1972; Thompson, 1973) that are regionally consistent, but different between regions (Thompson, 1984). Recent studies show that marine sediments can record these variations (Figure 8), providing a potentially continuous regional centennial to millennial stratigraphic tool (Lund & Keigwin, 1994; Lund & Schwartz, 1996; Kotilainen, Saarinen, & Winterhalter, 2000; Verosub, Harris, & Karlin, 2001; St-Onge et al., 2003; St-Onge, Piper, Mulder, Hillaire-Marcel, & Stoner, 2004; Stoner et al., 2007). Because most of the initial work was done in lake sediments, marine/terrestrial correlation (e.g., Ólafsdóttir, Stoner, Geirsdóttir, Miller, & Channell, 2005) has significant potential that, in some cases, maybe better than what can be achieved through radiocarbon dating alone (Lund, 1996; Hagstrum & Champion, 2002; Stoner et al., 2007).

Most PSV studies have been concentrated in the Holocene, though its potential impact as a correlation tool is possible for older records. For example, recent studies from Blake/Bahamas Outer Ridge (ODP Sites 1061) and the Bermuda Rise Site 1063 sediments document reproducible directional PSV records, at more than 1,000 km distance, for the interval 15,000–50,000 yr BP (Lund et al., 2005).

10. Some Perspectives

As a community, we are just beginning to document geomagnetic change during times of constant polarity. Though there is still much to learn, progress is being made at a rapid rate. Much of this growing new understanding is fueled by the availability of new instruments (i.e., the u-channel magnetometer) that allow rapid measurements of sediment sequences, providing new observational constraints from previously untapped archives. Additionally, both numerical geodynamo and data based spherical harmonic models are also making substantial progress. Each new record is potentially an important new observation, providing new insights toward defining the "true" record, while the modeling work allows us to better interpret these observations. Continual observations and cross correlation with other dating and stratigraphic techniques allow the "true" record to be uncovered

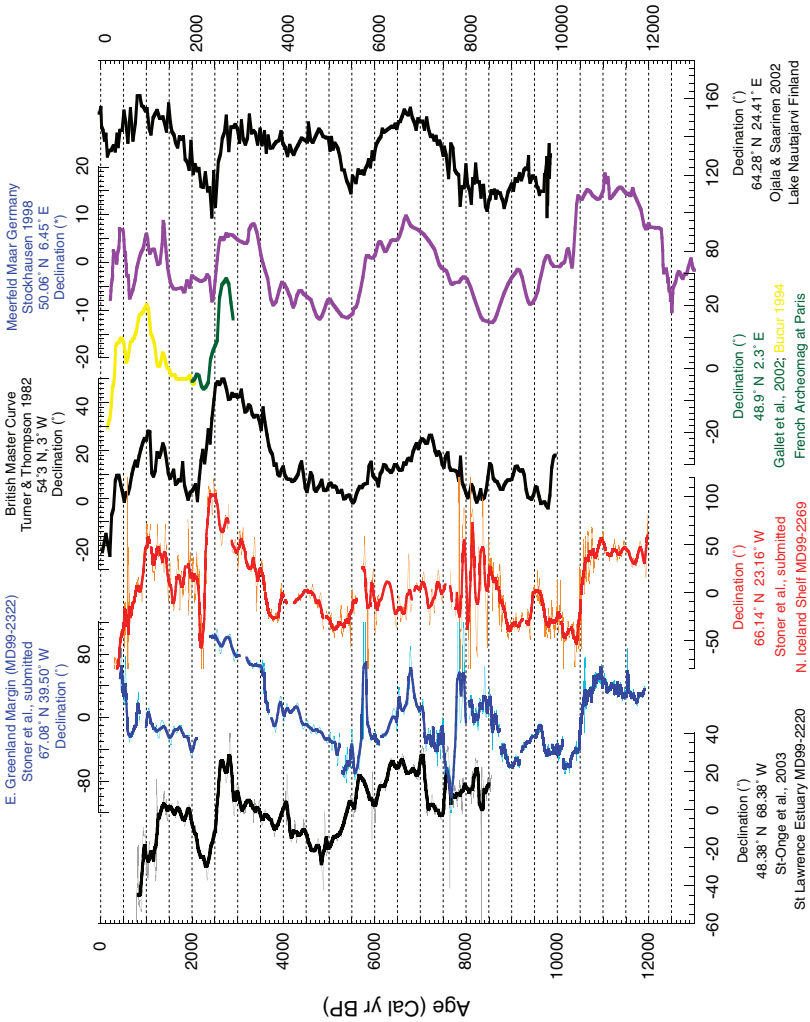

Figure 8 Comparison of Holocene paleomagnetic secular variation records from North America–North Atlantic–Europe. Only declination is shown. From left to right in each panel: St. Lawrence Estuary (MD99-2220; St-Onge et al., 2003); East Greenland Margin (MD99-2322; Stoner et al., 2007); North Iceland shelf (MD99-2269; Stoner et al., 2007); the western European archeomagnetic compilation (Bucur, 1994; Gallet, Genevey, & Le Goff, 2002); the British master curve (Thompson & Turner, 1979; Turner & Thompson, 1982); the German maar lake records (Stockhausen, 1998); and a Finish varved lake record (Ojala & Saarinen, 2002). All records are on their own chronologies and calibrated to calendar years using Stuiver et al. (1998).

piece by piece. The spatial coherence of the geomagnetic field allows prior information to be brought to each new observation and provides many stratigraphic opportunities. Yet, in many ways we are just scratching the surface. Global coverage over all timescales is still only fair, and as for high-resolution data detailed records are only found in a few restricted regions. Holocene observations suggest that if we are to understand the full dynamic range of the geomagnetic record, ultra high-resolution (>1 m/kyr) sediments, both within the Holocene and beyond, need to be targeted. As we seek to uncover the "true" geomagnetic record, our knowledge of geomagnetic stratigraphy, its applications, and its resolution will continue to improve, providing a rich future for paleomagnetic research.

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