SEDIMENTARY RECORDS OF RELATIVE PALEOINTENSITY OF THE GEOMAGNETIC FIELD: THEORY AND PRACTICE

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Abstract. Sediments have proved irresistible targets for attempts at determining the relative variations in the Earth's magnetic field because of the possibility of long and continuous sequences that are well dated and have a reasonable global distribution. The assumption underlying paleointensity studies using sedimentary sequences is that sediments retain a record reflecting the strength of the magnetic field when they were deposited. Early theoretical work suggested that because the time required for an assemblage of magnetic particles in water to come into equilibrium with the ambient magnetic field was quite short, no dependence on magnetic field was expected. Nonetheless, a number of experiments showed that sedimentary magnetizations varied in accordance with the field, albeit not always in a simple, linear fashion. Experiments in which the sediments were stirred in the presence of a field (to simulate bioturbation) showed a reasonably linear relationship with the applied field, and these results spurred the hope that variations in the Earth's magnetic field might indeed be recoverable from appropriate sedimentary sequences. Examination of existing paleointensity data sets allows a few general conclusions to be drawn. It appears that sedimentary sequences can and do provide a great deal of information about the variations in relative paleointensity of the Earth's magnetic field. The dynamic range of sedimentary data sets is comparable to those acquired from thermal remanences. Moreover, when compared directly with such independent measures of magnetic field variations as beryllium isotopic ratios and thermally blocked remanences, there is considerable agreement among the various records. When viewed over timescales of hundreds to thousands of years, relative paleointensity data sets from more than a few thousand kilometers apart bear little resemblance to one another, suggesting that they are dominated by nondipole field behavior. When viewed over timescales of a few tens of thousands to hundreds of thousands of years, however, the records show coherence over large distances (at least thousands of kilometers) and may reflect changes in the dipole field. On the basis of a sequence spanning the Brunhes and Matuyama chrons, the magnetic field has oscillated with a period of about 40 ka for the last few hundred thousand years, but these oscillations are not clear in the record prior to about 300 ka; thus they are probably not an inherent feature in the geomagnetic field, and the correspondence of the period of oscillation to that of obliquity is probably coincidence.

1. INTRODUCTION

One of the first motivations for investigating the magnetism of rocks was to study the behavior of Earth's magnetic field in the past. Paleogeomagnetism is almost unique among geophysical endeavors in that a historical perspective is possible, and the potential of the rock record was realized from very early on. The magnetic field is a vector field, having both direction and intensity, and a complete understanding of it requires study of the full vector properties. However, paleointensity determinations are much more difficult than directional ones, and the great majority of paleogeomagnetic studies are concerned solely with directional variability of the field. Nonetheless, intensity variations of the field are of great interest for a variety of reasons. For example, it has long been known that the Earth's magnetic field can partially deflect incoming cosmic rays [e.g., *Elsasser et al.*, 1956]. Cosmic rays are responsible for the creation of such radionuclides as ^{14}C and ^{10}Be, and the radiocarbon timescale is built upon the assumption that the rate of production is constant. Detailed calibration of the radiocarbon dates using tree rings and varved sediments of known age resulted in the discovery of rather large discrepancies [e.g., *Stuiver et al.*, 1986], and these deviations can be reasonably well modeled as resulting from changes in Earth's magnetic field (see Figure 1). This calibration is available only for the last 10 ka or so, and direct calibration prior to that time is based on a very few spot calibration points comparing radiocarbon ages with those derived from thorium decay [see *Bard*

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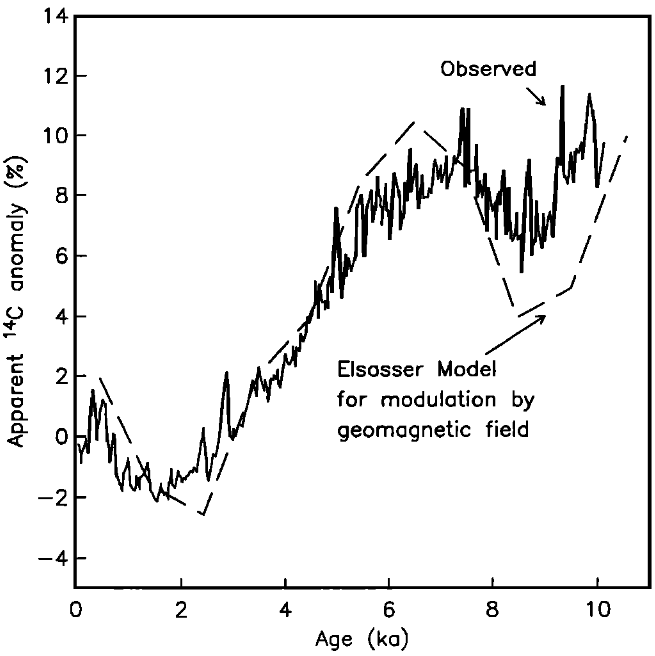

Figure 1. Comparison of observed variations in ^{14}C production by *Stuiver et al.* [1986] and those predicted by *Elsasser et al.* [1956] using the dipole moment data compiled by *McElhinny and Senanayake* [1982]. Figure redrawn from *Tauxe and Valet* [1989].

et al., 1990*a*, *b*; *Mazaud et al.*, 1991]. Better calibration of the radiocarbon timescale thus depends on an accurate knowledge of the intensity behavior of Earth's dipole field.

Direct observation of the geomagnetic field has been carried out for the last few centuries, and prior to that time, data must be culled from archeological and geological proxies. Figure 2 illustrates the present sources of paleointensity data. In principle, the best records for paleointensity studies are those in which the magnetization was acquired by cooling from high temperature in the presence of the magnetic field (volcanic rocks and fired archeological materials). Even these, however, can be controversial [see *Dunlop et al.*, 1987; *Aitken et al.*, 1988; *Walton*, 1988]. Problems inherent in such studies are the difficulty in reproducing the original cooling rate in the laboratory, alteration of the material during the experiment, nonideal rock magnetic properties (e.g., multidomain magnetizations), and so on. Furthermore, these so-called thermal remanences give only spot readings of the field, leading to a discontinuous record which is quite sparse prior to about 10 ka [see *Merrill and McElhinny*, 1983].

Far less straightforward for paleointensity is the use of sedimentary records. There are, of course, lake and marine sediments spanning virtually the whole of the Phanerozoic, but most published paleointensity records span at most the last few hundred thousand years. The advantage of sedimentary sequences, in particular marine sediments, is that they provide more or less continuous records that can be dated with relative precision. Good global coverage is also a possibility.

The difficulties in sedimentary paleointensity studies stem from the lack of a sound theoretical basis for the acquisition of remanence. The assumptions underlying sedimentary paleointensity are illustrated in Figure 3. As we shall see, under certain laboratory conditions, sediments acquire a remanence which is linearly related to the applied field. Thus the assumption is that the remanent magnetization carried by detrital grains, DRM, is a linear function of the applied field B under certain geological conditions as well. I will discuss the body of research relating the intensity of remanence in sediments to lithologic factors, such as mineralogy, concentration, and grain size of the magnetic phases and properties of the nonmagnetic matrix. This dependence on factors other than the field implies that sedimentary magnetizations must be normalized by some parameter (here called the "magnetic activity" [m_a]) that compensates for these nonfield

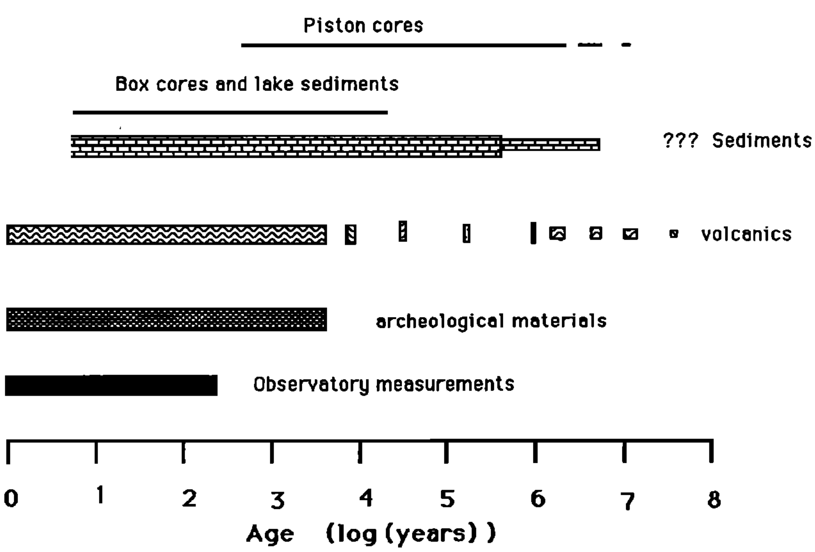

Figure 2. Potential sources of paleointensity information.

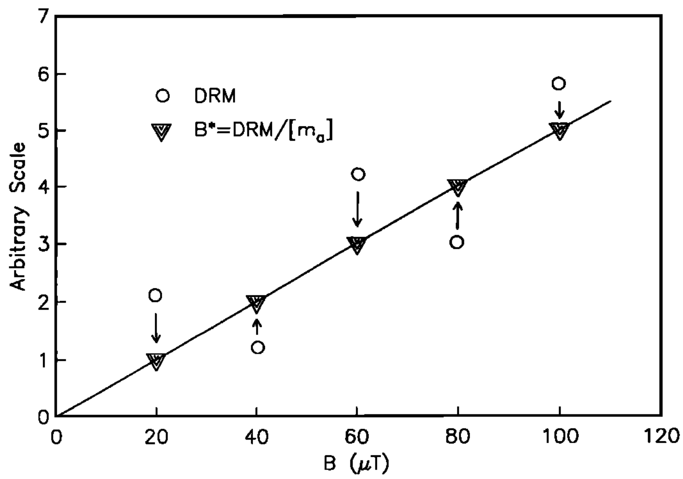

Figure 3. Assumptions in using sedimentary sequences for relative paleointensity. DRM is the detrital magnetization, B^* is the relative paleointensity estimate obtained by normalizing by the "magnetic activity" $[m_a]$, which reflects the contribution of changes in nonfield effects such as magnetic concentration, grain size, etc.

effects in order to obtain the contribution of the magnetic field. Under the ideal conditions shown in Figure 3, DRM normalized by magnetic activity yields estimates of the paleofield intensity B^* that is linearly related to the ancient magnetic field. Of course, the slope relating B^* and B may never be known.

The experimental and theoretical basis for estimating $[m_a]$ is relatively undeveloped. I will highlight results from landmark papers that serve as the foundations for all sedimentary paleointensity studies. These papers provide hope that there may be sediments that retain a recoverable record of paleofield intensities if properly normalized. I will describe the conditions under which such records are created and the available techniques for analyzing them. Finally, I have collected some representative records of relative paleointensity from sediments and will draw a few general conclusions from them.

2. SOME DEFINITIONS AND CONVENTIONS

The paleomagnetic literature suffers from confusion about units and terms and is therefore at times difficult to read and to interpret. I have redrafted all figures in order to standardize the units according to the Système International (SI). Table 1 lists some useful conversions among the various units in the original literature. I will describe all magnetizations in terms of moment m or magnetization M. Magnetic fields are denoted either as H or as B. B is usually referred to as the magnetic induction, but often in the paleomagnetic context B is also called the magnetic field; the two will be distinguished by using the appropriate units and associated letter. I note in passing that the change from using cgs units to SI units in the early 1980s has led to an enormous amount of confusion because many measurements are still made in cgs units and conversion is ubiquitous and frequently done incorrectly. Nearly every table in the paleomagnetic literature listing unit conversion contains confusing, misleading, or erroneous statements [see *Payne*, 1981; *King et al.*, 1983; *Butler*, 1992]. I will adhere to the usage in Table 1 throughout the paper.

Magnetic remanence is the magnetization measured in the absence of an applied field. I will call, by convention, natural remanence, detrital remanence, and anhysteretic remanence NRM, DRM, and ARM, respectively. A list of various acronyms and their meanings appears in Table 2.

3. EXPERIMENTAL AND THEORETICAL FOUNDATION

Magnetic grains tend to align with the ambient magnetic field as they fall through the water column. This alignment can be preserved through the process of deposition, becoming locked in at some depth below the sediment/water interface. The slight statistical alignment preserved results in a magnetic remanence

TABLE 1. Some Useful Conversions

Parameter	*Symbol*	*SI*	*cgs*	*Conversion*
Magnetic moment	m	A m^2	gauss cm^3 (G cm^3)	1 A m^2 = 10^3 G cm^3
Magnetization	M	A m^{-1}	G	1 A m^{-1} = 10^{-3} G
Polarization	J	tesla (T)	G	1 T = 10^4 G
Magnetic field	H	A m^{-1}	oersted (Oe)	1 A m^{-1} = $4\pi \times 10^{-3}$ Oe
Magnetic induction	B	T	G	1 T = 10^4 G
Permeability	μ_0	henry (H) m^{-1}	1	$4\pi \times 10^{-7}$ H m^{-1} = 1 cgs unit
Viscosity	η	pascal s (Pa s)	poise	10^{-1} Pa s = 1 poise
Susceptibility	χ			
Total (m/H)		m^3	G cm^3 Oe^{-1}	1 m^3 = ($10^6/4\pi$) G cm^3 Oe^{-1}
By volume (M/H)		...	G Oe^{-1}	1 SI unit = ($1/4\pi$) G Oe^{-1}
By mass ((m/mass)($1/H$))		m^3 kg^{-1}	G cm^3 g^{-1} Oe^{-1}	1 m^3 kg^{-1} = ($10^3/4\pi$) G cm^3 g^{-1} Oe^{-1}

Some additional conversions are as follows: 1 T = kg A^{-1} s^{-2}; 1 H = kg m^2 A^{-2} s^{-2}; 1 poise = 1 dyn s cm^{-2}; 1 emu = 1 G cm^3; $B = \mu_0 H$; and 1 Pa = 1 N m^{-2}.

TABLE 2. Some Definitions and Conventions Used in the Paper

Magnetization	*Symbol*	*Mode of Acquisition*
Natural remanence	NRM	acquired in nature
Thermal remanence	TRM	acquired by cooling from high temperature, in the presence of a dc bias field
Viscous remanence	VRM	time dependent magnetization acquired by sitting in a field
Detrital remanence	DRM	acquired by detrital magnetic grains in a sedimentary environment
Isothermal remanence	IRM	acquired by exposure to an instantaneous field
Anhysteretic remanence	ARM	acquired when subjected to an alternating field of decreasing peak amplitude, in the presence of a dc bias field
Stirred remanence	StRM	magnetization of a thick slurry acquired in the laboratory after stirring in the presence of a dc bias field
Saturation remanence	M_{sr}	saturation IRM
Saturation magnetization	M_s	magnetization measured in the presence of a saturating field
Equilibrium magnetization	M_e	magnetization of a population of grains suspended in a fluid
(Initial) susceptibility	χ	slope relating induced magnetization to a low applied field
Relative paleointensity	B^*	estimated relative intensity of the geomagnetic field based on normalized remanence

that is controlled in part by the magnetic field vector prior to locking in. The journey of a magnetic particle to its resting place in the sedimentary sequence thus begins as it settles through the water (see Figure 4). The behavior of magnetic particles in a viscous fluid has been considered by many [e.g., *Nagata*, 1961, 1962; *Collinson*, 1965; *Stacey*, 1972; *Rees and Woodall*, 1975; *Tucker*, 1980*a*, *b*; *Hamano*, 1980; *Denham and Chave*, 1982; *Tauxe and Kent*, 1984; *Yoshida and Katsura*, 1985]. While in the water, a magnetic grain is subjected to a hydrodynamic couple generated by fluid shear, a magnetic couple tending to align the magnetic moment with the ambient magnetic field, viscous drag and inertial forces tending to oppose motion, and thermally derived random motions. Furthermore, when a particle strikes the ground, it will also be subject to a gravitational couple which will tend to bring the particle into the nearest depression. After a grain crosses the sediment/water interface, it enters a region where the sediment undergoes initial consolidation during which the effective viscosity experienced by the grain increases rapidly and the grain becomes "stuck." While it is still near the sediment/water interface, however, there is a chance that the grain will experience a sudden drop in effective viscosity caused by bioturbation, slumping, etc., and the grain may again come into equilibrium with the prevailing magnetic and other forces. Finally, when a grain is buried deeply, it is in a region where the sediment undergoes compaction and the magnetic particles may rotate in response. I will consider the experimental and theoretical aspects of each of these regions in the following, beginning with the water column and initial deposition. Since the process is a continuum, I have abandoned the traditional separation of remanence into depositional and postdepositional flavors [*Irving and Major*, 1964] and call the remanence carried by detrital magnetic grains simply DRM (see Table 2).

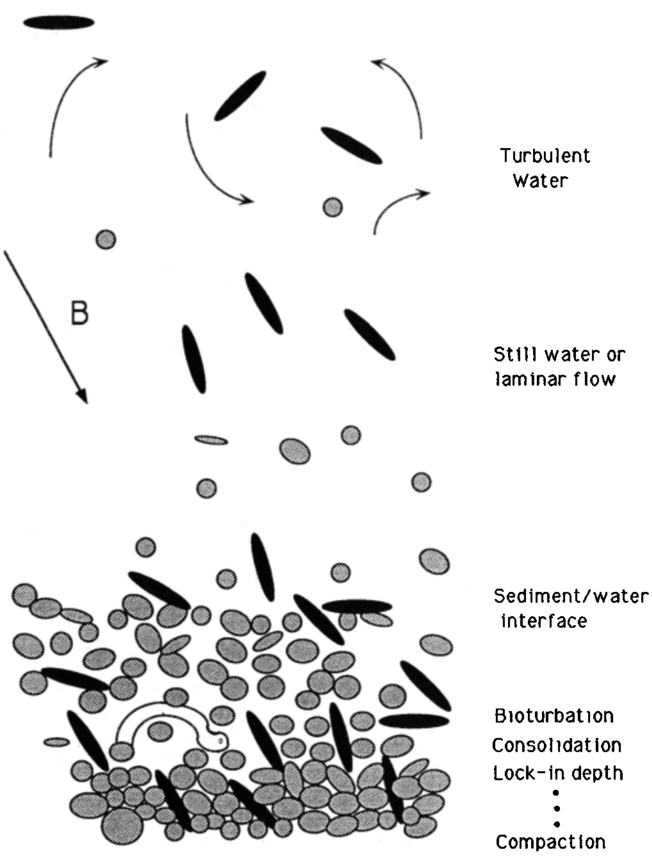

Figure 4. Schematic drawing of the journey of magnetic particles from the water column to burial.

4. BEHAVIOR IN THE WATER AND AT THE SEDIMENT/WATER INTERFACE

Much of the rather slim body of theoretical work on the magnetization of sediments [e.g., *Nagata*, 1961,

TABLE 3. Typical Values for Earth Materials and Fields

	Value
Magnetization	
Basalts	$>10^{-1}$ A m^{-1}
Red sediments	$\sim10^{-3}$ A m^{-1}
Limestones	$10^{-4}-10^{-5}$ A m^{-1}
Magnetite	4.8×10^{5} A m^{-1}
Field Strength	
Present field	20–70 μT
Past fields	~2–200 μT
Viscosity of Water	
At 0°C	1.792 cP
At 20°C	1.005 cP

1962; *Graham*, 1949; *King*, 1955; *Nozharov*, 1967; *Griffiths et al.*, 1960; *Irving*, 1957; *Collinson*, 1965; *Stacey*, 1972] and many experiments investigating what has been called "grain by grain" deposition [e.g., *Johnson et al.*, 1948; *Clegg et al.*, 1954; *Vlasov et al.*, 1961; *Vlasov and Kovalenko*, 1964; *Rusakov*, 1966; *Khramov*, 1968; *Levi and Banerjee*, 1975; *Blow and Hamilton*, 1978; *Barton and McElhinny*, 1979; *Barton et al.*, 1980; *Yoshida and Katsura*, 1985; *Lu et al.*, 1990; *Amerigian*, 1977] have considered the remanence acquired when magnetic particles fall through the water column and hit the sediment/water interface. The broader implications for paleomagnetism of sediments were ably reviewed by *Verosub* [1977], and I will review the aspects relevant to paleointensity studies here.

For now, I will be concerned with remanence acquired at the time of deposition and will ignore the effect of possible realignment after the particles cross the sediment/water interface. It has been repeatedly shown that there is little change of a magnetization acquired in such a manner, unless the sediment is significantly disturbed (see, for example, *Barton et al.* [1980]), and in the absence of such disturbance, postdepositional realignment (in the zone of initial consolidation) can be neglected for the most part. Thus in this section I consider what has been traditionally called "depositional detrital remanent magnetization."

Although primarily interested in magnetic fabric, *Rees and Woodall* [1975] attempted to quantify some of the forces acting on a particle falling through water and landing on the sediment/water interface. They found that for large grains, hydrodynamic and gravitational forces probably dominate, whereas for smaller grains, magnetic coupling can be quite important. Since large magnetic particles are also magnetically unstable, I will be primarily concerned with grains in which the magnetic couple is a substantial force.

Collinson [1965] was among the first to consider theoretically the motion of magnetic particles in water (see also *Nagata* [1961, 1962]), and I follow him by starting with the equation of motion for a magnetic particle with moment m in a viscous fluid in response to the magnetic couple generated by the field B at angle θ:

$$I\ddot{\theta} + \lambda\dot{\theta} + mB\sin\theta = 0 \tag{1}$$

where $\lambda\dot{\theta}$ is the viscous damping torque, I is the moment of inertia of a grain about an axis through the center, and θ is the angle that the direction of magnetization of the grain makes with the applied field.

By neglecting the inertial term and taking θ to be small, the solution to (1) is

$$\tan\frac{\theta}{2} = \tan\frac{\theta}{2}\, e^{(-mBt/\lambda)}$$

The time τ needed for θ to decay to $1/e$ of its initial value is

$$\tau = \lambda/mB$$

Stacey [1972] performed some useful substitutions for λ and arrived at the following equation for τ:

$$\tau = 6\eta/MB$$

where M is magnetization per unit volume and η is the viscosity of the fluid. Plugging in reasonable values for these parameters (see Table 3) shows that for small grains (say, diameter $d < 15$ μm), τ is of the order of a second or less. He further calculated how far a 10-μm grain would fall during this time using Stokes's law and estimated the distance to be about 0.25 mm. From this simplified approach one expects that the net moment of a magnetic population suspended in water, as well as the remanence acquired from magnetic particles settling out of still water, will be close to saturation, hence independent of field strength.

Yoshida and Katsura [1985] actually measured the magnetization of a population of magnetic grains suspended in water. Measurements made of the "equilibrium magnetization" M_e of resuspended calcareous ooze as a function of applied field are shown in Figure 5. First, it is clear that even for fields approaching Earth values (see Table 3), M_e is not independent of the applied field. Second, the saturation remanence of the material is about 3×10^{-3} A m^2 kg^{-1} or about a factor of 3 higher than that of the maximum equilibrium magnetization attained.

The work of *Yoshida and Katsura* [1985] showing a field dependence of M_e on B was naturally not available to *Collinson* [1965]. However, he did know from the classic work of *Johnson et al.* [1948] that remanence acquired by deposition from water was dependent on the magnetic field and was quite a bit lower than saturation remanence. The work of *Johnson et al.* [1948] contains the seeds of much of modern thought on sedimentary magnetization and is an excellent initial assessment of the reliability of sediments for paleogeomagnetic field studies. They redeposited natural

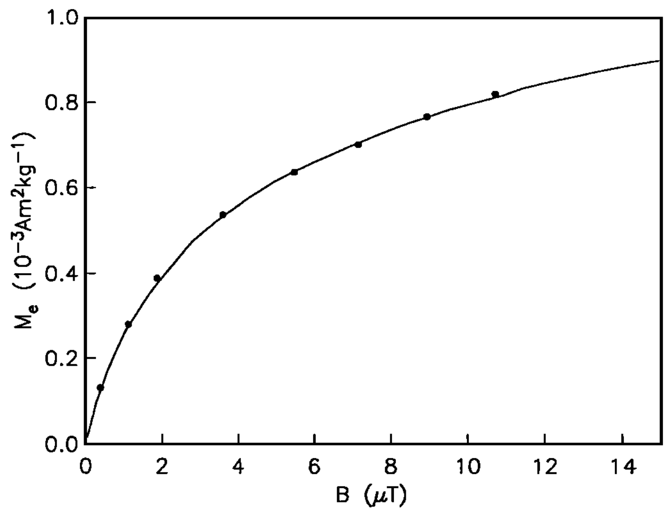

Figure 5. Plot of the equilibrium magnetization M_e of a suspension of carbonate ooze as a function of applied field B. Redrawn from *Yoshida and Katsura* [1985].

sediments and attempted to normalize remanence both by redeposition in the laboratory and with saturation isothermal remanence. Results from one experiment are shown in Figure 6, in which the magnetization acquired after redeposition of varved clays is plotted as a function of magnetic field (B_0 was the field present at the site of original deposition, or about 50 μT). They also noted that temperature dependence of the magnetization was below the limit of detection. This latter was questioned by *Løvlie* [1974] but reconfirmed by *Barton and McElhinny* [1979].

A field dependent remanence clearly demands that some additional factors be included in the theoretical treatment of detrital remanence. Although *Graham* [1949] noted the possible influence of Brownian motion, *Collinson* [1965] was the first to consider the effect quantitatively. He estimated the deflection in radians (θ) to be about

$$\theta^2 = kT/mB \qquad (2)$$

Furthermore, if Brownian motion is important, then sedimentary magnetization can be modeled using the Langevin theory for a paramagnetic gas or

$$DRM/M_{sr} = \coth\,(mB/kT) - (kT/mB) \qquad (3)$$

In order to estimate the grain size dependence of the effect, *King and Rees* [1966] substituted M, a volume-normalized moment and the grain diameter d into (2); thus

$$\theta^2 = kT/d^3MB$$

and they estimated the maximum grain size affected by Brownian motion to be no larger than 0.1 μm. They envisioned grains either as being entirely aligned with the field or, if they are small enough to be affected by Brownian motion, as being essentially random, a result not supported by experimental data.

According to *Stacey* [1972, p. 140], "Clearly there is no possibility of explaining the field dependence of detrital remanence [without some misaligning effect on the magnetic moments]." He favored Brownian motion and noted that if a uniform distribution of moments with some maximum value is plugged into (3), the results of *Johnson et al.* [1948] (plotted in Figure 6), could be reasonably well modeled. This he took as powerful support for the idea of pseudosingle domains with rock magnetic characteristics (such as specific thermal remanence) that vary smoothly between values typical of single-domain grains and those of true multidomain grains. The idea of a distribution of moments was refined later by *Yoshida and Katsura* [1985], who showed that a lognormal distribution of grain moments fits their data better than the uniform distribution assumed by *Stacey* [1972]; moreover, a lognormal distribution is certainly reasonable from a geological point of view.

Most of the work on the paleomagnetism of sediments has been inspired by a desire to study the directional behavior of the ancient magnetic field. One of the first observations made by redepositing sediments in the laboratory was that the observed inclination I_o was shallower than that of the applied field I_f [see *Johnson et al.*, 1948]. Although the so-called "inclination error" may appear to be irrelevant to paleointensity studies, one investigation [*Tauxe and Kent*, 1984] showed that records displaying inclination shallowing probably will not yield reliable intensity information either.

Tauxe and Kent [1984] collected mud from a riverbank that had been deposited in a recent rainstorm. The natural remanence of the sediments was shallower by 30° than the field at the sampling site (50°), a rare demonstration that inclination error occurs in nature

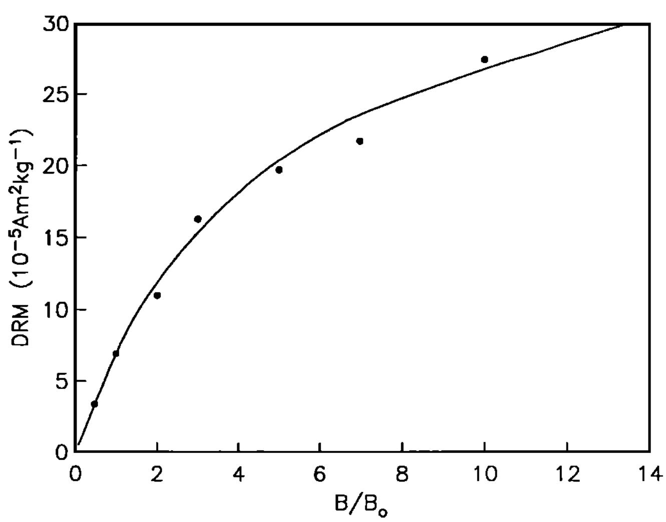

Figure 6. Plot of DRM versus applied field B for redeposited glacial varves. B_0 is the field at the site. Dilute slurries were allowed to settle in an applied field, dried, and measured. Redrawn from *Johnson et al.* [1948].

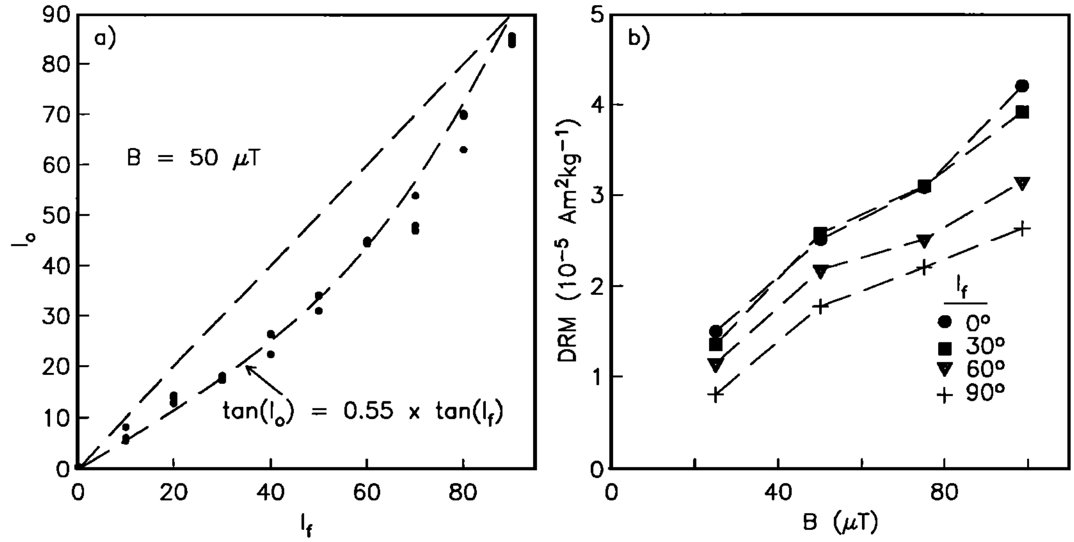

Figure 7. Data from redeposited recent river sediments, showing magnetization DRM versus applied field B for various inclinations. Dilute slurries were allowed to settle in an applied field; settling tubes were then inserted into a cryogenic magnetometer. (a) Observed inclination I_o versus inclination of the applied field (I_f). (b) DRM versus B. Redrawn from *Tauxe and Kent* [1984].

as well as in the laboratory. They then redeposited the sediments in a variety of conditions of the applied field. When allowed to settle in a 50-μT field of variable inclination, the observed inclinations followed the relationship first documented by *King* [1955] of $\tan(I_o) = f \tan(I_f)$ as shown in Figure 7a. Furthermore, when the sediment was deposited in various conditions of B and I_f, the resulting intensity was found to be a strong function not only of B but also of I_f, whereby I_f alone was responsible for factor of 2 changes in observed magnetization (Figure 7b). If this is generally true for sediments exhibiting inclination error, it will be practically impossible to separate out the contribution of field strength from that of inclination.

The suitability of remanence acquired by settling directly from the water column for relative paleointensity was called further into question by the work of *Barton et al.* [1980]. They studied a range of detrital remanences, and one series of experiments is of particular interest here. They made slurries from natural sediments and allowed them to settle in a magnetic field. No further disturbance of the sediment occurred after deposition. Typical results are shown in Figure 8, where the magnetization is plotted versus the field applied during settling. There is a general increase with increasing field strength as noted previously, but the curve is not linear (indeed, not easily fit with any simple curve) and the nonlinearity is not "noise" but appeared to *Barton et al.* [1980] to be reproducible (see also *Tucker et al.* [1980a]).

The extent to which these laboratory experiments are relevant in nature is a matter of some dispute. Certainly, the particle flux in the laboratory experiment is quite high, leading to a source of considerable disturbance. Nonetheless, such high particle fluxes are expected in many natural environments, such as in rivers, turbidity currents, and perhaps also in near-shore lake and marine environments. Slow-deposition rate lake and marine environments, however, are probably not well represented by these experiments, and certainly some of these sediments exhibit none of the telltale "inclination error" observed in the higher-energy deposits and laboratory analogues (see, for example, *Sprowl and Banerjee* [1989]). Redeposition experiments using very slow rates also show no inclination error [*Barton and McElhinny*, 1979], but the relationship of remanent intensity to that of the applied field has not been investigated.

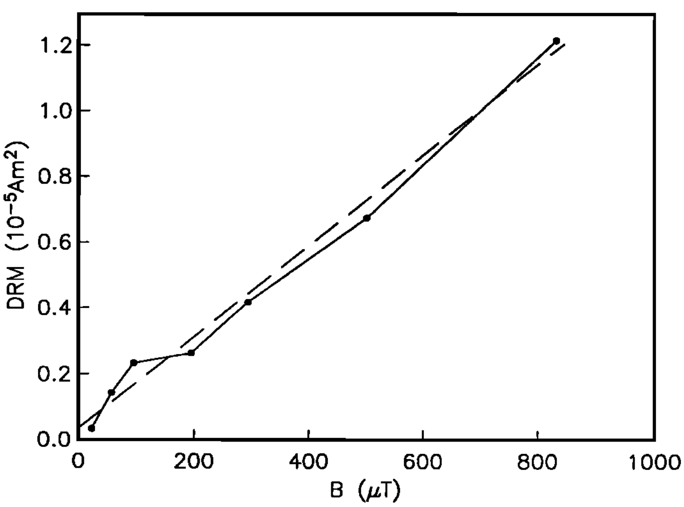

Figure 8. Magnetization versus applied field of redeposited lake sediments. Sediments were redeposited in a similar manner as those shown in Figure 7. Redrawn from *Barton et al.* [1980].

5. BEHAVIOR DURING INITIAL CONSOLIDATION

Let us follow our particles in their journey across the sediment/water interface into the region of initial consolidation (see Figure 4). This is the zone in which what has been termed "postdepositional remanent magnetization" is acquired. *Tucker* [1980*b*] separated the processes active during initial consolidation into two categories: (1) the case that there is no outside disturbance, and (2) the case that there are perturbations (such as stirring, bioturbation, slumping and so on). He notes (p. 153),

> In the absence of external perturbations, the internal constraints on grain movement (void size and rigidity of the sediment, entrapped gas, cohesive forces and friction, etc.) may be sufficient to largely inhibit realignment by any magnetic torque acting on the remanence carriers. It has been shown that for realistic field strengths (<200 A/m) only a small fraction of the carriers may be susceptible to realignment in this way. For the larger-scale realignments which have been proposed in order to account for the natural remanences of many fine-grain sediments, it is necessary to postulate the presence of additional time-dependent disturbances to the sediment. These may include local 'stirring' by for example bioturbation, shaking via earth movements or gross movement of the sediment during slumping. These mechanisms may temporarily reduce or remove the constraints on grain movement just as heating or the application of high alternating field reduce or remove the effective barriers to domain-wall movement (or domain rotation) in TRM or ARM acquisition respectively.

Since there has been persistent confusion in the literature caused by lumping remanence acquired by both processes into the single term "postdepositional remanent magnetization," I have avoided this term. I will instead discuss the remanence acquired during initial consolidation with and without disturbance separately. First, I will consider the theoretical and experimental data available relevant to the initial consolidation without disturbance (experiments in which slurries are shaken, not stirred).

5.1. Initial Consolidation Without Disturbance

Several types of experiments have been undertaken to establish the behavior of magnetization acquired just under the sediment/water interface, in the absence of outside disturbance. These include (1) wetting a powder and allowing it to stand in the presence of a field [e.g., *Irving and Major*, 1964], (2) deposition of a slurry in zero field and allowing it to stand in a field [*Clegg et al.*, 1954; *Barton et al.*, 1980; *Tauxe and Kent*, 1984; *Khramov*, 1968], (3) deposition of a slurry in a field, changing the field conditions, and then monitoring changes in remanence [*Johnson et al.*, 1948; *Løvlie*, 1974, 1976; *Graham*, 1974; *Barton and McElhinny*, 1979; *Tucker*, 1979, 1980*a*; *Verosub et al.*, 1979; *Payne and Verosub*, 1982], and (4) monitoring of remanence while the sediment undergoes compaction [e.g., *Blow and Hamilton*, 1978; *Hamano*, 1980; *Otofuji and Sasajima*, 1981; *Anson and Kodama*, 1987; *Deamer and Kodama*, 1990]. The magnetization acquired in this region has also been considered theoretically [*Tucker*, 1980*a*; *Hamano*, 1980, 1983; *Denham and Chave*, 1982; *Hyodo*, 1984]. I will call attention to the aspects of these experiments important for paleointensity studies.

The first type of experiment described above was attempted by *Irving and Major* [1964] in a classic paper on postdepositional remanence. Their hypothesis was that "soon after deposition, a sediment is likely to have within it voids that are as large or larger than the magnetic particles, which are therefore free to rotate" [*Irving and Major*, 1964, p. 136]. *Clegg et al.* [1954] and *Irving* [1957] had noted the possibility of such a remanence, and *Irving and Major* [1964] tested the physical plausibility of the phenomenon's occurring in nature. They made a synthetic sediment from quartz plus 0.5% magnetite using two grain sizes (silt and sand). The synthetic sediment was packed dry into cylinders in zero field, and flooded for 15–70 hours in fields from 25 to 113 μT. The samples were dried prior to measurement. The magnetization was quite stable, and contrary to experiments such as those, for example, of *Johnson et al.* [1948], it exhibited no inclination error. Also, the magnetization appeared to be roughly proportional to the applied field, although the data on this question were quite scanty (two or three fields for each type of sediment). The relevance of these synthetic samples and the contrived rewetting experiment has been questioned [see *Barton et al.*, 1980]; nonetheless, these results have been held up as proof that postdepositional alignment can occur and that natural sediments need not suffer from inclination error if indeed such postdepositional rotation occurs.

Redeposition in zero field and then exposing the sediment to a field was first done by *Clegg et al.* [1954]. They deposited powdered sandstone in zero field, decanted the water, and then set the sediment in a field for a given time. The mud was dried, subsampled, and measured. *Barton et al.* [1980] and *Tauxe and Kent* [1984] performed a similar experiment with one important difference; they used cryogenic magnetometers which allowed measurement of the sediments without drying or disturbance and also allowed remanence acquisition to be more easily measured over time. *Tauxe and Kent* [1984] monitored remanence after exposure to a field and again after allowing the magnetization to "relax" for an equivalent amount of time to the exposure time. This was done to determine the truly fixed moment as the remanence behaved quite "viscously." Tauxe and Kent's published and unpublished results are shown in Figure 9. The results without allowing the remanence to relax are similar to those of *Barton et al.* [1980], in that there is a quasi-logarithmic growth in magnetization with exposure time. After relaxation, however, the sediments lost some 30% of the initial magnetization, and the remanence growth curve is not logarithmic but follows a

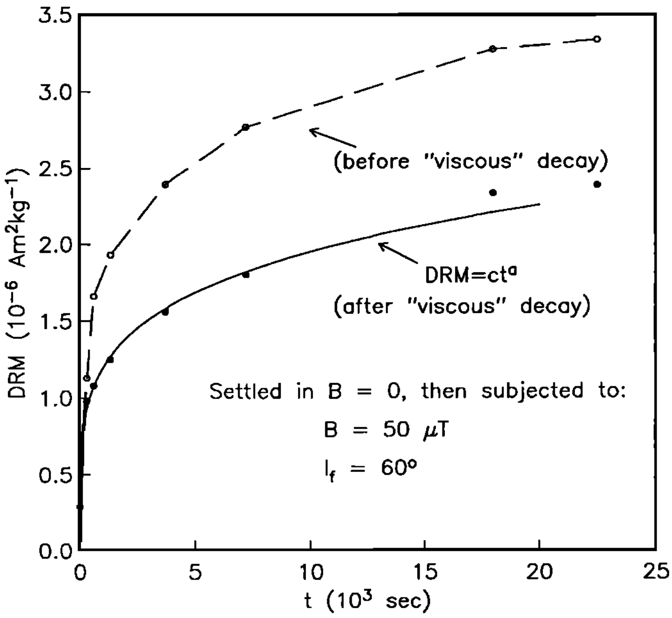

Figure 9. Magnetization versus time for redeposited recent river sediments. Dilute slurry settled in zero field, and was then exposed to a field for time *t*. Dashed curve is data measured immediately after exposure, and solid line is data measured after the sediment sat in zero field for a time equivalent to the exposure time. Parameters *c* and *a* were fitted as 2.7×10^{-7} and 0.211, respectively. Data from *Tauxe and Kent* [1984, also unpublished data, 1984].

power law as shown in the solid line in Figure 9. Another important result demonstrated by this and other experiments was that the remanence acquired by postdepositional grain rotation is some 10% of that acquired by deposition in the same field. It is likely that the "viscous" remanence observed by *Tauxe and Kent* [1984] is not a true viscous remanent magnetization (in the sense described in Table 2 whereby the grain moments rotate, not the grains themselves) but is carried by grains that remain mobile in the sense envisioned by *Irving* [1957]; most importantly, the mobile fraction (some 3%) is a very small component of the total magnetization acquired during deposition. The magnetization acquired by the power law represents grains that are rotating against barriers, an apparently irreversible process. I will consider this phenomenon further when I discuss the work of *Tucker* [1980*a*], *Hamano* [1980], and *Otofuji and Sasajima* [1981].

The third and most common type of experiment on so-called postdepositional remanence (without disturbance) is deposition in one field followed by monitoring remanence acquired in some different field. Experimental details vary substantially among the many attempts. The principal features of the experiments are, however, that a natural or artificial sediment is magnetized in some fashion, either by deposition or by some postdepositional disturbance such as stirring. Then the sediment is exposed to a second field for a period of time. The sediment is usually dried prior to measurement. It has long been known [e.g., *Henshaw and Merrill*, 1979] that drying can alter the magnetic remanence, and experiments involving drying naturally compound the effects of drying with those of rotation in the wet sediment. The latter is the process under study, and results from experiments in which the sediment was dried are more difficult to interpret.

The first of the "rotation" experiments was an accident described by *Johnson et al.* [1948]. One redeposition column was rotated inadvertently and no postdepositional remagnetization was observed. *Løvlie* [1974] showed that some postdepositional realignment of grains persisted for several days in his experiment and later [*Løvlie*, 1976] proposed that there was a zone of consolidation within the sediment in which the largest grains would be blocked first. *Tucker* [1979] showed that smaller grains remained mobile longest in his experiments. *Barton and McElhinny* [1979], *Verosub et al.* [1979] and *Payne and Verosub* [1982] all found that for the most part, very little magnetization was realigned by subsequent fields. The most complete, elegant, and elaborate experiments on this type of behavior were done by *Tucker* [1980*a*], and I will review these in some detail.

Key diagrams from *Tucker* [1980*a*] have been redrawn in Figure 10. In his experiments, Tucker deposited synthetic sediments (silica and carefully sized magnetite) in applied fields and then rotated the field 90° and monitored remanence growth in this new direction. He used a sensitive gradiometer, obviating the need for drying, subsampling, or even moving the mud during the experiment. As noted in previous experiments, the magnetization acquired in this way is but a fraction of the original remanence. Nonetheless, the results shed considerable light on possible mechanisms controlling grain mobility in the zone of initial consolidation.

In Figure 10*a*, I replot some of Tucker's data on magnetization acquired versus log (time). These measurements are made immediately after the field was "switched off" and thus are comparable to the dashed line in Figure 9. Tucker reported logarithmic decay of the magnetization with time, but only a small fraction is lost in this fashion (unlike the results shown in Figure 9, where some 30% is lost). First, he observed rapid buildup or decay of magnetization as soon as the field was switched on or off (data not shown). Next, he fit the data shown in Figure 10*a* with several straight-line segments. The magnetization acquired in a 19-μT field, for example, was interpreted to show two linear segments, one up to about 300 s, after which there was no additional growth of remanence. These two line segments are also seen in the data acquired in a 63-μT field, but the "kick point" is shifted to about 100 s. After about 120 s in a 63-μT field, there is another phase of remanence growth and a second saturation, occurring at close to 1000 s. The top two curves were interpreted to show a third phase of remanence growth

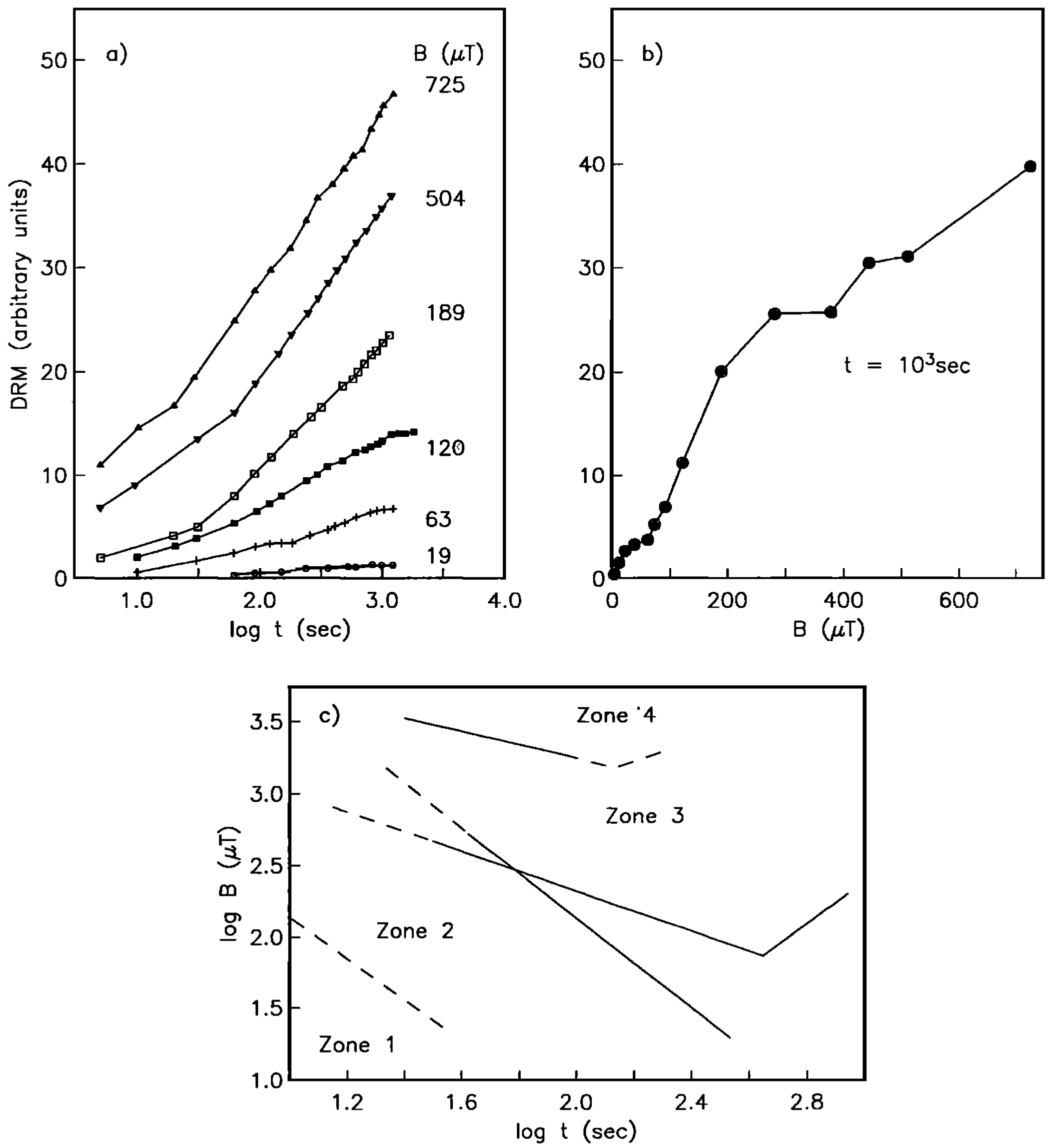

Figure 10. (*a*) Magnetization versus time for redeposited synthetic sediments. Dilute slurry was allowed to settle in a field; then a perpendicular field was applied, and the magnetization along this axis was monitored versus time and using a variety of applied fields. (*b*) Magnetization acquired after 1000 s versus applied field. (*c*) Schematic diagram illustrating modes of remanence acquisition as a function of time and field strength (see text and *Tucker* [1980*a*]). (Figures 10*a*–10*c* are redrawn from *Tucker* [1980*a*].)

beginning at about 100 s for fields of 500 μT. A final phase of remanence growth was observed for fields of 725 μT after about 4 hours (not shown).

Figure 10*b* shows remanence acquired after 1000 s as a function of the applied field. There are many similarities between this complex curve and that shown in Figure 8 for the magnetization acquired during original deposition [*Barton et al.*, 1980] as noted by *Tucker* [1980*a*]. Indeed, if sediments are magnetized in such a way in nature, relative paleointensity estimates are a pipe dream.

Tucker interprets his data in terms of four behavioral zones (shown schematically in Figure 10*c*). The first zone is that in which there is a rapid and reversible buildup of magnetization. This he attributes to rotation of very loosely bound grains such as larger grains against the still elastic matrix. This would be the same carrier as that responsible for the "viscous" magnetization lost in Figure 9 after "relaxation." In the second zone, magnetization is built up over minutes and decays over tens to hundreds of hours. He sees this zone as being controlled by rotation of grains in fluid-filled voids. Zone 3 behavior was observed in fields higher than about 44 μT; here grains may actually move against the matrix, shoving aside small obstacles. This magnetization is irreversible and inhibited by compaction and is probably the type responsible for the solid curve in Figure 9, fit by *Tauxe and Kent* [1984] with a power law curve. A fourth zone was evident in rather high fields (above 441 μT) which was interpreted by *Tucker* [1980*a*] as rotation against surface tension forces because the behavior was suppressed in fluids with lower surface tensions.

Tucker [1980*a*] considered only very briefly the role of compaction in blocking the magnetization. Several investigations have sought to constrain the effect of compaction and dewatering on the magnetization of sediments (see, for example, *Blow and Hamilton* [1978], *Hamano* [1980], *Otofuji and Sasajima* [1981], *Stober and Thompson* [1979], *Tucker* [1980*b*], *Anson and Kodama* [1987], and *Deamer and Kodama* [1990]). *Anson and Kodama* [1987] and *Deamer and Kodama* [1990] were more concerned with effects occurring during burial, and I will discuss their results later. The work of *Stober and Thompson* [1979] failed to consider possible chemical changes in their sulfide-rich sediments as a consequence of drying out (known nowadays as chemical demagnetization), and their conclusions about the magnitude of the effect of drying on remanence are suspect.

The works of *Hamano* [1980] and *Otofuji and Sasajima* [1981] are similar in that they both attempt to constrain the remanence acquired during initial consolidation. *Hamano* [1980] compacted the sediments with an overburden pressure, and *Otofuji and Sasajima* [1981] used a centrifuge. *Hamano* [1980] found that for a given grain shape, there is a critical void ratio at which the time constant of alignment goes essentially infinite, independent of grain size. *Otofuji and Sasajima* [1981] found that the remanence blocked between two given void ratios was independent of that acquired between a different pair of void ratios and postulated a law of additivity for such remanences.

Experimental work on the blocking of magnetic remanence gave rise to several theoretical papers [*Denham and Chave*, 1982; *Hamano*, 1983; *Hyodo*, 1984]. *Denham and Chave* [1982] started with (1) and solved it using the small-angle approximation where $\sin\theta \simeq \theta$ and grain radius is r:

$$\ln \frac{\theta(t)}{\theta_0} = -\int_{t_0}^{t} \frac{dt}{\tau(t)} \tag{4}$$

where

$$\tau(t) = (8\pi r^3/mB)\eta(t)$$

Echoing earlier work, they pointed out that τ is very short in water, but they underscored the point that it is proportional to the effective viscosity $\eta(t)$. This was the basis of their "viscosity theory" of postdepositional remanence.

They took (4) as the step response R of a particle to a change in field. This then can be differentiated to get the impulse response S. Assuming a profile of τ allows the calculation of a system response for a given step change in field. They took as a first stab an exponential curve for τ, presuming that viscosity varies linearly with particle concentration; hence, τ would vary exponentially in sediments as

$$\tau(t) = \tau_0 e^{At}$$

where A is the characteristic growth rate of the alignment time constant. So at time t_1 after a step change in θ

$$\ln R = \ln \frac{\theta(t)}{\theta_0} = \frac{1}{A\tau_0} e^{-At_1}$$

Since any profile of $\tau(t)$ can be plugged in, they proceeded to consider an empirical viscosity profile for sediments based on Mooney theory. The start of Mooney theory is constant-size spheres. Given these, the viscosity of a suspension is a function of the particle concentration x and the maximum permitted concentration x_{max}. *Denham and Chave* [1982] chose the form

$$\frac{x}{x_{max}} = 1 - e^{Ct}$$

where C is the characteristic rate of consolidation. Putting it all together, they got

$$\ln \frac{\tau}{\tau_0} = 2.5x_{max}(e^{Ct} - 1)$$

This can be manipulated to get R and S. The lock-in depth is the maximum of S.

In a later attempt at modeling the magnetization acquired during initial consolidation, *Hyodo* [1984] made the point that most workers had considered only an increase of viscosity due to compaction as a factor in determining the lock-in depth of magnetic grains. He noted that *Otofuji and Sasajima* [1981] demonstrated that rotation of magnetic particles does not occur steadily "but in a halting fashion and that the rotation is dominantly controlled by a maximum friction with matrix grains but not by the viscosity" [*Hyodo*, 1984, p. 46]. This realization is supported by the work of *Tucker* [1980*a*] as well.

Hyodo [1984] began by assuming (based on the work of *Hamano* [1980]) that the amount of magnetic moment blocked at t, ($r(t)$), is exponential with depth:

$$r(t) = Ce^{-At}$$

Further assuming that each incremental moment fixed at a particular time is aligned parallel to B and that each moment is additive, the magnetization of a package of age τ which has been exposed to the geomagnetic field variation $f(t)$ since time t is

$$M(\tau) = \int_0^{\tau} \frac{M_0}{f_0} f(t) r(\tau - t)\, dt$$

This model implies that the remanence can be calculated by a convolution integral of a magnetic field variation $f(t)$ and the moment-fixing function $r(t)$. If this is true, then (1) the magnetization never coincides exactly with the external field unless it is constant or $r(t)$ is Dirac's δ function, and (2) the magnetization intensity results from a component proportional to the field intensity and a component of intensity reduction caused by the cancellation due to the superposition of magnetic moments integrating over $f(t)$.

He posed the problem as a convolution of $f(t)$ and $r(t)$. If $M(\omega)$, $F(\omega)$, $R(\omega)$ are the spectra of $m(t)$, $f(t)$, $r(t)$ respectively, then

$$M(\omega) = F(\omega)R(\omega)$$

and $R(\omega)$ gives the details of the filter. Making some approximations in $r(t)$, he got

$$R(\omega) = 1/[1 + i(\omega/A)]$$

This he split into the real and imaginary parts; substituting in $T = 2\pi/\omega$ and $\alpha = 2\pi/\ln(0.5)$, the amplitude is given by

$$|R| = 1/[1 + (\alpha T_{1/2}/T)]^{1/2}$$

and the phase is given by

$$\theta = \tan^{-1}(\alpha T_{1/2}/T)$$

where $T_{1/2}$ is the half fixing period. He suggested that because secular variation is small (less than 20°) the declination and inclination can be operated on as if they are linear functions and calculated the amplitude decrease and phase lag for different $T_{1/2}$. For example, if $T_{1/2}$ is 500 years, a directional swing would be reduced in amplitude to 0.13 and the phase lag would be 115 years. If $T_{1/2}$ is 250, the reduction is 0.25 and the delay is 53 years.

Hyodo [1984] proposed the use of this model to quantify the alteration of the secular variation recorded in sediments. His model predicts (1) amplitude attenuation of declination and inclination records as well as a phase lag and (2) intensity variation resulting from the cancellation of magnetic moments aligned in the opposite directions.

5.2. Initial Consolidation With Disturbance

It was pointed out by *Kent* [1973] that most sediments in the deep sea are intensively burrowed and that the magnetization acquired during initial deposition and consolidation may well be completely irrelevant to the natural remanence of such sediments. Because the remanence acquired during rapid deposition often exhibits inclination error and responds in a complex manner to the applied field (see Figures 7*b*, 8, and 10*b*), it would be a very poor recorder of the Earth's magnetic field. *Kent* [1973] was fortunate to find an alternative mechanism of magnetization which was not only more appropriate to many sediments investigated by paleomagnetists, but which behaved in a much more tractable fashion.

Kent [1973] suggested that the magnetization acquired when a water-saturated sediment is stirred in the presence of a field was a useful laboratory analogue to the magnetization acquired as a result of bioturbation. His results are shown in Figure 11*a*. He redeposited deep sea sediments as concentrated slurries and stirred them in the presence of fields ranging up to 120 μT. The stirred magnetization was somewhat stronger than the original depositional remanence and had no inclination error. More important, these stirred remanences were a linear function of the applied field. *Barton et al.* [1980] found linear behavior in their stirred remanences up to field strengths of 900 μT.

Verosub et al. [1979] gave a stirred remanence to several types of sediment. He then allowed the sediments to dry, turning the samples through 90° at various stages during the drying process. He found that the stirred remanence was quite stable against subsequent field changes.

Tucker [1980*b*] noted the similarity of the experiments of *Kent* [1973], *Verosub et al.* [1979], and *Games* [1977]. Games magnetized mud by slinging it into a box and allowing it to dry in the sun in order to simulate the remanence acquired during the making of adobe bricks. In each of these experiments the magnetization was acquired during the disturbance and was resistant to further change. *Tucker* [1980*b*] devised a series of experiments to investigate the behavior of stirred remanence. As before, he used a gradiometer to measure magnetization during and after stirring as a function of stirring rate, applied magnetic field, and water content. Ultimately, the sediments were dried, subsampled, and measured on a spinner.

Tucker's [1980*b*] principal results may be summarized as follows: (1) For low fields and low stirring rates the remanence was acquired during stirring, and there was only a slight increase after stirring ceased. The stirred remanence was linearly related to the applied field and was an order of magnitude higher than that acquired with no stirring (data shown in Figure 11*b*). The nonstirred remanences, however, were more stable against alternating field demagnetization. (2) High stirring rates resulted in lower intensity during stirring (synstirring magnetization) and a huge increase after stopping. The behavior versus the applied field was nonlinear for moderate fields. (3) High fields produced both high synstirring and poststirring remanences. (4) The poststirring remanences were stable over days; if put in an opposite field, a reduction in intensity of only 10% was observed. (5) Lower water contents reduced remanence acquisition.

Tucker [1980*b*] modeled his results as follows: He started with an equation similar to (1), but with the addition of the contribution of the stirring $f(\psi)$:

$$f(\psi) + I\ddot{\theta} + \lambda\dot{\theta} + mB\sin\theta = 0$$

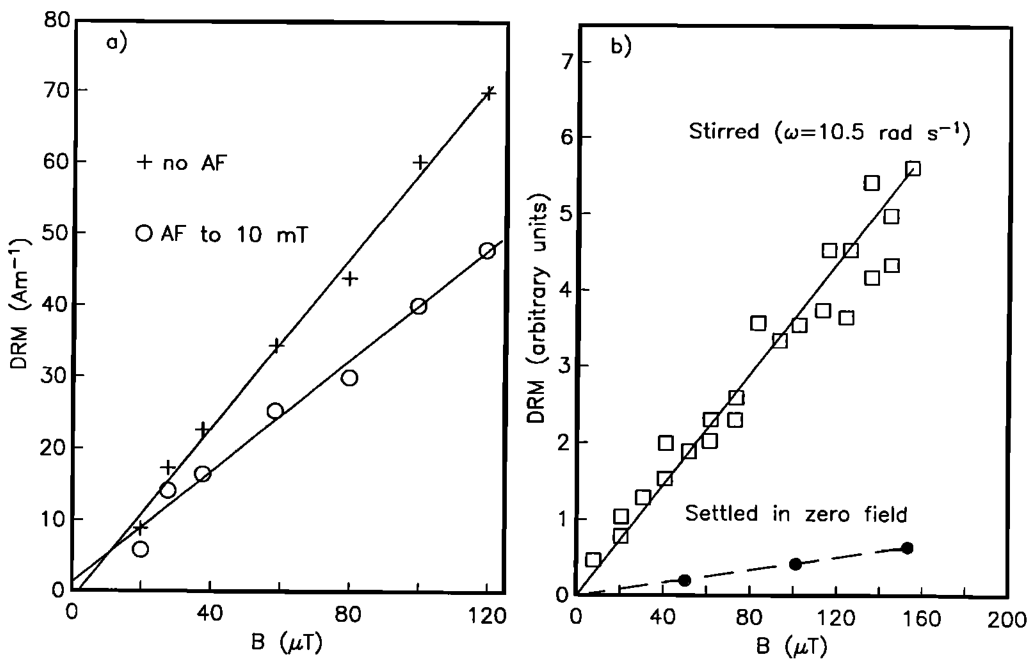

Figure 11. (*a*) Deep sea sediment redeposited in the laboratory. Thick slurry was stirred in the presence of field B. Redrawn from *Kent* [1973]. (*b*) Comparison of remanence acquired by stirring in an applied field (squares) versus that acquired when settled in zero field and then exposed to field B (solid circles). Redrawn from *Tucker* [1980*b*].

He assumed that stirring periodically randomizes the grains. These may then realign and get fixed after some time τ, and in this case, $f(\psi)$ can be neglected. The characteristic time τ is some function of mB/λ and stirring rate ω. After some manipulation, he came up with a model for the various magnetizations, which was in excellent agreement with his data. Using a parameter F that is related to the fraction of alignment,

$$F(\theta_o, k, t) = \left[\frac{1 - \tan(\theta_o/2)e^{-2kt}}{1 + \tan(\theta_o/2)e^{-2kt}}\right]$$

the synstirring magnetization (M_{syn}) can be modeled as

$$M_{\text{syn}} = m \sum_{\theta_o} \left(\sum_{t=0}^{\tau} F(\theta_o, k, t) + \sum_{t=\tau}^{\pi/\omega} F(\theta_o, k, t) \right)$$

for $\tau < \pi/\omega$, and

$$M_{\text{syn}} = m \sum_{\theta_o} \sum_{t=0}^{\pi/\omega} F(\theta_o, k, t)$$

for $\tau > \pi/\omega$.

The poststirring magnetization M_{post} can be described by

$$M_{\text{post}} = m \sum_{\theta_0} F(\theta_0, k, t)$$

Tucker's model relies heavily on the following two experimentally determined facts: (1) Constraining forces must be eliminated in order for grains to realign themselves. This is achieved by stirring, for example. (2) A characteristic time τ after stirring ceases, the constraining forces again become important, and the particle is immobilized, fixing the magnetization.

6. OTHER CONTROLS OF SEDIMENTARY INTENSITY

It is worthwhile considering what other factors might be important for determining the intensity of sediments. One particularly interesting paper was published by *Lu et al.* [1990]. They measured remanence in synthetic samples as a function of clay content and conductivity (salt content). They mixed a saturated slurry of water and clay in an electric blender and poured it into 1-inch-cylinder cups. These samples were then dried (resulting in a huge volume loss). The effect of drying on remanence was neglected.

Their results were as follows: (1) There was a strong effect of the conductivity of the solution on the magnetizations of kaolin + magnetite + silica slurries, whereby added salt decreased the DRM. No effect was observed for slurries in which the clay was montmorillonite. (2) The magnetization decreased and the inclination error increased with addition of kaolin, whereas χ and ARM were all constant with conductivity and kaolin changes. (3) Finally, the longer the slurry was allowed to "sit" before the experiment, the lower the DRM. This latter effect was held to support coagulation as the controlling factor.

These results point to the role of gels, coagulation, or flocculation as being of critical importance to the blocking of magnetic remanence. Coagulation may be encouraged as compaction occurs and may contribute to the locking in of DRM. When a later disturbance occurs, these bonds are broken momentarily as described by *Tucker* [1980*b*], allowing realignment; then coagulation again occurs, fixing the remanence.

Let us now consider what might happen to remanent intensity during burial. The work of *Anson and Kodama* [1987] primarily focused on the effect of compaction on inclination, finding that the inclination error increased with increasing compaction. They mentioned in passing that there is a 20–30% decrease in intensity, suggesting that this is due to increased particle interaction. They presented intensity data in a table, and I have plotted these in Figure 12. In Figure 12*a*, I show magnetization versus volume change (ΔV). There is an excellent linear relationship as shown by the line with an inferred initial remanence (M_o and not measured) of some 44×10^{-7} A m^2. These data are typical of the entire data set, and in Figure 12*b*, I plot all the data of *Anson and Kodama* [1987] for acicular magnetite, normalized to M_o and averaged over 0.05-wide bins of ΔV. In Figure 12*c*, I plot the slope of the regression line (an estimate of the severity of the effect) against M_o (an estimate of magnetite concentration). The slopes of the regression lines are remarkably similar and completely independent of initial magnetization; hence the reduction in intensity is probably not controlled by increased particle interaction, as this would certainly be higher for higher initial concentrations. It is more likely due to random rotations about a horizontal axis, leading to reduced inclination and reduced intensity. Since the slope is more or less uniform, it is possible to hope that even if this occurs in nature in sediments otherwise suitable for paleointensity measurements, it occurs uniformly and will not severely affect relative paleointensity measurements, except perhaps for altering the long-term trend of the data.

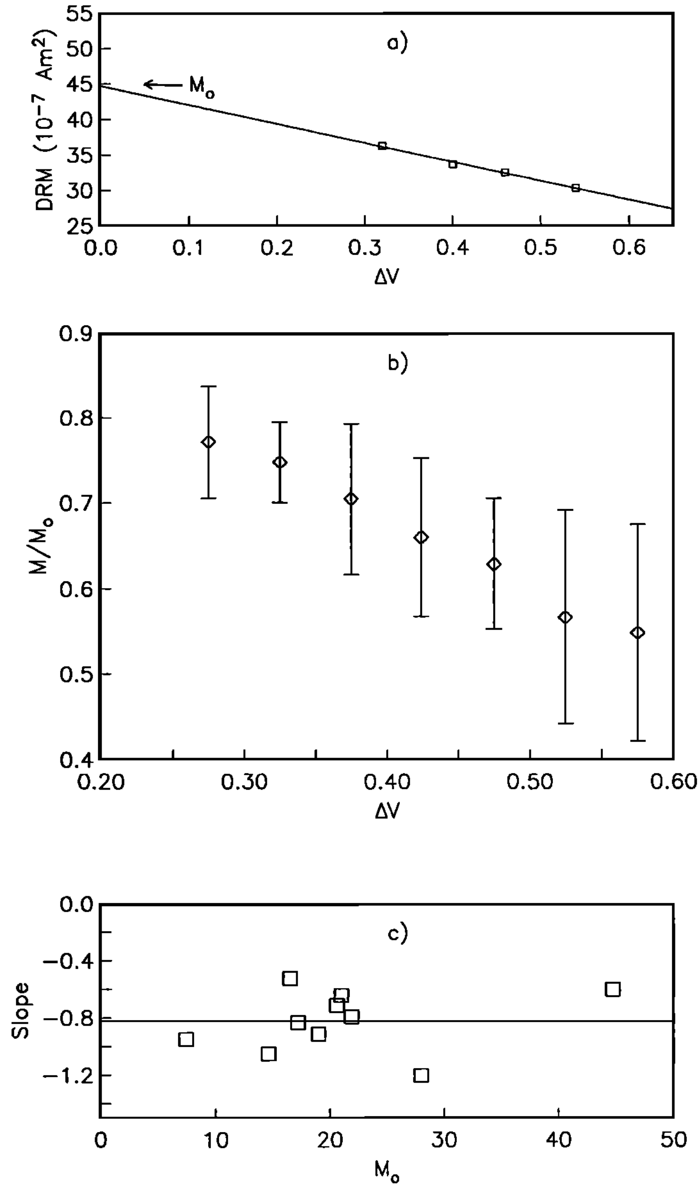

Figure 12. Data from *Anson and Kodama* [1987]. (*a*) Magnetization versus degree of compaction as represented by volume change (ΔV). M_0 is the *y* intercept of a regression line through the measured data points (squares). (*b*) Same as Figure 12*a*, but all acicular data are normalized by M_0 and averaged over 0.05-wide bins of ΔV. (*c*) Slopes of all regression lines (see Figure 12*a*) versus M_0.

7. NORMALIZATION

In the foregoing it was established that under certain conditions likely to occur in nature, there is a linear relationship between field and magnetization in sediments. However, the magnetization must also be related to the quantity and type of magnetizable material within the sediment, and if these factors vary within a sedimentary sequence then intensity of magnetization alone will not provide a reliable estimate of relative paleointensity. The proper parameter for normalization is still a matter of some debate, and I will guide the reader through the efforts at establishing a firm theoretical and experimental basis for choosing a normalizer. Normalized remanence yields an estimate of the relative paleointensity of the geomagnetic field. This estimate of the relative intensity is here denoted B^*. The normalizer will be indicated as a subscript where appropriate. For example, NRM/ARM is B^*_A, NRM/IRM is B^*_I and NRM/χ is B^*_χ.

The earliest attempt at using sediments for paleointensity was by *Johnson et al.* [1948]. They sought to normalize their remanence and chose saturation isothermal remanent magnetization (M_{sr} in Table 2) to account for variability in the "magnetization potential" of the sediments. They also attempted to calibrate their relative estimates into absolute estimates through laboratory redeposition. It is interesting to note that they concluded that the intensity (and direction for that matter) of the field had not changed much over the last million years; this was taken as support for Blackett's "fundamental theory" of magnetization that a dipolar field was inherent to rotating bodies.

Harrison [1966] also realized that magnetization must be adjusted for changes in concentration of magnetic phases and divided the average magnetization for each of his deep sea sediment cores NRM by the average magnetic susceptibility $\bar{\chi}$. He stated (p. 3041, notation altered to conform to usage in this paper),

> If the dipole model of the earth's magnetic field is correct, I would expect the value of the ratio of the intensity of NRM to susceptibility to show a variation with latitude [λ] of the form
>
> $$\bar{NRM}/\bar{\chi} \propto (1 + 3 \sin^2 \lambda)^{1/2}$$
>
> owing to the variation of the strength of the earth's magnetic field. No such variation is apparent.

Of course, we now realize that such a normalization would fail to take into account differences in matrix, grain size, mineralogy, and depositional environment, not to mention changes in the field itself, and it is not surprising that such a relationship was not observed.

Most of the early attempts at using sediments for relative paleointensity were associated with records of reversals of Earth's magnetic field. *Harrison and Somayajulu* [1966] were among the first to study reversals in deep sea sediments and used χ as a normalizer. They attempted to show that the normalized intensity was independent of lithologic variations by calculating correlation coefficients among NRM, χ, and NRM/χ. This approach foreshadows more sophisticated methods of testing correlation in the frequency domain by *Tauxe and Wu* [1990]. *Nesbitt* [1966] also used χ as a normalizer, but on red sandstones which cannot be considered as reliable records [see *Tauxe and Kent*, 1984]. *Opdyke* [1972] and *Opdyke et al.* [1973] advocated the use of IRM or ARM as a normalizer.

The dangers of failing to consider carefully aspects of normalization are illustrated by the work of Wollin and colleagues [e.g., *Wollin et al.*, 1971]. They proposed a correlation between climatic changes and variations in magnetic inclination and in magnetic intensity in deep sea sediment cores. They claimed that the NRM directly reflected changes in magnetic field strength as follows [*Wollin et al.*, 1971, p. 181]: "Since the intensity measurements in [the] cores ... are per gram of sediment and not per gram of magnetizable material in the samples, some of the variability in the intensity measurements in the cores may be due to differences in composition and abundances of magnetic minerals in the sediment samples." Instead of following the lead of *Harrison* [1966] and showing that there is no relationship between their field estimate, NRM, and for example χ, they considered its relationship to less direct measures of magnetizability such as grain size. The correlation of the NRM to carbonate and other climatic indicators served as the basis for their claim that the Earth's field is somehow linked to climate. This claim was debunked by the work of *Amerigian* [1974], *Chave and Denham* [1979], and *Kent* [1982], who demonstrated that NRM was profoundly influenced by concentration of magnetic material as measured by ARM, IRM, and χ.

The contenders in the choice of a parameter reflecting magnetic activity have been χ, IRM, ARM, and stirred remanence (StRM) (see Table 2). Other possibilities will be discussed later. Several key papers provide the foundation for a rational choice. Advocates for ARM include *Johnson et al.* [1975*b*], *Levi and Banerjee* [1976], and *King et al.* [1983]. The case for stirred remanence was laid out by *Kent* [1973] and *Tucker* [1980*b*, 1981], and arguments in favor of IRM have been made by *Hartl et al.* [1993].

Levi and Banerjee [1976] laid out the assumptions for paleointensity in sediment as follows: (1) the magnetization must be linearly related to applied field, (2) variations in pressure, temperature and water content are either small or unimportant, and (3) judicious selection of normalizing procedure ("It is also required that the particular normalizing procedure chosen for the determination of the relative paleointensity activate the same relative spectrum of magnetic particles which are also responsible for the NRM." [*Levi and Banerjee*, 1976, p. 221]). They pointed out that susceptibility and saturation magnetization are measured in the presence of a field and hence are hard to relate to remanence which is measured in zero field. Furthermore, these parameters are disproportionately influenced by superparamagnetic and multidomain grains which play no role in the stable remanence. Thus they argued that some form of room temperature remanence (such as viscous, isothermal, anhysteretic, and so on) would be preferable. *Levi and Banerjee* [1976, p. 221] wrote,

> One criterion for choosing the normalization parameter is to use the remanence whose demagnetization curve most closely approximates that of the NRM because that process magnetizes particles with a stability (or coercivity) spectrum which most closely resembles the NRM.

Johnson et al. [1975*a*] had advocated the use of anhysteretic remanence because it is particularly sensitive to the single-domain particles. Part of the reason for the preference of anhysteretic remanence is the argument that the detrital magnetic grains in a sediment presumably acquired their moments as a thermal remanence. Since anhysteretic remanence is very similar to thermal remanence [see *Gillingham and Stacey*, 1971], ARM is a logical choice to normalize what amounts to a redeposited thermal remanence. *Johnson et al.* [1975*b*], however, cautioned that detrital remanence is in fact quite different from thermal remanence and care should be exercised in carrying the analogy too far.

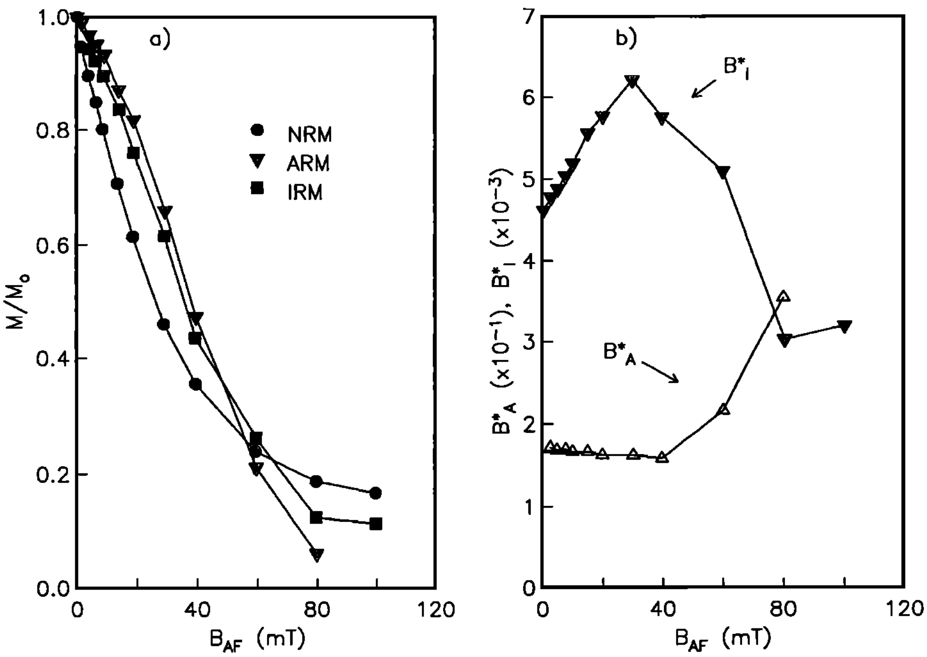

Figure 13. Replotting of figure from *Levi and Banerjee* [1976], showing (*a*) magnetization versus demagnetizing field and (*b*) normalized remanence versus demagnetizing field.

Levi and Banerjee [1976] amplified the argument for using ARM by noting that (at least in their cores) the coercivity spectrum of the anhysteretic remanence was similar to that of the natural remanence and also both of these were not very much like that of the IRM. A consequence of this dissimilarity in coercivity spectra between IRM and NRM is that the ratio of the two is very unstable during demagnetization. In Figure 13, I have replotted data from *Levi and Banerjee* [1976]. Figure 13*a* shows the intensity decay of the various remanences during alternating field demagnetization. Figure 13*b* shows the ratios of the various remanences during demagnetization. The ratio of natural to anhysteretic remanence B_A^* is quite stable over a broad range of demagnetization steps, whereas B_I^* is unstable. This, combined with the argument that the magnitude of ARM is a few tens to hundreds of times that of NRM, whereas IRM is typically thousands of times larger, is the basis for preferring B_A^*. *Levi and Banerjee* [1976] favored using natural remanence normalized by anhysteretic remanence, demagnetized in the same way as for the relative paleointensity study. *Denham* [1976] went a step further and advocated using vector differences between two demagnetization steps (the portion of each remanence isolated between two demagnetization steps).

In his paper on postdepositional remanence in deep sea sediments, *Kent* [1973] suggested that it might be possible to use a laboratory-stirred remanence as a proxy for magnetic activity. *Tucker* [1980*b*] made a very strong experimental and theoretical case for such a normalization and then attempted to apply it to deep sea sediments [*Tucker*, 1981]. The problem with his application was that he chose to use unbioturbated turbidites whose magnetizations may not have been well duplicated by the stirring experiments.

Barton and McElhinny [1979] tried laboratory redeposition (with no postdepositional disturbance) in order to calibrate their relative intensity studies, but as was shown later [*Barton et al.*, 1980; *Tucker*, 1980*a*] this sort of deposition is not necessarily linear with the field and cannot be considered a reliable record of even relative paleointensity. *Thouveny* [1987] compared remanences normalized by StRM with other types of normalizers in his sediments from Lac du Bouchet and found good agreement among the various methods.

There have been hints at other possible laboratory analogues for normalization [see *Yoshida and Katsura*, 1985], but laboratory analogues have never gained wide acceptance in the paleomagnetic community. Reasons for this ambivalence are many. The experiments are time consuming, they require a large amount of material, and they are destructive of the sedimentary fabric. Furthermore, there is a great deal of skepticism among paleomagnetists that it could ever lead to true "absolute" paleointensity estimates. Nonetheless, it is the opinion of this author that the case was very strongly made by *Tucker* [1980*b*] and that it has never been systematically and properly tested. At the very least, StRM or M_e may prove to be superior normalizers for relative paleointensity information to any of the others in common use.

By the early 1980s, therefore, there was a strong preference in paleomagnetic circles for using ARM as an estimate of magnetic activity. The sense of the community was expressed by *Johnson et al.* [1975*b*, p. 415], who stated,

> An accurate measure of relative intensity can only be obtained for a given section of a core if (1) there are no significant trends in chemical effects and magnetic stability down the section, (2) there has not been substantial mixing of the

sediment in the section, and (3) there are no significant overlaps of normal and reversed polarities in a sample. If these conditions are ever demonstrated to exist, then ARM acquisition should be a useful method for obtaining relative intensities.

Against this backdrop came a series of papers culminating in a set of criteria for reliability of sediments for relative paleointensity measurements [*Banerjee et al.*, 1981; *King et al.*, 1982, 1983]. These have been the starting point for most recent paleointensity work and I will now consider them in some detail, with a particular eye to examining the foundations for the criteria and modifying them where appropriate.

Drawing on the work of *Dankers* [1978], among others, that determined basic rock magnetic parameters for carefully sized magnetites, *Banerjee et al.* [1981] exploited the fact that these parameters show a smooth variation with grain size. In particular, they noted that ARM is enhanced in fine grains and χ in coarse ones. Thus the ratio of the two could be used as a crude guide to the grain size of the magnetic fraction. They introduced the idea of plotting ARM against χ as a quick look at relative grain size variations, a style of plotting I will call the "Banerjee plot." Rock magnetic parameters such as the ratios of M_{sr}/M_s and B_{cr}/B_c (coercivity of remanence and coercivity respectively), plotted in what I will call "Day plots," were already in use for determining grain size [*Day et al.*, 1977], but these parameters were difficult and time consuming to determine. In contrast, ARM and χ were readily measurable without the need for magnetic separation. I note that now, with the advent of the alternating gradient force magnetometer, hysteresis parameters can be measured in minutes on very small bulk samples.

The most recent version of the Banerjee plot relating anhysteretic magnetization and susceptibility [*King et al.*, 1983] is shown in Figure 14. Anhysteretic magnetization is calculated as "anhysteretic susceptibility" (χ_A) or the anhysteretic magnetization acquired per unit bias field. In this empirical model the ratio χ_A/χ is related to grain size, and the distance from the origin is related to concentration. *King et al.* [1983] cautioned the reader against a too literal interpretation of their "phenomenological model," but the temptation has been great.

The most important contribution of the *King et al.* [1983] article was that they critically evaluated the hypothesis that normalized remanence (in particular, B_A^*) is a measure of relative paleointensity in two ways: (1) they looked at the rock magnetic assumptions involved in the normalization method, and (2) they compared B_A^* data with contemporaneous data obtained from thermal remanence (an entirely different recording medium).

As a motivation for their discussion of the rock magnetic basis for relative paleointensity, they summarized the requirements for the method as (1) uniform mechanism of magnetization, (2) uniform magnetic mineralogy, and (3) uniform grain size. The question they sought to answer was how to define and detect "uniformity." First, since most rock magnetic studies have been carried out on magnetite, they restricted the mineralogy of sediments to magnetite.

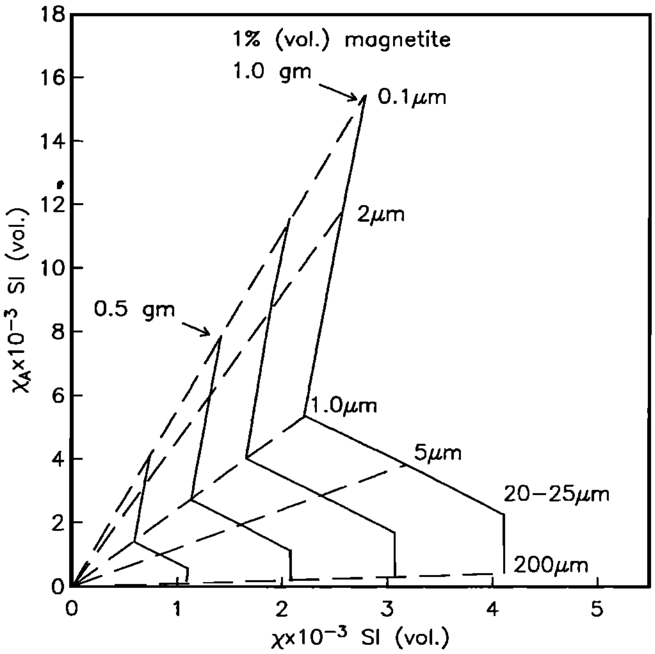

Figure 14. "Phenomenological model" of *King et al.* [1983] relating anhysteretic susceptibility (χ_A) and χ with magnetic grain size and concentration.

Next, they drew on the work of *Amerigian* [1977], who determined a rough grain size dependence of DRM and ARM. *King et al.* [1983] pointed out that you can only use sediments whose grain size variations do not affect the ratio. DRM is inefficient at low grain size because of Brownian motion and at large grain size presumably because the particles get blocked faster. The peak in efficiency is in the 2 μm range. ARM, on the other hand, is efficient at both small and large grain sizes with a low in the middle (troughing at about 2 μm). In order for the normalization to be relatively insensitive to small changes in grain size, the slopes of each parameter must be relatively constant (certainly not change sign, as at about 2 μm); hence the grain size must be larger than a few microns. Furthermore, *King et al.* [1983, p. 5912] stated that "... magnetite larger that ~15 μm is not usually suitable for paleomagnetic studies because it is magnetically too 'soft' ... to retain a stable primary remanence." Thus, *King et al.* [1983] placed size limits of 1–15 μm on the magnetite for "uniformity" with respect to B_A^*. They recommended the use of the Banerjee plots to test this, but of course Day plots could also be used effectively.

Finally, *King et al.* [1983] cited the work of *Sugiura* [1979] (shown in Figure 15*a*), to show that for low particle interaction, the ARM acquisition curve is dis-

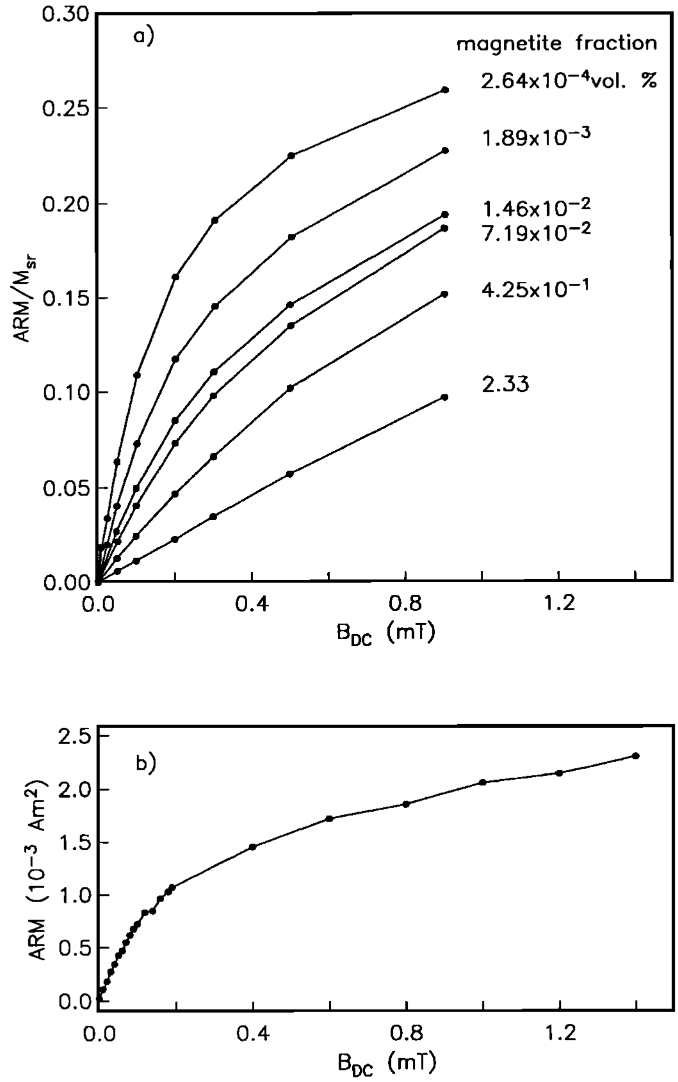

Figure 15. (*a*) Anhysteretic remanence normalized by saturation remanence versus the bias field applied during alternating field demagnetization B_{DC}. Redrawn from *Sugiura* [1979]. (*b*) Anhysteretic remanence versus B_{DC} for marine carbonates from the Ontong-Java Plateau.

tinctly nonlinear and the curvature is strongly affected by concentration. If concentration varies, then the ARM acquisition curve will be different. On the basis of these data, *King et al.* [1983] placed a limit on the amount that the concentration may vary (no more than a factor of 20–30) for the sediments to be uniform in B_A^*. Variations in concentration can also be "guesstimated" from the Banerjee plot. A factor of 30 change in concentration, however, results in a difference of some 10% in the ARM acquired in a given field. This will generate considerable scatter in the data and is perhaps too liberal a limit. I would prefer a limit of more like a factor of 10 or less for changes in concentration, when using ARM as a normalizer.

King et al. [1983] tested their criteria using sediments from LeBoeuf Lake (in Pennsylvania). These were found to satisfy the criteria outlined above (that the mineral be magnetite with grain sizes between 1 and 15 μm and that concentration not vary by more than a factor of 20 or 30). Their B_A^* data were compared with the contemporaneous absolute paleointensity data of *Champion* [1980] and were said to agree. *King et al.* [1983] stressed the importance of the cross-check for two reasons: (1) B_A^* does not have the internal checks that the thermal remanence allows and (2) it is really only a relative tool and requires calibration to absolute scale.

Hilton [1986] considered the basis for the Banerjee plot and showed that the ratio of two bulk parameters is not independent of concentration if more than one magnetic mineral is present. If the number of minerals is unknown, no information on domain size changes and mineralogy can be deduced. Of course, for the purposes of the present discussion it is essential that there be a single magnetic mineral present and, moreover, that that mineral be magnetite, so the comments of *Hilton* [1986] per se are irrelevant. However, there is one fundamental problem with the Banerjee plot that is important to discussions of paleointensity in sediments. That is the problem pointed out by *Sugiura* [1979] and reiterated by *King et al.* [1983] that the acquisition of ARM is heavily dependent on concentration.

Figure 15*a* shows the behavior of anhysteretic remanence as a function of dc bias field and concentration. The diagram shown in Figure 14 was calibrated using synthetic samples with magnetite concentrations of the order of 1% by volume, or close to the bottommost curve. In this concentration, ARM is linear with bias fields up to at least 1.2 mT. Most sediments, however, have magnetite concentrations measured in parts per million, or near the upper two curves. In this region, ARM is strongly nonlinear, even in very low fields. To highlight this fact, I plot in Figure 15*b* ARM acquisition curves for a marine carbonate sample from the Ontong-Java Plateau, known to satisfy the "King criteria" [see *Constable and Tauxe*, 1987; *Tauxe and Wu*, 1990]. The nonlinearity is striking even for fields as low as 0.1 mT (the bias field recommended by *King et al.* [1982]). The curvature is consistent with concentrations in the range of a few times 10^{-4} percent, or the upper curve of Figure 15*a*.

The first problem that the concentration dependence of ARM brings up is that the analogy between anhysteretic magnetization and susceptibility (which is defined as the slope between induced magnetization and the applied field) implied by the term "anhysteretic susceptibility" is strained. The acquisition is not linear even in quite low fields. It is common to assume a linear relation and combine data derived in different bias fields. Use of anhysteretic susceptibility seems to obviate the need for even specifying the actual bias field used.

A second, and more serious problem is that at the low concentrations typical for sediments, ARM is

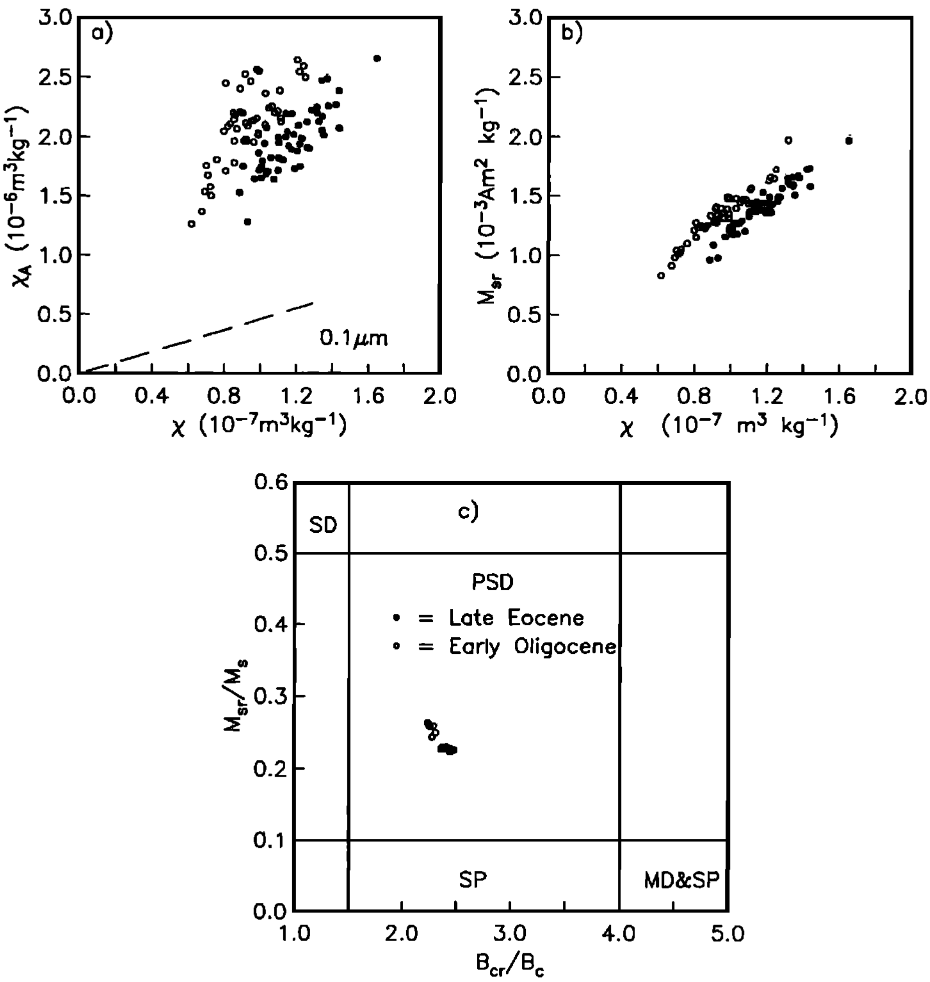

Figure 16. (*a*) Banerjee plot of data from marine carbonates from Site 522 (P. Hartl, unpublished data, 1992) and grain size predicted from *King et al.* [1983] (see Figure 14). Solid symbols are from the late Eocene and open symbols are early Oligocene. (*b*) Plot of M_{sr} versus χ for same samples as in Figure 16*a*. (*c*) Day plot [*Day et al.*, 1977] of hysteresis parameters for samples from two tracks shown in Figure 16*b*.

nearly twice as efficient as for high concentrations, and the relationship of ARM to χ will be far from linear with concentration. Thus the lines radiating from the origin in Figure 14 must in fact be curves. Furthermore, a Banerjee plot of data derived from a sequence of marine carbonates (P. Hartl, unpublished data, 1992) is shown in Figure 16*a*. Day plots for these sediments (Figure 16*c*) indicate a pseudosingle domain grain size (i.e., 1–15 μm), yet they plot well above even the line for 0.1 μm in Figure 14. One possibility for the discrepancy is a difference in converting from cgs units into SI units. There are several errors in the table of *King et al.* [1983] listing the conversions. I checked their calculations, however, and they apparently employed the correct conversion in constructing Figure 14, so the explanation probably lies in the difference in concentrations as previously described.

Finally, the data in Figure 16*a* are somewhat scattered. In Figure 16*b* I show M_{sr} versus χ for the same samples. These data fall along two clearly defined tracks, one made up of samples from the lower Oligocene in the core, and the other from samples in the Eocene. A Day plot (Figure 16*c*) for specimens from the upper track and lower tracks clearly shows a very slight difference in grain size. These trends, seen in the M_{sr} versus χ data, are much less clear in the plots of ARM versus χ. The scatter in ARM is not instrumental noise but could relate to the nonlinearity of ARM depending on concentration. Because the M_{sr} used here is a saturation parameter, it is relatively insensitive to changes in the inducing field and is a simple function of concentration in these low-concentration sediments. Of course, in high concentrations, M_{sr} will be strongly affected by particle interaction as well [see *Hamano*, 1983].

It is also undeniable, however, that M_{sr} is biased by large grain sizes and is very much stronger than the original remanence and that the coercivity spectrum is usually somewhat different. For these reasons we seek a broader palette of parameters to choose from. *Yoshida and Katsura* [1985] suggested their method of examining the magnetization of resuspended sediments as a measure of concentration of magnetic phases. This may be a very promising approach. As already stated, stirred remanence has been largely overlooked. Perhaps saturation ARM would provide a more stable estimate. Lastly, low field IRM, when plotted against the log of inducing field is often linear, and the ratio of these might provide an entirely new and different parameter to normalize remanence. These are but a few suggestions for further investigation, and there are surely other appropriate methods of normalization.

Perhaps the best approach is to compare different methods of normalization, and where agreement is found, investigators can take heart [e.g., *Thouveny*, 1987; *Tric et al.*, 1992; *Hartl et al.*, 1993]. In cases of poor agreement, records should be viewed with extreme caution.

8. SOME REALITY CHECKS

Lacking a firm theoretical basis for sedimentary paleointensity, perhaps the strongest arguments that can be made for reliability of records are statistical in nature. First, as was pointed out by *Harrison and Somayajulu* [1966] and *Kent* [1982], the paleointensity estimate should not be correlated with bulk rock magnetic parameters. Second, when placed on a common timescale, multiple records from a given region should be quite similar.

One straightforward means of testing if two parameters are correlated is to calculate a linear regression and a correlation coefficient. This was done, for example, by *Harrison and Somayajulu* [1966], who showed that while remanence was correlated with susceptibility, their normalized remanence was not. However, because data in a time series are inherently correlated and not independent, calculation of a correlation coefficient is inappropriate; hence this method is rather limited. A more suitable approach is to seek correlations within particular frequency bands using coherency testing. The benefits are enormous. For example, there may be a subtle control of a rock magnetic nature at one frequency band (say, with a period of 20 ka) which could well be obscured by calculating a simple correlation coefficient.

The generalization of the correlation coefficient to the frequency domain is the coherence function spectrum. *Tauxe and Wu* [1990] illustrated the power of the technique for paleointensity studies. They used it in two ways. First, if paleointensity records are strongly affected by such climatically controlled factors as changes in grain size, then they will be coherent with some rock magnetic parameter reflecting grain size (say χ), over some range of frequencies. If the two are not significantly coherent, then one may feel confident that the relative paleointensity estimate is not primarily controlled by lithologic factors. Furthermore, choice of the coherence function over the correlation coefficient has one great advantage: The range of frequencies for which the data contain paleointensity information (as opposed to noise or lithological information) is readily apparent from the plots of squared coherence between two parameters. Moreover, tests of significance are available. Second, the coherence function can be used to compare two paleointensity records. When placed on a common timescale, the records should be coherent, if they both reflect changes in the geomagnetic field.

The technique was described by *Tauxe and Wu* [1990], and the approach is outlined here. Squared coherence, γ^2, is evaluated for a specific frequency, f, by comparing the cross-spectral density function for two data series x and $y(S_{xy}(f))$ with the autospectral density functions of each series, S_{xx} and S_{yy}, respectively. Thus the squared coherence function is defined as

$$\gamma^2_{xy} = \frac{|S_{xy}(f)|^2}{|S_{xx}(f)||S_{yy}(f)|}$$

and lies between zero (no coherence) and one (complete coherence). A zero coherence test for the above function is given by the relation [*Chave and Filloux*, 1985]

$$c^2 = 1 - (1 - \alpha)2/(\nu - 2)$$

where $\alpha = 0.95$ for the 95% level of confidence and ν is the equivalent degrees of freedom (here, ν is 24). Thus if γ^2 exceeds about 0.24, the squared coherence of the two time series is significantly different from zero at the 95% level of confidence.

Although coherence can be calculated from a variety of methods for estimating spectral density, *Tauxe and Wu* [1990] advocated the use of multitaper techniques in which the data are multiplied in turn by a set of tapers which are designed to maximize resolution and minimize bias [*Thomson*, 1982]. In addition to minimizing the bias while maintaining a given resolution, the multitaper approach allows an estimate of the statistical significance of certain features (such as spectral lines) in the power spectrum by comparing the character of the spectral density estimates calculated for different data windows. These techniques are now in routine use in a variety of applications in earth science (see *Park et al.* [1987], for example).

I show examples of the use of the coherence function in Figure 17. In Figures 17*a*–17*c* I plot γ^2 for the paleointensity estimators B^*_χ, B^*_I, B^*_χ respectively versus the normalizer for three cores taken from the Ontong-Java Plateau. In the case of ERDC 113p, *Tauxe and Wu* [1990] decided that for frequencies lower than about 0.06 (or periods longer than about 17 ka) the relative paleointensity record was uncontaminated by lithological control. The record of ERDC124p is clearly affected by lithological variations in the fre-

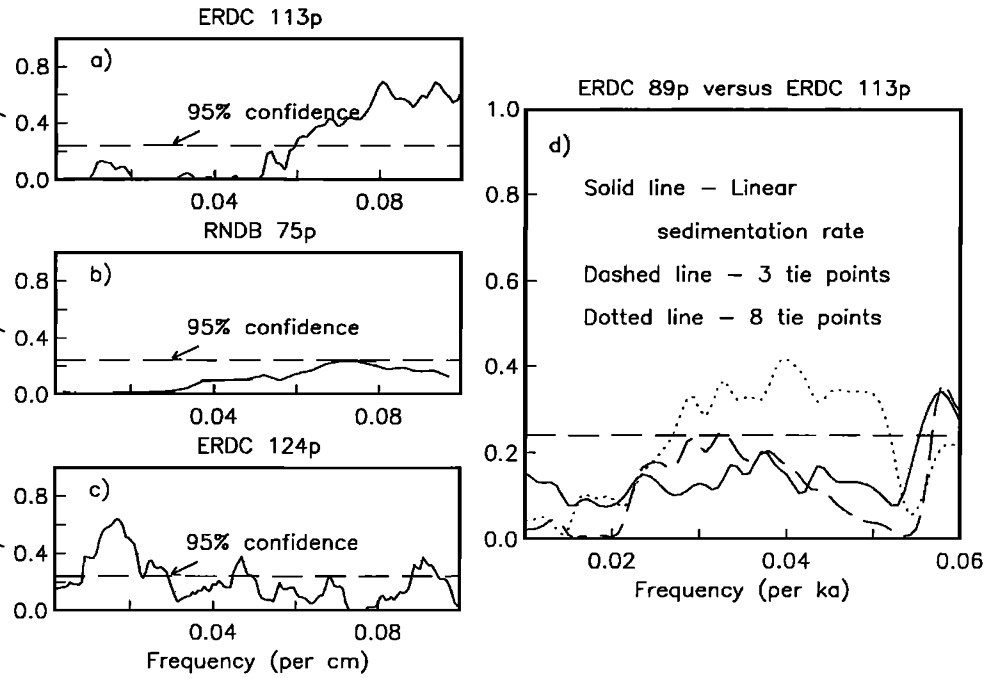

Figure 17. Plots of coherence versus frequency. (*a*)–(*c*) Coherence between B^* and $[m_a]$, where $[m_a]$ was χ in Figures 17*a* and 17*c* and M_{sr} in Figure 17*b*. (*d*) Coherence between two B^*_χ records as a function of increasingly accurate time control.

quencies of interest, and this record should never be used for paleointensity studies. The record of RNDB 75p appears to be satisfactory and will be discussed later. In Figure 17*d*, I show the other use of the coherence function, to test the correlation of two independent paleointensity records. In this case, I have used the data of *Tauxe and Wu* [1990] for cores ERDC 113p and 89p. When put on a common time scale using oxygen isotopes, the two records become significantly coherent.

9. SUMMARY OF DESIRABLE ATTRIBUTES FOR SEDIMENTARY PALEOINTENSITY RECORDS

I have reviewed the experimental and theoretical considerations for assessing the reliability of paleointensity data derived from sediments and am now in a position to establish criteria by which such data can be judged. These are summarized below.

1. The natural remanence must be carried by stable magnetite, preferably in the grain size range of about 1–15 μm. Furthermore, the portion of the natural remanent vector used for paleointensity should be a single, well-defined component of magnetization.

2. The detrital remanence must be an excellent recorder of the geomagnetic field and exhibit no inclination error. Although some unbioturbated sediments do seem to record faithfully the inclination of the field, particularly when the particle flux is low during deposition, there is to date no experimental evidence that the remanent intensity would be a simple linear function of field intensity. Thus bioturbated sediments are preferable to laminated sediments for relative paleointensity studies.

3. Concentration variations of more than about an order of magnitude should be avoided.

4. Normalization is preferably done by several methods, all yielding consistent results.

5. The relative paleointensity estimates that are coherent with bulk rock magnetic parameters should be treated with caution.

6. Records from a given region should agree within the limits of a common timescale.

10. RELATIVE PALEOINTENSITY RECORDS

Considering the level of interest in secular variation of the magnetic field, there are not very many papers concerned with intensity variations (see Table 4). More important, few consider very seriously the above listed suggestions for reliable data. Nonetheless, I think it appropriate in a review article such as this to examine the available records and draw preliminary general conclusions about their meaning. In Table 4, I have listed the records available to me, and the locations of the studies are plotted in Figure 18. I will discuss them in order of increasing time span, starting with the highest resolution, shortest duration records and finishing with records that span several million years. I consider records of relative paleointensity during the Brunhes/Matuyama (B/M) reversal separately. I have taken a conservative approach to data

TABLE 4. Paleointensity Records

Record	*Reference*	*Latitude, °N*	*Longitude, °E*	$[m_a]$
RC10-167	1	33	150	ARM
LeBoeuf Lake	2	~42	~−80	ARM
Lake Eacham	3	−17	175	ARM
Lake Barrine	3	−17	175	ARM
ERDC 83Bx	4	1	157	ARM
ERDC 102Bx	4	−4	161	ARM
Lac du Bouchet	5	45	4	StRM, ARM
Lake Mbo	6	5	10	ARM
INMD 48Bx	7	30	−43	ARM
ERDC 113p	8	2	159	χ
ERDC 89p	8	0	156	χ
KET82629	9	39	14	χ
MD84629	9	36	33	χ
DED8707	9	40	13	χ
DED8708	9	40	13	χ
RNDB75p	10	2	160	IRM
ODP 130-803A	11	2	160	χ
DSDP 73-522	12	−26	−5	IRM, ARM, χ
MD85-668	13	−1	46	ARM
MD85-669	13	2	47	ARM
MD85-674	13	3	50	ARM
ODP 767B	14	5	123	χ
ODP 769A	14	9	121	ARM
V16-58	15	−4	30	ARM
ODP 609B	16	50	336	ARM
ODP 664D	17	0	−23	ARM
ODP 665A	17	3	−20	ARM

References are as follows: 1, *Kent and Opdyke* [1977]; 2, *King et al.* [1983]; 3, *Constable* [1985]; 4, *Constable and Tauxe* [1987]; 5, *Thouveny* [1987], *Thouveny et al.* [1990], and *Creer et al.* [1990]; 6, *Thouveny and Williamson* [1988]; 7, *Tauxe and Valet* [1989]; 8, *Tauxe and Wu* [1990]; 9, *Tric et al.* [1992]; 10, L. Tauxe and N. Shackleton (unpublished data, 1993); 11, *Gallet et al.* [1993, also unpublished data, 1992]; 12, *Hartl et al.* [1993]; 13, *Meynadier et al.* [1992]; 14, *Schneider et al.* [1992]; 15, *Clement and Kent* [1991]; 16, *Clement and Kent* [1986]; 17, *Valet et al.* [1989].

analysis using a minimum amount of data processing. In general, since each normalized paleointensity record is relative and not absolute, I have taken the liberty to scale every record, such that the plotted data have a mean of unity. I have not smoothed the data or manipulated amplitudes (apart from the scaling just described). Wherever possible, I have used data supplied by the authors, but in certain instances, I digi-

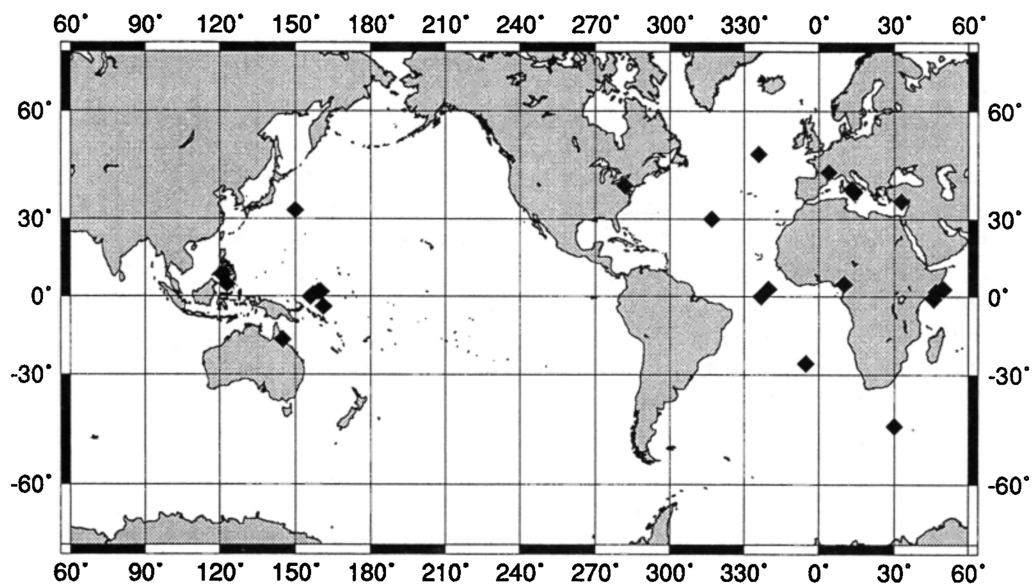

Figure 18. Locations of records discussed in text. See Table 4.

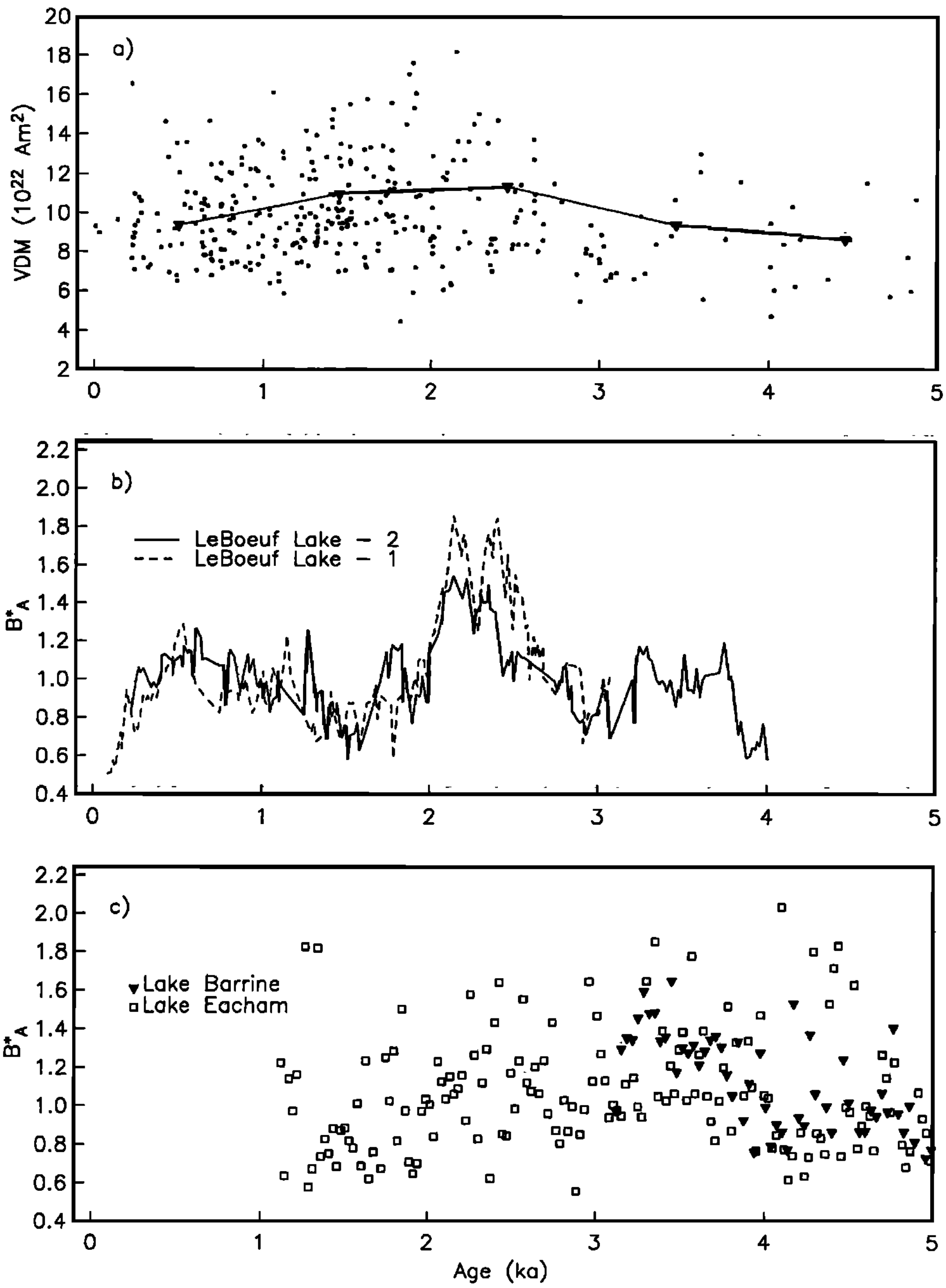

Figure 19. (*a*) Virtual dipole moment (VDM) data from Thellier-Thellier technique compiled by *Pick and Tauxe* [1993] and globally averaged dipole moments of *McElhinny and Senanayake* [1982]. (*b*) Data from *King et al.* [1983] for LeBoeuf Lake sediments. (*c*) Data from *Constable* [1985] for two Australian lakes (see Table 4). Data have been scaled to have a mean of unity.

tized data from figures in the published articles. I have adjusted the timescales to the new standard of *Cande and Kent* [1992*a*] and *Shackleton et al.* [1990].

In Figure 19 I plot paleointensity information available for the last 5 ka. The absolute paleointensity data determined by the Thellier-Thellier technique [*Thellier and Thellier*, 1959] as compiled by *Pick and Tauxe* [1993] are shown in Figure 19*a*. Individual data points are plotted as small circles and the global averages calculated by *McElhinny and Senanayake* [1982] are plotted as triangles. In Figure 19*b* are B^*_A data from two cores retrieved from LeBoeuf Lake in Pennsylvania presented by *King et al.* [1983]. Figure 19*c* shows the B^*_A data from two nearby Australian lakes [*Constable*, 1985]. There is a fair degree of agreement between the two records from each region, and each compared favorably to the absolute paleointensity data from the same continent. However, the Pennsylvanian and Australian lake records bear little resemblance to one another. Furthermore, there is little in common between the lake records and the global paleointensity data set. This lack of agreement is perhaps to be expected from short, high-resolution records, because the data reflect local geomagnetic field changes, and as such are heavily influenced by the nondipole field.

In Figure 20 I show data spanning the last 30 ka. Figure 20*a* shows the B^*_X data from Lac du Bouchet of *Thouveny* [1987], *Thouveny et al.* [1990], and *Creer et al.* [1990] as supplied by T. Williams. These have been smoothed by the various authors. The Lake Mbo data were digitized from *Thouveny and Williamson* [1988]. The data from INMD 48Bx are those of *Tauxe and*

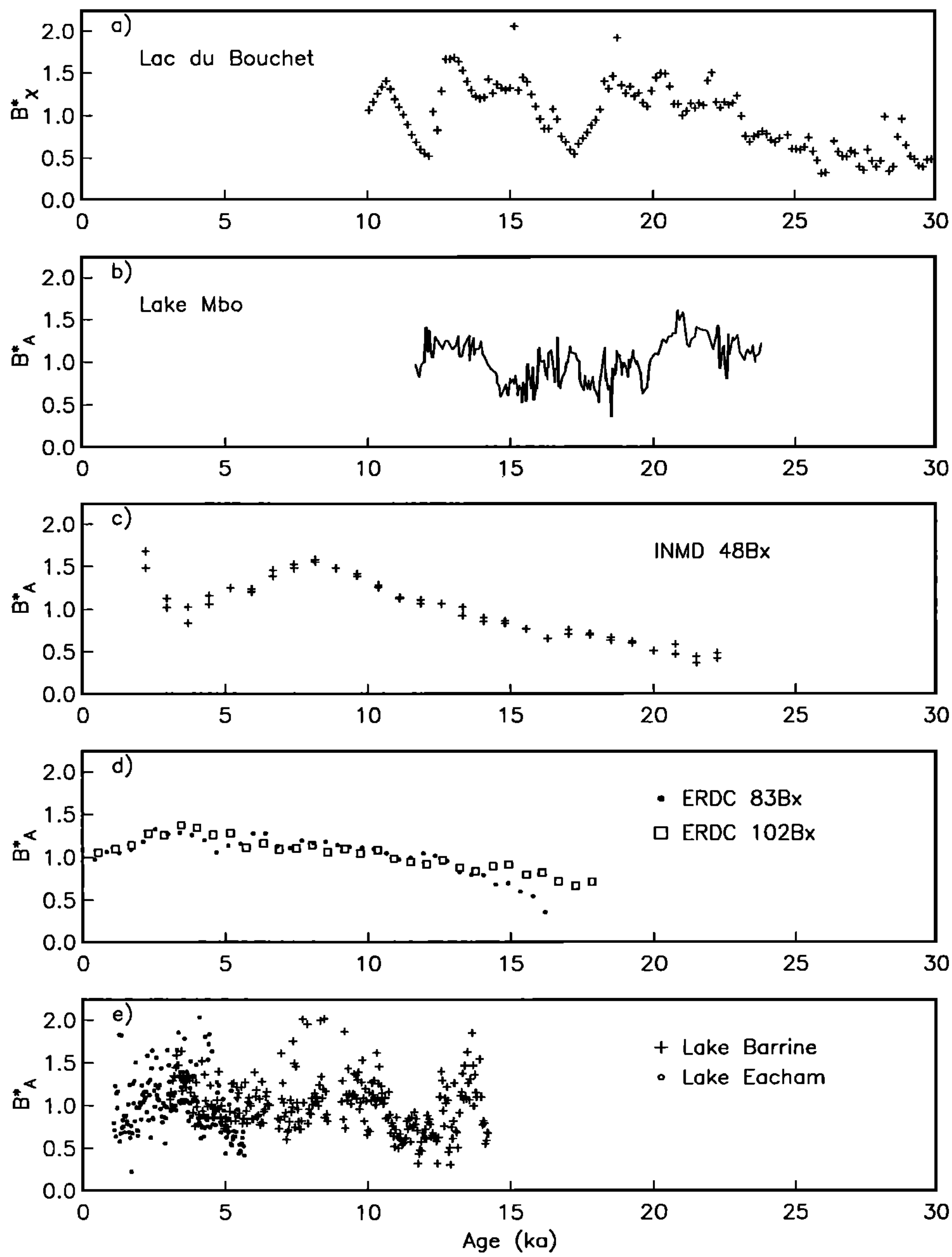

Figure 20. Paleointensity data (scaled to unit mean) spanning the last 30 ka (see Table 4 for data sources).

Valet [1989], and the ERDC box cores are from *Constable and Tauxe* [1987]. Figure 20*e* is the complete data set from the two Australian lakes of *Constable* [1985]. As noted for the 0–5 ka record, the individual records from a given region (e.g., ERDC 83Bx and 102Bx) are quite similar, but there is poor agreement among the various regions. *Constable and Tauxe* [1987] showed that the northeastern Australia data, when plotted as 500-year medians, are quite similar to the box core data from the Ontong-Java Plateau in that both records tend to show a peak in paleointensity at about 4 ka with a general decreasing trend back to some 15 ka. The scatter around 8 ka and the high at around 13–14 ka in the Lake Barrine data are not seen in the Ontong-Java records, but these are also associated with abnormal values in the Banerjee plot and may not be "real." However, there are high values around 8 ka in the North Atlantic box core record (INMD 48Bx) and a high around 13 ka in the Lake Mbo record from Cameroon. Although certain features may appear in more than one record, the degree of correspondence among all five records is not great.

Figure 21 shows data spanning the period 10–140 ka. Figure 21*a* shows the absolute paleointensity data compiled by *McElhinny and Senanayake* [1982]. Figure 21*b* is an extension of the Lac du Bouchet data [*Thouveny et al.*, 1990; *Creer et al.*, 1990]. The dashed lines in Figure 21*c* are the whole core B_χ^* data from four cores presented by *Tric et al.* [1992]. Also shown is their stacked record (solid line). In Figure 21*d* I show the data of *Meynadier et al.* [1992]. At the bottom are the whole core B_χ^* from ERDC 113p (solid

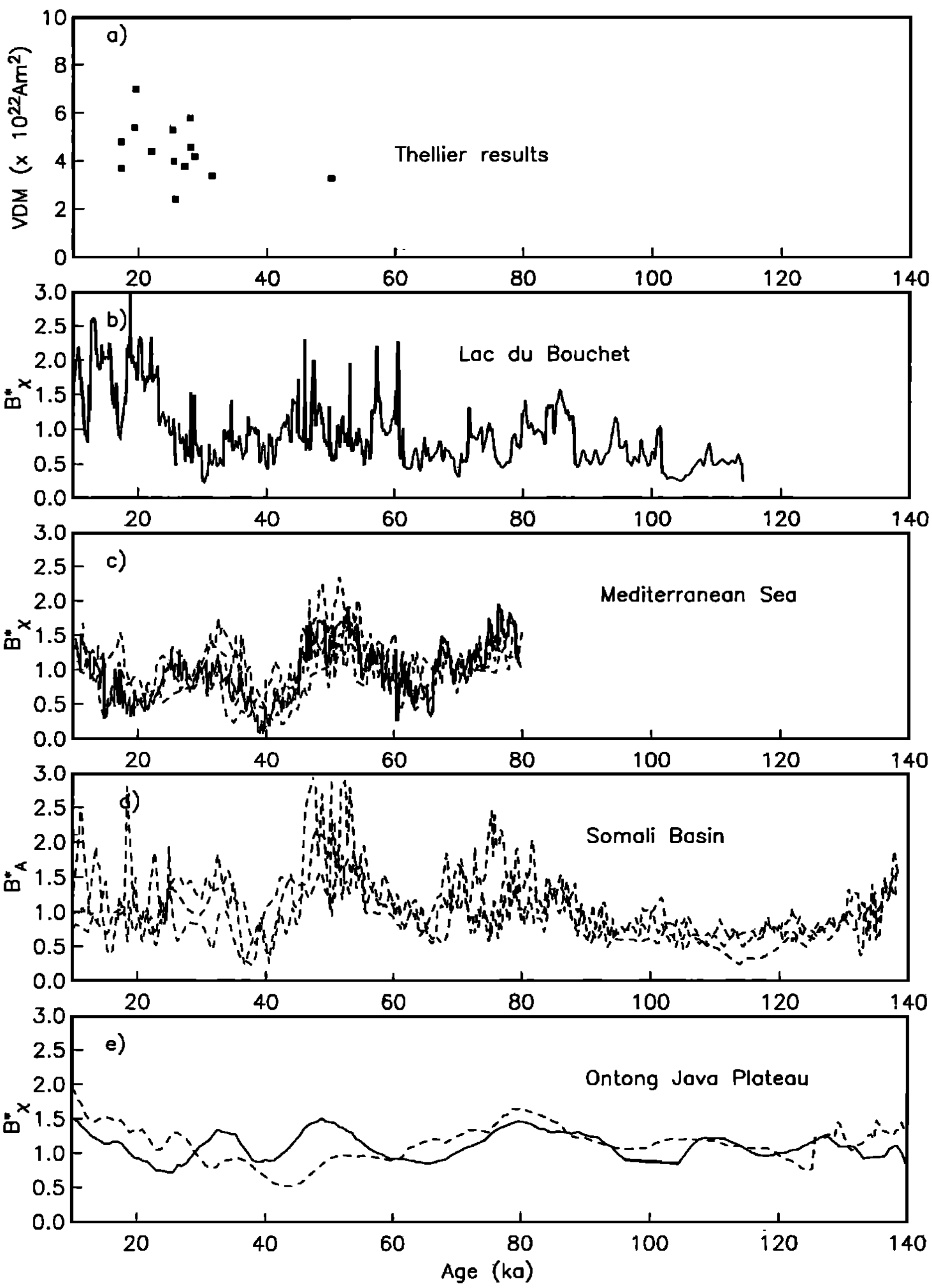

Figure 21. Same as Figure 20 but spanning 10–140 ka (see Table 4).

line) and ERDC 89p (dashed line) of *Tauxe and Wu* [1990] using an age model of N. Shackleton (personal communication, 1993). The eager eye can find similarities among portions of the five data sets. The Somali Basin data are quite similar to the Mediterranean Sea data. All of the data sets show lows at around 40 and 60 ka and most show a high at around 80 ka.

Let us turn now to have a look at records spanning the Brunhes Chron. In Figure 22 I plot data from three cores from the Ontong-Java Plateau (ERDC 89p, 113p and RNDB 75p) and one from the North Pacific (RC10-167) (see Table 4). The Ontong-Java records [*Tauxe and Wu*, 1990; also L. Tauxe and N. Shackleton, unpublished data, 1993] were dated using oxygen isotopes. RC10-167 was dated using only the position of the B/M boundary [*Kent and Opdyke*, 1977]. All records were converted to absolute ages using the new timescale proposed by *Shackleton et al.* [1990] and *Cande and Kent* [1992*a*]. In contrast to the data sets viewed on shorter timescales the four records in Figure 22 show more coherence, particularly when small changes in sedimentation rate are allowed between RC10-167 and the oxygen isotope calibrations of the others. The coherence between ERDC 113p and ERDC 89p was plotted in Figure 17*d* and is significant. Without isotopic calibration of RC10-167 (which is impossible in this very low carbonate core), coherence between it and the other records is not meaningful, but many common features can be identified by eye.

Two additional data sets are shown in Figures 23

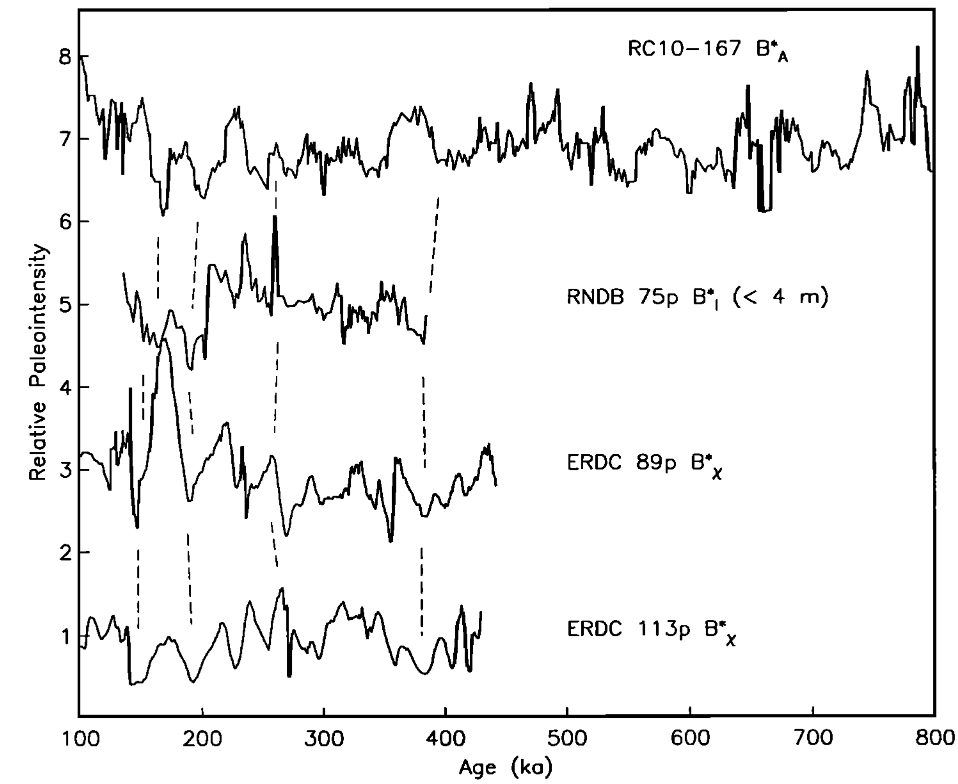

Figure 22. Relative paleointensity data spanning the Brunhes Chron (see Table 4).

and 24 from the Matuyama and Chron C12R, respectively. Neither of these has yet been duplicated, so the reliability cannot be fully assessed. Figure 23 is an extension of the Ontong-Java records into the Matuyama as acquired by *Gallet et al.* [1993, also unpublished data, 1992]. They are NRM (demagnetized to 300°C) normalized by χ. There are several periods of unusually low relative paleointensity values associated with the boundaries of the major known subchrons as well as several other suspected "excursions" (see *Champion et al.* [1988] for a review). In Figure 24 I have plotted the data from *Hartl et al.* [1993] from the early Oligocene. These are NRM demagnetized to 17.5 mT normalized by M_{sr} (B_I^*). Again several low relative paleointensity values are associated with the chron boundaries. In addition to these periods of low relative intensity, there are four other periods with equally low values of B_I^* within Chron C12R. These are associated with periods of "directional instability" marked by dashed lines and a shaded zone in the remanent directions.

Finally, although paleomagnetic reversals are not a focus of this paper, it is appropriate to discuss briefly the relative paleointensity records associated with some transitional data sets. Most publications on reversals in sediments include some intensity data. However, until recently, these were usually not systematically normalized, with bulk parameters measured only on pilot specimens (see, for example, *Clement and Kent* [1984]). In the last few years it has become more popular to provide fully normalized data (see, for example, *Clement and Kent* [1991]). In gen-

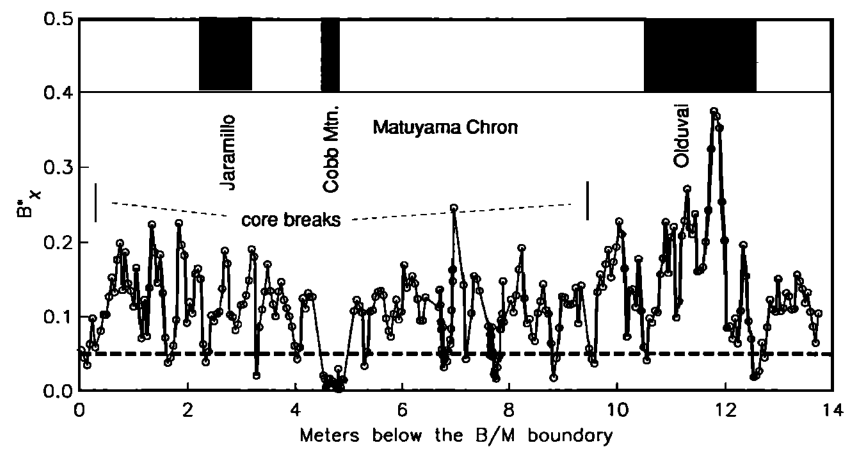

Figure 23. Relative paleointensity data from the Matuyama Chron of *Gallet et al.* [1993; also unpublished data, 1992] (see Table 4).

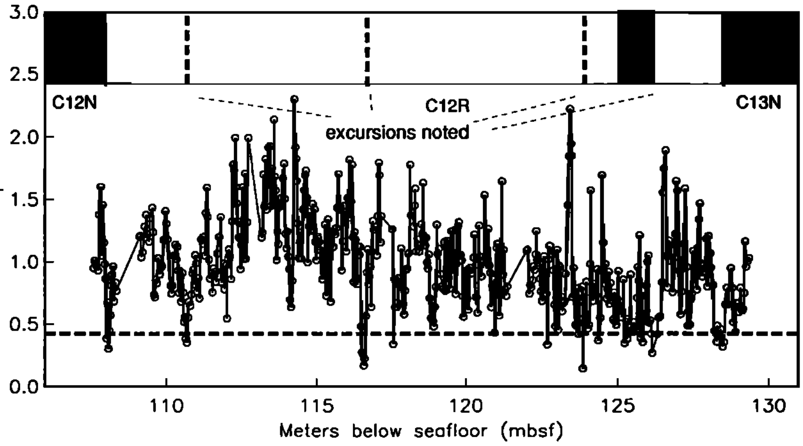

Figure 24. Relative paleointensity data Chron C12R of *Hartl et al.* [1993] (see Table 4).

eral, there are only a few adequate records for a given reversal, but the Brunhes/Matuyama transition is an exception. I have gathered seven records of the relative paleointensity data in Figure 25. These are plotted against depth, with the regions associated with intermediate directions bounded by vertical bars. (In several cases the transition is so narrow that the bars superpose.) Each relative paleointensity record was scaled by

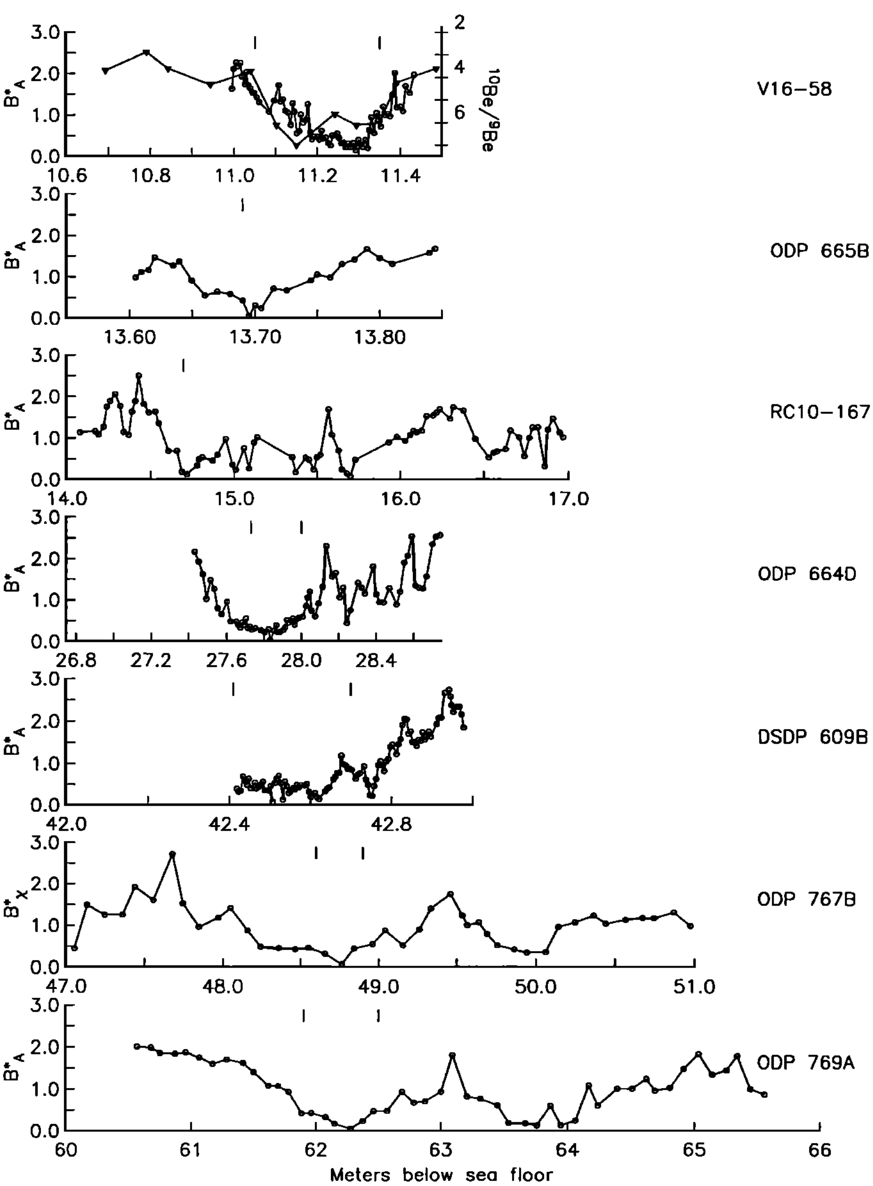

Figure 25. Relative paleointensity data from the Brunhes/Matuyama polarity reversal. Data are plotted versus depth and the transitional intervals are enclosed by vertical lines (see Table 4).

its mean as before. In the topmost record I have also plotted the beryllium isotope data available from *Raisbeck et al.* [1990], as quoted by *Clement and Kent* [1991]; note the inverted axes for the beryllium data.

The relative paleointensity data from the Brunhes/Matuyama reversal are broadly similar to each other in that the reversal itself is associated with low values of the relative paleointensity. Also, the longer records show equally low values some 20,000–40,000 years prior to the actual transition as noted by *Schneider et al.* [1992]. Perhaps most encouraging for those who wish to interpret normalized remanence as relative paleointensity is the agreement between the B^* data and the beryllium isotopic data reported by *Clement and Kent* [1991]. Beryllium 10, like ^{14}C, is produced in the upper atmosphere by bombardment of cosmic rays. Its production, like that of radiocarbon, is modulated by the dipole moment of the geomagnetic field. Thus the variations in beryllium isotopes plotted in Figure 25 could reflect changes in the paleofield intensity, and the general peak in production corresponding to low relative intensity values in all the records supports the hypothesis that the normalized remanence reflects relative paleofield variations.

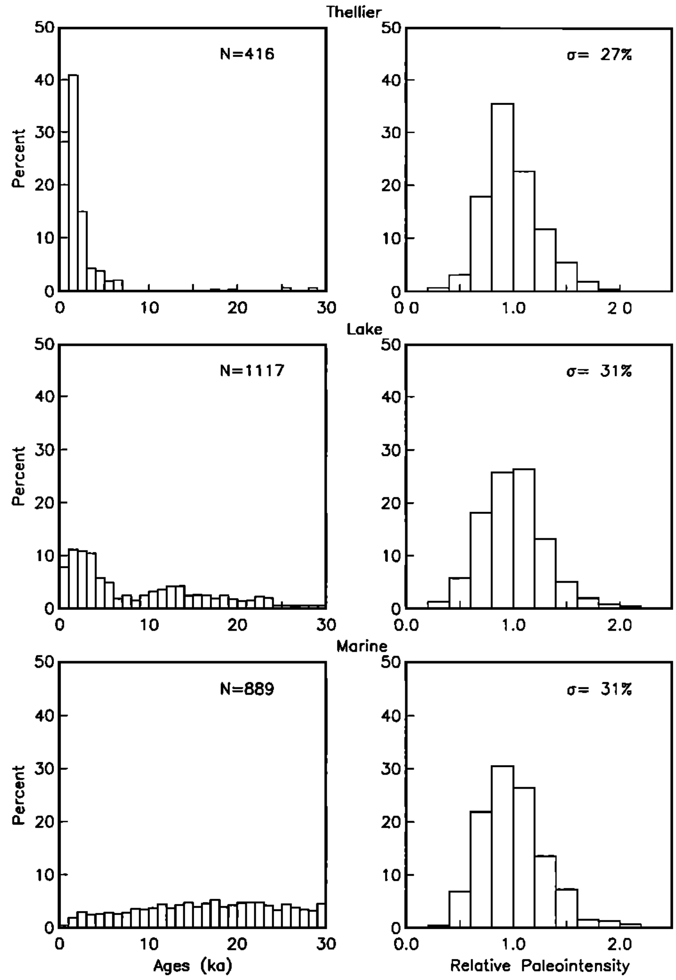

Figure 26. Comparison of data derived from absolute paleointensity methods (Thellier) and relative paleointensity from lake and marine sources. See text.

11. DISCUSSION

11.1. Comparison of Absolute and Relative Paleointensities

Having considered the bulk of the published records, it is now worthwhile discussing what, if anything, these records mean. The question I address now is, Do these records reflect genuine geomagnetic field variations or are they dominated by mineralogical effects? There is no direct method for verifying sedimentary paleointensity data because observational data are available for only the last few hundred years or so. Absolute paleointensity data with sufficient temporal density are also sparse, but occasionally are adequate for a meaningful comparison. *King et al.* [1983] and *Constable* [1985] compared their sedimentary records with the local archeomagnetic data sets and in both cases they were found to agree remarkably well. *Tric et al.* [1992] compared their record with the paleointensity data from Mount Aetna and were satisfied with the general agreement. Apart from these few studies, direct comparison of time series of thermally acquired and sedimentary paleointensity data has been hindered by the lack of absolute paleointensity data with sufficient age control.

Given the spotty nature of the absolute paleointensity data set derived from volcanic and archeological materials, we are forced to employ less direct methods of comparison. For example, *Tauxe and Wu* [1990] considered whether the distribution of relative paleointensity data was similar to that of lava flow data. They compared their records with durations of some few hundred thousand years derived from a single region to the more or less globally distributed lava flow data spanning the last 5 million years. In Figure 26 I have taken a similar approach to that of *Tauxe and Wu* [1990], but have tried to compare data sets that were more similar in terms of time span and global coverage. First, I separate the relative paleointensity data into lake and marine sediments spanning the last 30,000 years (shown in Figures 19 and 20). The absolute paleointensity data are the Thellier-Thellier results shown in Figures 19 and 21. This combined data set has also been scaled by the mean for comparison with the sedimentary records. The data are plotted both as histograms of the ages (to the left) and as histograms of the relative paleointensities (to the right). As noted frequently in the literature and in this paper, the age distribution of the Thellier data set is heavily biased to the last few thousand years, whereas sedimentary records are more evenly distributed in time. Nonetheless, the distributions of the relative intensity data derived from the different sources are extremely similar. The dynamic range of the three data

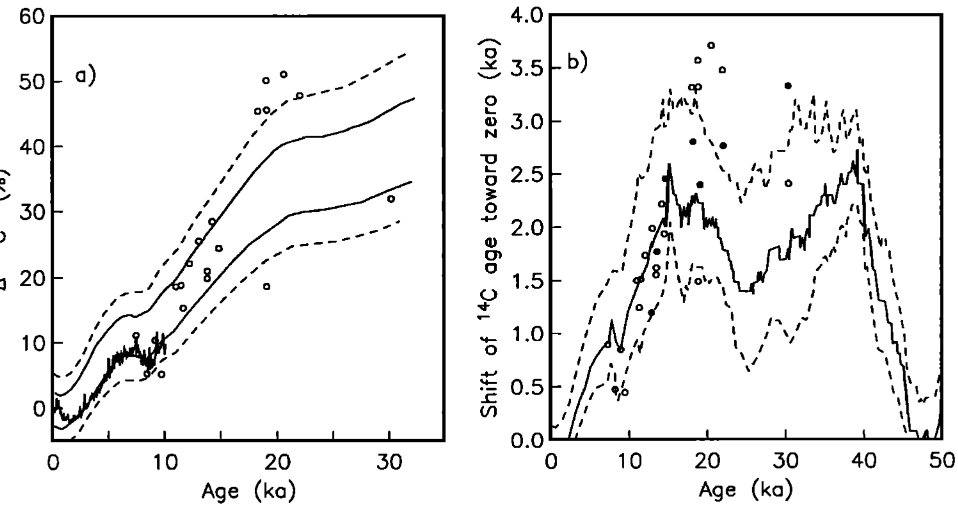

Figure 27. (*a*) $\Delta^{14}C$ data are plotted versus age. Solid (dashed) lines assume uncertainties of 5% (10%) on dipole moment data of *McElhinny and Senanayake* [1982]. Circles are based on comparing U-Th ages with radiocarbon ages. Uncertainties have been left off for clarity, but intersect the dashed lines. Redrawn from *Bard et al.* [1990*a*]. (*b*) Solid line is predicted age discrepancy (see text) with uncertainties shown by dashed line; open circles are from U-Th/^{14}C data [*Bard et al.*, 1990*a*] and filled circles are from *Bard et al.* [1990*b*]. Redrawn from *Mazaud et al.* [1991].

sets is virtually identical, suggesting that sediments are able to record the full range of geomagnetic field fluctuations (including quite weak fields of 10–20% of the mean value). The slight difference in scatter (as reflected by the standard deviation expressed as a percent of the mean) could be caused by the skewed age distribution of the Thellier data set.

11.2. Comparison of Radiocarbon Production and Relative Paleointensity Data

Another source of information about changes in the geomagnetic field is the variation in production in radiogenic isotopes found in the upper atmosphere as mentioned briefly in the introduction and in discussing the data from the last reversal. Of particular interest is the period of time back to about 35 ka, as there are data on the discrepancies between the radiocarbon timescale assuming constant production and the ages actually observed (see summary by *Mazaud et al.* [1991]). For the last 10 ka there is an abundance of absolute paleointensity measurements (see Figure 19*a*), and a reasonable estimate of the global (dipole) field variations was made by *McElhinny and Senanayake* [1982]. Taking these data and assuming that the dipole field is entirely responsible for changes in radiocarbon production Q in the manner estimated by *Elsasser et al.* [1956], i.e.,

$$Q/Q_o = (M/M_o)^{-1/2}$$

yields a predicted $\Delta^{14}C$ variation shown by the dashed line in Figure 1. The variation observed by "dating" tree rings and varved lake sediments of known calendar age is shown by the solid line, and the agreement is quite good [*Tauxe and Valet*, 1989].

Bard et al. [1990*a*, *b*] took this approach a few steps farther. They extended the $\Delta^{14}C$ record back to some 30,000 years by comparing U/Th ages (assumed accurate) to the ^{14}C ages in Barbados corals. Using the data of McElhinny and Senanayake for the last 30,000 years (Figure 21*a*) and the calculations of *Lal* [1988] to predict changes in ^{14}C production, as well as the two-box model of *Houterman et al.* [1973] that attempts to simulate the filtering of the ocean/atmosphere system on ^{14}C production, they calculated a predicted $\Delta^{14}C$ record. Comparing the $\Delta^{14}C$ data predicted from the magnetic paleointensity data as modulated by the models of *Lal* [1988] and *Houterman et al.* [1973] with the observed $\Delta^{14}C$ data from tree rings, varved sediments, and U-Th calibrations yielded the results shown in Figure 27*a*. The principal problem with this approach is that the data of *McElhinny and Senanayake* [1982] are very sparse prior to about 10 ka, consisting of a handful of individual data points (see Figure 21). Each data point is a spot reading of the field (similar to the open circles in Figure 19*a*) and is expected to have a scatter about the true dipole moment of some 20%. Production of ^{14}C is modulated by variations in the dipole moment; the uncertainty in the paleofield estimate assumed by *Bard et al.* [1990*a*] was 5–10% (solid and dashed lines, respectively, in Figure 27*a*). The question is, Do so few data points from so few places represent the dipole moment of the Earth's magnetic field?

Seizing an opportunity to refine the estimates of dipole field variations and their effect on radiocarbon production, *Mazaud et al.* [1991] used the data shown in Figure 21*c* (solid line). They used the stacked record, some additional $\Delta^{14}C$ data from *Bard et al.*

[1990*b*], and the same production and Earth filter models as *Bard et al.* [1990*a*] to produce Figure 27*b*.

Both Figures 27*a* and 27*b* show discrepancies between the predicted and observed ^{14}C anomalies. Examination of the relative paleointensity data used to calculate the solid line in Figure 27*b* sheds some light on the matter. Taking a close look at the data in Figure 21 one can observe the following: First, the absolute paleointensity data, the data from Lac du Bouchet, and the data from ERDC 113p all show a general decrease in paleointensity from about 10 ka to about 30 ka. At least some of the records from the Mediterranean Sea and the Somali Basin also show this trend. However, two cores presented from the Mediterranean Sea show a sharp increase in paleofield at around 30 ka. In the stacked record, processed to produce relative variations in ^{14}C concentrations, this results in the trough in predicted ^{14}C age anomalies in Figure 27*b*, at variance with those observed. Combining data from all the sources shown in Figure 21 would yield results in much better agreement with the observed age anomalies. No doubt, the addition of more and better data sets will prove invaluable in the detailed calibration of the radiocarbon timescale in the future.

11.3. Spectral Analysis of Long Time Series and Periodicities of the Field

Kent and Opdyke [1977] were the first to perform spectral analysis on time series of relative paleointensity data. They found a broad peak in amplitude centered around 43 ka in the Brunhes portion of RC10-167 (data shown in Figure 22). They noted the similarity in this period to that of the variations in obliquity of the Earth's orbit (41 ka [e.g., *Hays et al.*, 1976]) and suggested that perhaps the Earth's orbit could in some way influence the geodynamo.

Tauxe and Wu [1990] performed a similar analysis on a stacked data set combining ERDC 113p, 89p and RC10-167 (all in Figure 22). They broke the record down into early Brunhes and late Brunhes and found a broad peak in amplitude centered on about 33 ka in the late Brunhes portion but no significant peaks in the early Brunhes portion. They suggested that there are apparent periodicities that are not permanent features of the geomagnetic field and are probably unrelated to Milankovitch forcing.

It is now possible to extend this line of investigation, incorporating the new data for the Matuyama of *Gallet et al.* [1993, also unpublished data, 1992]. Each record is scaled by its mean and interpolated at equal temporal spacing using a spline. Then, by piecing together the Brunhes portion of the RC10-167 record and the Matuyama portion of the ODP 803A record, we have a 2.2-million-year time series of normalized remanence.

I have analyzed this time series using the multitaper spectral techniques described by *Tauxe and Wu* [1990]. The data set is analyzed in 500,000-year pieces

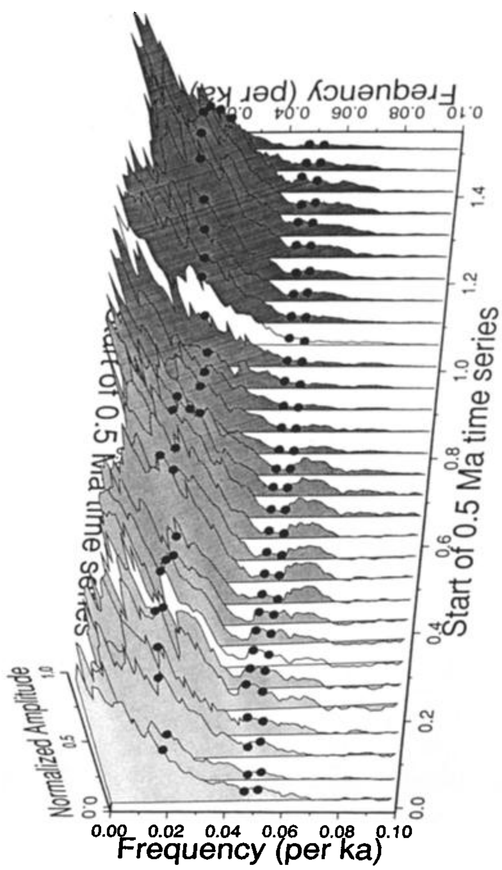

Figure 28. Spectral analysis of RC10-167 and ODP 803A time series. Data windows are 0.5 Ma long starting at 50-ka intervals along the time series. Milankovitch frequencies corresponding to periods of obliquity and precession are denoted by solid circles. The unshaded time slices are examined in more detail in Figures 29 and 30.

starting at 10,000 years. The analysis is repeated by sliding the data window back at 50,000-year intervals. The resulting 31 spectra are plotted in a three-dimensional perspective in Figure 28. The frontmost spectrum is from a piece of the time series, starting at 10 ka and ending at 510,000 ka. For purposes of discussion, solid circles are plotted at frequencies corresponding to the Milankovitch periods of precession (19 ka, 21 ka) and obliquity (41 ka). As noted by *Tauxe and Wu* [1990], there are indeed some portions of the late Brunhes that exhibit amplitude peaks at Milankovitch frequencies. These are less evident in the earlier Brunhes, where a peak centered at around 17 ka dominates. Finally, in the Matuyama record, there are some spectra that appear to have convincing peaks near the frequency of obliquity.

Before any claims are made concerning the possible control of the Earth's orbit on the geodynamo, I should follow the example of *Tauxe and Wu* [1990] and test the paleointensity records for coherence with the normalizer. I will examine in detail two time series of

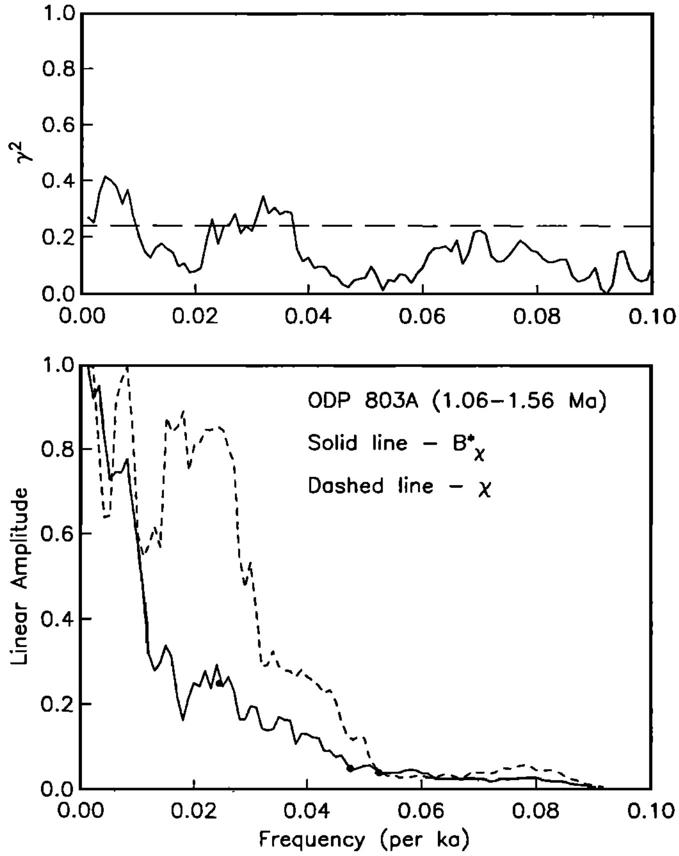

Figure 29. Spectral analysis of portion of ODP 803A data spanning 1.06–1.56 Ma is shown in the bottom panel. The solid line is the spectrum of the normalized remanence data and the dashed line is the spectrum of the susceptibility data alone. Solid circles are the frequencies corresponding to obliquity and precession. The squared coherence of the two time series is plotted in the top panel. The dashed line is the 95% confidence limit for zero coherence.

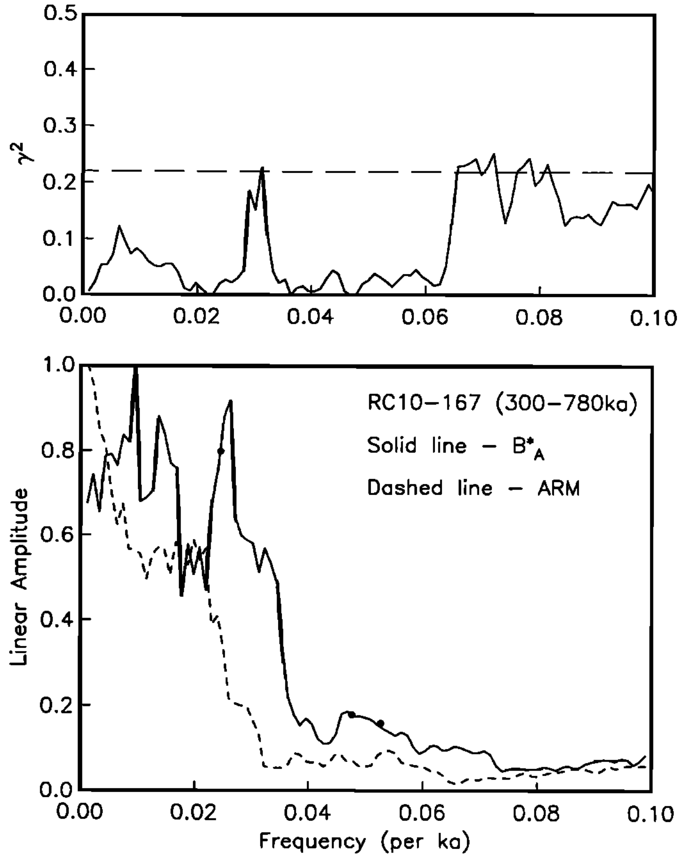

Figure 30. Same as Figure 29 but for RC10-167 spanning from 300 to 780 ka.

the 31 shown in Figure 28. These are highlighted by having the spectra plotted as an unshaded panel and are those starting at 310 ka and 1.06 Ma from RC10-167 and ODP 803A, respectively. Considering the 1.06 Ma panel first, I plot the spectrum of the relative paleointensity time series again in Figure 29 as the solid line in the lower panel. The spectrum of the susceptibility is plotted as the dashed line; the two are not very similar, but portions are significantly coherent, particularly in the obliquity band, suggesting that the peak near the frequency of obliquity could be controlled in part by uncompensated lithologic variability.

Turning now to the late Brunhes time series, I plot again the spectrum of the time series of RC10-167 data starting at 310 ka (frontmost unshaded panel in Figure 28) as the solid line in the lower panel of Figure 30. The spectrum of the corresponding ARM data is shown as the dashed line, and coherence is plotted in the upper panel. In this time series the normalized remanence is not coherent with ARM except possibly at quite high frequencies. The peak in B^* at about 40 ka is the same as that identified by *Kent and Opdyke* [1977] and again by *Tauxe and Wu* [1990], and I echo the conclusions of the latter that portions of the Brunhes relative paleointensity time series appear to have amplitude peaks near frequencies corresponding to Milankovitch forcing. However, not only are these not permanent and stable features of the entire time series, but there are many other peaks as well. I remain skeptical of the notion that the Earth's orbit plays a role in modulating the geodynamo.

11.4. Comparison of Marine Magnetic Anomalies With Relative Paleointensity Data

Cande and LaBrecque [1974] and *Blakely* [1974] noted the presence of correlatable small-wavelength features in marine magnetic anomaly data (known as "tiny wiggles"). They can be modeled either as reversals of the Earth's magnetic field keeping the field intensity constant [*Blakely*, 1974] or as fluctuations in the intensity of the field keeping direction constant [*Cande and LaBrecque*, 1974]. *Cande and Kent* [1992*b*] recently elaborated on the intensity fluctuation model. Their model derives from an analysis of paleointensity data from lavas done by *McFadden and McElhinny* [1982], whereby dipole moment data were shown to have a normal distribution, truncated because of the impossibility of having negative field values. *Cande and Kent* [1992*b*] demonstrated that if crustal magnetizations are distributed in the same fashion and these are viewed through an "Earth filter" (that is upward continued to the sea surface), then features quite similar to tiny wiggles can be produced. Their model then is that the Earth's dipole moment has behaved as it did for the last 5 Ma, throughout the Cenozoic, that the magnetizations in the ocean crust are proportional to these field values, and that upward continuation of the crustal magnetizations results in tiny wiggles in the magnetic anomalies. This model is difficult to test. A proper test would require drilling numerous holes through the ocean crust and acquiring absolute paleointensity data. This simply is not feasible for many reasons. However, *Hartl et al.* [1993] suggested that it may be possible to assess the plausibility of the model using sedimentary relative intensity data.

Some of the most prominent tiny wiggles were observed between anomalies 12 and 13 on the Juan de Fuca plate [*Cande and LaBrecque*, 1974]. *Hartl et al.* [1993] presented relative paleointensity data of exceptionally high quality through this time interval (data shown in Figure 24). As noted previously, there are several periods of unusually low relative intensity associated with directional "instability." One of the key assumptions in the Cande-Kent model is that the field during Chron C12R behaved in a similar manner to the last 5 Ma. The data set of *McFadden and McElhinny* [1982] was composed of absolute paleointensity data from lava flows. Data associated with intermediate directions were not included. Since the field during Chron C12R may have exhibited excursionary behavior, I have included all the data compiled by McFadden and McElhinny (as well as those published later by *Roberts and Shaw* [1984]) in the analysis here, including those associated with intermediate directions. The data set is also scaled by its mean to facilitate comparison with the relative paleointensity of *Hartl et al.* [1993]. The distribution of the paleosecular variation in lava flows (PSVL) is shown in Figure 31 (bottom panel) and is quite similar to that published by *McFadden and McElhinny* [1982] but with a larger standard deviation (58% as opposed to 47% of the mean). The distribution of the relative paleointensity data from Chron C12R (shown in Figure 24) is plotted in the top panel of Figure 31. Also shown are the relative paleointensity data derived using ARM and χ as normalizers; these are essentially identical. On the basis of this treatment, it appears that the model of *Cande and Kent* [1992*b*] for the origin of the tiny wiggles between anomalies 12 and 13 is certainly plausible.

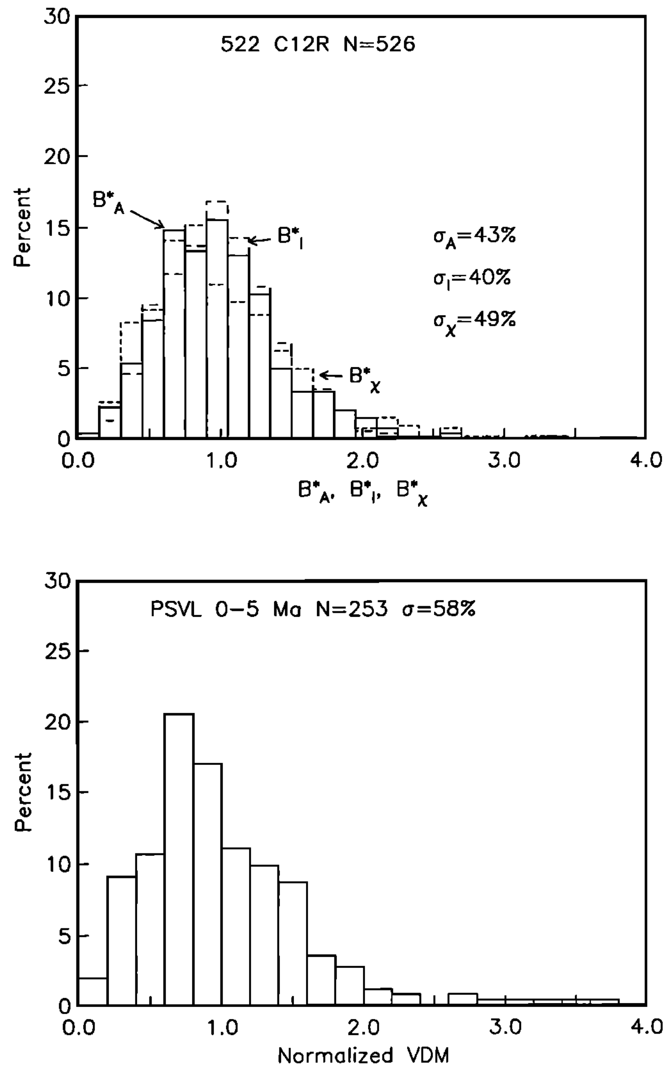

Figure 31. (Top) Histograms of relative paleointensity data from DSDP 522 [*Hartl et al.*, 1993]. (Bottom) Data from lava flows (PSVL) as compiled by *McFadden and McElhinny* [1982] as well as those of *Roberts and Shaw* [1984]. Data are scaled to have a mean of unity.

12. CONCLUSIONS

In this paper we have reviewed the experimental and theoretical foundations for using sedimentary sequences as recorders of the variations in paleointensity of the Earth's magnetic field. On the basis of this review, several attributes are considered desirable for a given sedimentary sequence to yield reliable paleointensity information, including magnetite mineralogy with relatively homogeneous concentration and grain sizes throughout the sequence. The directions of magnetization should retain excellent records of the geomagnetic field variations with uncomplicated behavior during demagnetization. Normalization should be attempted with several parameters, and good agreement among the various estimates of relative paleointensity is required. Furthermore, the relative paleointensity data should not be coherent with the bulk magnetic parameters. Finally, records from a given region should be similar within the limits of the time control.

I have also reviewed existing paleointensity data sets. When viewed over timescales of a few thousands of years, data sets from more than a few thousand kilometers distance bear little resemblance to one another, suggesting that they are dominated by nondipole field behavior. When viewed over timescales of tens to hundreds of thousands of years, however, the records show coherence over large distances (at least thousands of kilometers) and may reflect changes in the dipole field. Comparison of such paleointensity data sets with independent measures of dipole field intensity (such as production of cosmogenic nuclides) requires averaging of records taken over the globe or averaging over sufficient time (apparently several thousand years) in order to estimate changes in dipole moment as opposed to local nondipole field effects. Finally, on the basis of several sequences spanning the Brunhes Chron, the dipole field has oscillated with a period of 30–40 ka for the last few hundred thousand years, but these oscillations are not apparent in the record prior to about 300 ka; thus they are not an inherent feature in the geomagnetic field, and the correspondence of the period of oscillation to that of obliquity is probably coincidence. Existing data from the Matuyama Chron are too contaminated with lithological variations to make a meaningful analysis of that time period. Finally, sedimentary paleointensity data from Chron C12R demonstrate the plausibility of the model of *Cande and Kent* [1992*b*] in which tiny wiggles observed in marine magnetic anomalies are assumed to result from the fluctuations in intensity of the geomagnetic field.

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