PALEOMAGNETISM, DEEP-SEA SEDIMENTS

Introduction

Deep-sea (pelagic) sediments, deposited remotely from sources of continental detritus, have been a very important source for learning about the direction and intensity of the ancient geomagnetic field (paleomagnetism) because they often carry a primary natural remanent magnetization (NRM) acquired at the time of deposition, or shortly thereafter. The age of a primary magnetization can be determined from the accompanying biostratigraphy. The magnetization directions of known age have been used to assess plate motion (continental drift) or to record the characteristics of the ancient geomagnetic field, such as the sequence of polarity reversal in the geologic past. The record of geomagnetic polarity reversal (magnetostratigraphy) in deep-sea sediments, first practiced by Opdyke et al. (1966), has become important in paleocanography and biostratigraphy, and in geologic timescale construction. The geomagnetic polarity time scale (GPTS), based on the sequence of polarity reversal through time, is the central thread to which the other facets of geologic time (radiometric, isotopic, biostratigraphic) are correlated in the construction of geologic time scales. The sequence of polarity reversal through time, is the central thread to which the other facets of geologic time (radiometric, isotopic, biostratigraphic) are correlated in the construction of geologic timescale (GPTS). Prior to ~150 Ma, the GPTS is less well defined, due to the lack of in situ oceanic crust and hence marine magnetic anomaly records, and is not adequately correlated to the biozonations that define geologic stages. The global synchronicity of polarity reversals (of the main axial dipole field) means that magnetic polarity stratigraphies can be used to correlate environmental (isotopic) and biostratigraphic events among contrasting environments and remote locations. The stochastic (unpredictable) occurrence of polarity reversals, and our inability to distinguish one normal (reverse) polarity chron from another, means that precise correlation through magnetic polarity stratigraphy is only possible at polarity reversals, with interpolation between these tie points. Within individual polarity chron (time intervals between polarity reversals) magnetic records can be used for correlation if geomagnetic directional “secular” variation and/or paleointensity are adequately recorded. The conditions for adequate recording of secular variation and geomagnetic paleointensity are more stringent than for the recording of polarity reversals as contamination (overprint) of primary magnetization is less crippling for polarity records than for the more subtle changes that define secular variation and paleointensity.

Origin of primary magnetizations

What makes deep-sea sediments efficient recorders of the geomagnetic field at time of deposition? Sediments can acquire a detrital remanent magnetization (DRM) at the time of deposition by mechanical orientation of fine grained magnetite (Fe₃O₄) or titanomagnetite (±Fe₂TiO₅ [1-x] Fe₃O₄) into line with the ambient geomagnetic field at the sediment–water interface. The natural remanent (permanent) magnetization (NRM) of ferrimagnetic (titanom) magnetite results in a torque that statistically orients the magnetic moment of the grain population into line with the ambient geomagnetic field. The mechanical orientation of grains may be achieved either at the sediment–water interface (DRM) or in the uppermost few centimeters or decimeters of the sediment in which case the resulting remanence is referred to as DRM (post-depositional detrital remanent magnetization).

Figure P22 The sigmoidal function based on tanh(x) used to model the mixing zone (thickness M) and the underlying lock-in zone (thickness L) (after Channell and Guyodo, 2004).
et al. (2000) estimated mixed layer thicknesses of 10–20 cm in box cores collected close to the Rockhall Plateau. These values are greater than the 2–13 cm mixed layer thicknesses obtained from farther south in the North Atlantic using $^{14}$C in foraminifera (Trauth et al., 1997; Smith and Rabouille, 2002). Estimates of mixed layer thickness are grain size sensitive (see Band, 2001) and would be expected to be lower for the coarse fraction (foraminifera) than for the bulk carbonate (nannofossils). For this reason, mixed layer thickness estimates based on $^{14}$C and other isotopic tracers should be considered as minimum estimates for the fine (PSD) grains that carry stable magnetization. The main control on the mixed layer thickness appears to be organic carbon flux derived from surface water productivity (Trauth et al., 1997; Smith and Rabouille, 2002). A recent redposition study has suggested that, at least for some lithologies, intergranular interactions overcome the magnetic aligning torque so that bioturbation would not enhance, but rather disrupt, the remanent magnetization (Katari et al., 2000). This proposition is in agreement with Tauxe et al. (1996) who, based on an analysis of the position of the Matsuyama–Brunhes boundary relative to oxygen isotope records, concluded that magnetization lock-in depth is insignificant in marine sediments. Other studies (DeMenocal et al., 1990; Lund and Keigwin, 1984; Kent and Schneider, 1995; Channell and Guyodo, 2004) invoked pDRM, and hence a finite (decimeter-scale) lock-in depth, to explain reversal–isotope correlations and observed attenuation of secular variation records.

Magnetite and titanomagnetite grains carrying DRM or pDRM in marine sediments may be of detrital or biogenic origin. Titanomagnetite is likely to have a detrital origin from the weathering of igneous rocks (such as mid-ocean ridge or oceanic island basalts) or from volcanic ash falls. A magnetic remanence known as thermal remanent magnetization (TRM) is acquired as the (titano)magnetite grain cools through a blocking temperature spectrum within its igneous host. The detrital grains retain this TRM during erosion, transport, and subsequent incorporation into the deep-sea sediment. The interaction of the TRM of individual grains with the ambient geomagnetic field at the time of sediment deposition generates the orienting torque that produces DRM or pDRM.

Magnetite (Fe$_3$O$_4$), with low or insignificant Ti (titanium) substitution, is commonly associated with biogenic sources. Magnetotactic bacteria are ubiquitous in freshwater and marine environments (e.g., Kirschvink and Chang, 1984; Petersen et al., 1986; Vali et al., 1987). They inhabit aerobic and anaerobic sediments and form intracellular chains of fine-grained (single domain) “magnetosomes” of magnetite (Blakemore and Blakemore, 1990) or, less commonly, goethite (Mann et al., 1990; Heywood et al., 1990). Magnetite and goethite magnetosomes have been observed to coexist within the same bacterium (Bazylinski et al., 1995). The intracellular magnetite and goethite crystals are formed by a process referred to as biologically controlled mineralization (BCM) (Lowenstam and Weiner, 1989; Bazylinski and Frankel, 2003). Intracellular magnetite and goethite are usually structurally well ordered and have a narrow (single domain) size distribution that is optimal for the retention of magnetic remanence in ferrimagnetic minerals such as magnetite and goethite (Figure P23).

Other Fe(III)-reducing bacteria secrete magnetite, and a range of other iron minerals, outside the cell by a process referred to as biologically induced mineralization (BIM) (Lowenstam and Weiner, 1989; Frankel and Bazylinski, 2003). Magnetite and other iron minerals produced by BIM are often poorly crystalline, fine-grained, and not structurally ordered. In the case of Geobacter metallo reducing (also referred to as bacterium strain GS-15), magnetite is produced outside the cell as fine (superparamagnetic) grains produced by the oxidation of organic matter and Fe(III) reduction (Lovley, 1990). Unlike magnetotactic bacteria, these iron reducers are nonmagnetotactic (nonmotile), unconcerned about the size, shape, or composition of the iron mineral product, and can produce a range of iron minerals depending on the nature of the surrounding medium.

Magnetotactic bacteria apparently utilize the internal magnetite–goethite chains to navigate along geomagnetic field lines (magnetotaxis) to find optimal redox conditions in the near-surface sediment. In the absence of sensitivity to gravity due to neutral buoyancy in seawater, the ambient magnetic field provides up-down orientation, either in the northern or southern hemisphere (Kirschvink, 1980). The single-domain grain size typical for magnetite and goethite produced by BIC indicates that magnetic remanence is central to magnetosome function. Living magnetotactic bacteria tend to be concentrated at depths of few tens of centimeters below the sediment–water interface at the transition from the iron-oxidizing to iron-reducing conditions (Karlin et al., 1987). The conditions under which BIM and BCM of magnetite takes place, and the depth in the sediment column in which these microbes are active (see Liu et al., 1997), are clearly of great importance in understanding the origin of the magnetic signature in sediments. Magnetic methods for the recognition of magnetosome chains, and individual magnetosomes, have been proposed by Moskowitz et al. (1988, 1994).

Apart from bacteria, other biogenic marine sources of magnetite are chiton (mollusk) teeth, fish, whales, and turtles. The role of magnetite in fish and marine mammals is thought to be navigational (Kirschvink and Lowenstam, 1979; Kirschvink et al., 2001; Walker et al., 2003). As for bacterial magnetite, these biogenic magnetite grains are often in the SD grain size range, the optimal grain size range for a stable magnetic remanence and usually pure magnetite. The grains acquire a so-called chemical remanent magnetization (CRM) as they grow through their “critical” volume within the host organism.

Biogenic magnetite has been commonly observed in marine sediments on the basis of particle shape and size (e.g., Kirschvink and

![Figure P23](image-url)
Chang, 1984; Petersen et al., 1986; Vali et al., 1987). The small elongate (SD) grain size of biogenic magnetite, typically ~0.1 ± 0.05 μm across (Figure P23), and the resulting high surface area to volume ratio makes these grains susceptible to diagenetic dissolution. A large proportion of the biogenic magnetite in surface sediment may not survive sediment burial. Detrital magnetite (often diagenetic) appears to have greater survivability perhaps due to larger initial grain size and/or less intimate contact with pore waters. Stable primary NRMs in deep-sea sediments are completely associated with PSD phases, rather than SD or MD magnetite. Larger multidomain (MD) magnetite, with dimensions in excess of a few microns, carry a lower coercivity magnetization (than PSD or SD grains) and are more prone to remagnetization during the history of the sediment or sedimentary rock.

Origin of secondary magnetizations

Reduction diagenesis

CRM acquired during sediment diagenesis is the principal process that generates secondary magnetizations in marine sediments. The other important secondary magnetization is viscous remanent magnetization (VRM), the time-dependent acquisition of magnetization by the ambient (usually Brunhes Chron) geomagnetic field. A CRM postdates sediment deposition (usually by an unknown amount of time) and must be differentiated from the primary magnetization if sedimentary magnetic data are to be correctly interpreted. The primary and secondary magnetizations are differentiated in the laboratory by stepwise progressive demagnetization using either temperature or alternating fields. The differentiation can be successful if the two magnetization components respond differentially to the demagnetization process (see Dunlop and Ozdemir, 1997).

A wide range of reactions, many of which are mediated by microbes, occur during deep-sea sediment diagenesis. Studies of pelagic pore waters indicate a sequence of electron acceptors used in the oxidation of sedimentary organic matter: O2, nitrite NO3 , manganese (Mn4+) and sulfite SO3 2– (Dunlop et al., 1993). The utilization of sulfate has been found to be particularly important in the dissolution of magnetite grains in pelagic sediments. Microbial reduction of sulfate (sulfate derived from seawater) yields sulfide ions that combine with the most reactive iron phase often magnetite to produce iron sulfide (Canfield and Berner, 1987). In most marine sedimentary environments, authigenic iron sulfides are dispersed throughout the sediment or associated with organic-rich burrows. In visual core descriptions of deep-sea sediments recovered during the 20-year duration of the Ocean Drilling Program, “pyrite” and amorphous iron sulfides are widespread and these iron sulfides probably originated through microbial sulfate reduction. The initial products are such as mackinawite, and further diagenesis may produce greigite (Fe3S4), pyrite (FeS2) and pyrrhotite (Fe1−xS), where x = 0–0.14). Pyrite is paramagnetic and therefore does not carry magnetic remanence. Pyrrhotite (e.g., Fe0.7S) and greigite, on the other hand, are ferrimagnetic and capable of carrying a stable magnetization.

A secondary magnetization (CRM) carried by authigenic pyrrhotite often develops at the expense of a primary magnetization carried by magnetite. The dissolution of magnetite will be accentuated where sufficient pore water sulfate and labile organic matter support microbial sulfate reduction. In organic rich, reducing, diagenetic conditions, such as the Japan Sea, pyrite and pyrrhotite are readily formed (Kobayashi and Nomura, 1972), at the expense of magnetite and other forms of reactive iron. Iron sulfide formation in marine sediments is limited either by the availability of pore-water sulfate (usually derived from sea-water) or the availability of organic carbon to sustain sulfate-reducing bacteria (Canfield and Berner, 1987). On the Ontong Java Plateau, increased C34 to glacial isotopic stages has led to enhanced magnetite dissolution (Tarduno, 1994). In Taiwan, Pleistocene sediments contain detrital magnetite, authigenic greigite and pyrrhotite (Horng et al., 1998). Authigenic greigite has been detected in lake sediments (Giovanoli, 1979; Snowball and Thompson, 1988; Roberts et al., 1996) and in marine environments (Triè et al., 1991; Reynolds et al., 1994; Lee and Jin, 1995).

The dissolution of magnetite within the upper few decimeters of hemipelagic sediments has been documented by a reduction in coercivity and magnetization intensity as finer grains undergo preferential dissolution (Karlin and Levi, 1983; Leslie et al., 1990). In pelagic, high sedimentation rate, “drift” sediments from the Sub-Antarctic South Atlantic (see Figure P24a) and Iceland Basin (see Figure P24b), the pore water profiles indicate steady sulfate depletion with depth.
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Smectite is widespread in deep-sea sediments, there are a number of different pathways for its formation and magnetite is one of a number of different possible precursors.

Degradation of the primary magnetic signal in deep-sea sediment cores can be attributed to factors associated with drilling and recovery. For example, magnetite in calcareous oozes from ODP Leg 154 (Ceara Rise, equatorial Atlantic) carry a secondary magnetization imposed by drilling that is apparently oriented radially in the core cross-section (Curry et al., 1997). No primary magnetization was resolved in these sediments. Magnetite grains are, however, not apparently greatly affected by sulfate reduction as pore water sulfate remains high (> 20 mM) to ~200 m depth and magnetic susceptibility does not decrease with depth (Curry et al., 1997; Richter et al., 1997). Low activity of sulfate reducing microbes may be due to low levels of (labile) organic matter in these sediments. The magnetic susceptibility record is not affected by diagenetic dissolution or authigenesis of magnetic minerals, and the cycles in susceptibility, produced by variations in surface water productivity, have provided a robust astrochronology (Shackleton et al., 1999). The fidelity of the cyclostratigraphy supports other rock magnetic data (Richter et al., 1997), indicating that magnetite has remained unaltered in ODP Leg 154 sediments. The magnetite grains are at least partially in the pseudo-single domain (PSD) size range (Richter et al., 1997) and therefore capable of carrying stable primary magnetization. The observed radial (re)magnetization is attributed to an isothermal remanence (IRM) imposed by the coring procedure (Curry et al., 1997). In pelagic and volcaniclastic sediments of

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**Figure P24** Volume susceptibility, NRM intensity after demagnetization at peak fields of 25 mT, anhysteretic susceptibility divided by susceptibility ($k_{max}/k$), and pore water sulfate concentrations: (a) ODP Site 1089 (sub-Antarctic South Atlantic at 40.9°S, 9.9°E, 4620 m water depth) and (b) ODP Site 984 (Iceland Basin at 61.4°N, 24.1°W, 1648 m water depth). Data from Stoner et al. (2003), Channell (1999), and Shipboard Scientific Party (1996, 1999). At both sites, the 90 m composite depths (mcd) shown here represent ~600 kyr, at mean sedimentation rates of 15 cm ka⁻¹.
ODP Leg 157, from off the Canary Islands, PSD magnetite appears to carry an inward-directed radial magnetization (Herr et al., 1998; Fuller et al., 1998). The coercivity of this radial remagnetization is greater than that of a simple IRM acquired in the few tens of mT fields associated with the steel core barrels and the remagnetization is more pronounced in poorly lithified sediments (Fuller et al., 1998). We speculate that the nonmagnetic matrix in these sediments, when disturbed or shocked by drilling, allows physical re-orientation of magnetite grains in ambient magnetic fields emanating from the core barrels, bottom hole assembly or cutting shoe.

At any time during the history of the sediment or sedimentary rock, secondary magnetizations may be acquired (as CRMs) by chemical alteration of existing magnetic minerals, or by growth of new magnetic minerals. Apart from the diagenetic changes noted above, events such as uplift, weathering, and deformation can all trigger the development of secondary CRMs. For example, much of the Paleozoic sequence of North America and Europe was remagnetized coeval with the Hercynian (Late Carboniferous) orogenic pulse. This “orogenic” remagnetization is very widespread in North America, extending thousands of kilometers into the continental interior from the Appalachian margin. In carbonate rocks in North America and Europe, the Hercynian remagnetization is a CRM carried by magnetite. Authigenesis of magnetite may be triggered by orogenic activity in the Appalachians and the migration of hydrocarbon rich fluids into the hinterland (McCabe et al., 1983, 1989; McCabe and Channell, 1994). Weathering and uplift can result in the growth of magnetic iron oxides and oxyhydroxides (such as hematite and goethite) from preexisting Fe-bearing minerals such as clays or iron sulfides. The high coercivity of CRMs carried by hematite and goethite makes them difficult to remove by alternating field demagnetization techniques.

Conclusions

On the basis of a survey of the global distribution of high-quality magnetic stratigraphies in deep-sea sediments, Clement et al. (1996) concluded that terrigenous sediment input is a factor that contributes to the data quality. The North Atlantic and the North Pacific, and the Indian Ocean, have received abundant terrigenous supply that appears to have aided preservation of a primary magnetization in these sediments and hence enhanced the magnetostratigraphic records from these regions. Although biogenic magnetite is ubiquitous in modern deep-sea sediments, the fine SD particles are particularly susceptible to dissolution due to their high surface area to volume ratio. Much
PALEOMAGNETISM, DEEP-SEA SEDIMENTS


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