Introduction: Magnetic Susceptibility and Environmental Magnetism in Paleoceanography

Magnetic susceptibility (MS) is one of the most commonly employed proxies in paleoresearch and is routinely measured on materials collected by the Integrated Ocean Drilling Program (IODP, www.iodp.org) and international piston coring cruises. MS measurements are not restricted to the marine realm and are commonly employed in terrestrial drilling programs like the Antarctic Geological Drilling Program (Andrill, www.andrill.org) and the International Continental Drilling Program (ICDP, www.icdp-online.org), as well as many lake coring terrestrial stratigraphic, Quaternary geological, and archaeological projects. The widespread use of MS reflects its sensitivity as an indicator of lithologic variation, coupled with the practicality and versatility of measurement. MS measurements can be made on many types of samples, from sediment cores and boreholes to discrete samples of any shape or even directly on an outcrop. Setup costs are low and running costs are negligible, data acquisition is fast, sample preparation is minimal, and as it is both nondestructive and does not interfere with other measurements, it is often the first measurement made in many studies. These factors make MS uniquely applicable to a host of problems, yet MS is still poorly understood by many in the Quaternary and geological communities. Even for those who specialize in magnetism, determining what exactly drives MS variations within any particular record is a nontrivial undertaking. MS is just the simplest of many measurements made in many studies. These factors make MS uniquely applicable to a host of problems, yet MS is still poorly understood by many in the Quaternary and geological communities. Even for those who specialize in magnetism, determining what exactly drives MS variations within any particular record is a nontrivial undertaking. MS is just the simplest of many measurements that make up the tool kit known as environmental magnetism (Thompson and Oldfield, 1986) and often many other magnetic measurements are needed to truly understand what drives MS changes (e.g., Table 1). Here we review and assess both the potential and the opportunities MS and environmental magnetism have for deciphering the Quaternary paleoceanographic record.

What Is MS?

All substances respond to a magnetic field, the question is by how much and in what direction. MS is the relationship between an induced magnetization (M) to an applied magnetic field (H), or simply how magnetizable a material becomes when placed in an external magnetic field.

Low-field (∼0.1 mT) volumetric MS (κ) is the type typically measured in most paleostudies:

\[ \kappa = \frac{M}{H} \]

where \( M \) is the magnetic moment induced per unit volume (amperes per meter; Am\(^{-1}\)) and \( H \) is the applied field (Am\(^{-1}\)). As both \( M \) and \( H \) have the same systeme international d’units (SI) units, \( \kappa \) is dimensionless. Centimeter–gram–second (cgs) units are also dimensionless, where 4π SI = 1 cgs. Unfortunately, it is often not explicitly stated which system is used in many MS studies, making comparisons difficult. To exclude density-related effects and permit comparison between diverse samples, volumetric susceptibility can be density-corrected to produce mass-normalized MS values (\( \chi \)).

\[ \chi = \frac{\kappa}{\rho} \]

where \( \rho \) is the sample density. SI units of \( \chi \) are m\(^3\) kg\(^{-1}\); cgs units are emu Oe\(^{-1}\) g\(^{-1}\).

Influences on MS Values

Different materials, their concentration, and their magnetic grain size and shape, can all affect the MS values of any sample. Resulting from the orbital and spin motions of electrons and their interactions, all materials produce a magnetic response when exposed to a magnetic field, though the nature and strength of this response varies between materials. In the transition metals (Fe, Co, Ni), and their compounds, the magnetization can be particularly strong with parallel (but unequal) coupling of unpaired electrons within a crystal lattice producing a strong magnetization and ferromagnetic or ferrimagnetic behavior (Figure 1; Table 2). True ferromagnetism only exists in the elemental state of these transition metals. As these elements are commonly oxidized (or reduced) into various ferrimagnetic forms in the environment, ferromagnetic behavior is extremely rare in natural samples. Common ferrimagnetic minerals include magnetite, maghemite, and greigite (Table 2). These possess strong positive magnetic susceptibilities and, when present, these minerals will often dominate the MS signal (Figure 1). When atomic arrangements are equal, but opposite, there is no (theoretical) net magnetization as parallel pairings cancel out (antiferromagnetism). However, for a variety of reasons, alignment may not be perfectly parallel in antiferromagnetic minerals and a residual magnetization may result, inducing canted antiferromagnetic behavior (Figure 1). Hematite and goethite are common canted antiferromagnetic minerals possessing weak positive susceptibility (Figure 1; Table 2). Paramagnetic and diamagnetic behavior (Figure 1; Table 2) is induced in samples that have some or no unpaired electrons, respectively. Unlike ferro/ferrimagnetic and canted antiferromagnetic behavior, any induced magnetization in a paramagnetic or diamagnetic material is nonpermanent and lost upon removal of the field. Common examples of paramagnetic magnetic materials are ilmenite, biotite, lepidocrocite, siderite, and pyrite, all exhibiting weakly positive magnetic susceptibilities. Common diamagnetic materials include quartz, calcite, and water with weakly negative magnetic susceptibilities (Figure 1; Table 2). In the absence of ferrimagnetic minerals, paramagnetic or diamagnetic minerals may drive the MS signal.

In many environments, the concentration and type of magnetic minerals are often the dominant controls on MS.
### Table 1

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Notes</th>
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<tbody>
<tr>
<td>$\kappa$, $\chi$</td>
<td>Magnetic susceptibility (MS): the ratio of magnetization induced to the intensity of the magnetizing field. Measured within a small (~0.1 mT) reversible magnetic field either volumetrically ($\kappa$) or on a mass-normalized basis ($\chi$). Roughly proportional to the concentration of ferrimagnetic minerals within a sample. Units: $\kappa$ (dimensionless), $\chi$ (m$^3$ kg$^{-1}$)</td>
</tr>
<tr>
<td>$\chi_{fd}$</td>
<td>Frequency-dependent susceptibility: variation of MS with frequency. Measured at 0.46 and 4.6 kHz on a Bartington MS2B. Indicative of superparamagnetic (SP) grains around the SP/single domain (SD) boundary (~0.03 μm). At higher frequencies ($\chi_{hf}$), grains at the SP/SD boundary respond as SD grains; thus, high frequency MS is proportionally lower than low frequency ($\chi_{ld}$) MS. Units: expressed as a % of $\chi_{ld}$</td>
</tr>
<tr>
<td>ARM</td>
<td>Anhysteretic remanent magnetization: acquired when a sample is placed in a decreasing alternating magnetic field (e.g., 100 mT) in the presence of a small steady biasing field (e.g., 0.1 mT). Sensitive to both ferrimagnetic concentration and ferrimagnetic grain size.</td>
</tr>
<tr>
<td>(S)IRM</td>
<td>(Saturation) isothermal remanent magnetization (SIRM): magnetic remanence induced in a sample upon removal of the field. Usually attained in increasing field strengths (e.g., IRM 20 mT, IRM 100 mT) up to the saturating field (SIRM), which is usually an arbitrary term for largest field generated in the laboratory (~1000 mT). SIRM is an indicator for the concentration of magnetic minerals but also responds to grain size variations. Units: Am$^2$ kg$^{-1}$</td>
</tr>
<tr>
<td>SIRM/$\chi$</td>
<td>This ratio can be diagnostic of either mineralogy, or if mineral type and concentrations are similar, magnetic grain size. Units: Am$^{-1}$</td>
</tr>
<tr>
<td>ARM x mT/SIRM</td>
<td>This ratio is dominantly used to determine magnetic grain size variations, but it is also somewhat sensitive to concentration. Units: Am$^{-1}$</td>
</tr>
<tr>
<td>IRM x mT/SIRM</td>
<td>Ratio of remanence acquired in a smaller field than SIRM to the SIRM. Different field sizes can reveal different aspects of the magnetic signal independent of magnetic concentration. For example, low fields such as 20 mT are sensitive to coarse-grained magnetite, whereas acquisition at higher fields (e.g., 300 mT) can indicate hematite contributions. Units: dimensionless</td>
</tr>
<tr>
<td>(Bo)cr</td>
<td>Demagnetization parameters: obtained by applying reversed magnetic fields to a previously saturated sample. The coercivity of remanence (Bo)cr is the reverse field (mT) at which the SIRM returns to zero. Loss in other fields SIRM-x mT/SIRM can be expressed as a ratio. Loss at a field of 100 mT (which discriminates between ferrimagnetic and canted antiferromagnetic minerals) has been termed the S-ratio. Units: (Bo)cr (mT), SIRM-x mT/SIRM, and the S-ratio are dimensionless</td>
</tr>
</tbody>
</table>

However, magnetic grain size and shape may also influence the MS signal. Ferrimagnetic grains organize their magnetization into different regions, or domains, to achieve minimum energy states; this results in a grain size dependence of MS values. For example, in magnetite ultrafine particles known as superparamagnetic (SP) grains (<0.03 μm) and larger particles known as multidomain (MD) grains (>~10 μm) possess higher MS values than smaller single domain (SD) and pseudo SD (PSD) grains in the 0.03 to ~10 μm size range (Figure 2). Anisotropy also affects MS values as not all grains are spherical. Magnetic energies favor ordering along the easiest direction, the long axis of elongate grains, with grain shape having as much as a factor of 2 influence on MS. For more information on the factors affecting MS measurement, see Thompson and Oldfield (1986), Dearing (1999), Evans and Heller (2003), and Tauxe (2010).

### MS Measurement

MS is the ratio of the induced magnetic moment to the applied field. Changes to the strength and frequency of the applied field and the temperature at which it is measured induce non-linear magnetic responses in a sample. These differences can provide information about the composition of the materials carrying the magnetization, which is important for their interpretation.

Most common measurements of MS are made in low fields (~0.1 mT), which do not permanently magnetize the sample, allowing them to return to their original magnetic state on removal of the field. When ferrimagnetic minerals dominate the mineral assemblage, low-field MS is often assumed to be proportional to their concentration. In greater field strengths, a variety of different minerals and magnetic grain sizes can contribute to the MS signal. However, above a certain critical field, the induced magnetization becomes nonreversible and a magnetization, or remanance, is retained by the sample. Measurement of the magnetization in different field sizes permits construction of remanence parameters and ratios (when measured in zero field) (i.e., isothermal remanent magnetization (IRM; Table 1)) and construction of hysteresis loops (when measured infield) that can be diagnostic of magnetic mineralogy and/or magnetic grain size. MS is dependent on the frequency of the applied field, which has some magnetic grain size dependence. Many systems allow measurements to be made at variable frequencies providing information on the proportion of ultrafine SP grains. MS also displays strong temperature dependence, which is a function of intrinsic magnetic mineralogical properties. Regular measurements made through warming or cooling cycles can produce characteristic curves indicative of different magnetic mineral or grain size populations. In summary, MS is an aggregated function of magnetic concentration, mineralogy, and grain size and...
shape that is dependent upon the applied field, frequency, and temperature of measurement.

**MS Measurement Systems**

MS ($\kappa$ or $\chi$) can be measured on a variety of field and laboratory equipment providing MS measurements of discrete samples, whole cores, split cores, u-channels, or in situ field samples, and can be made in varying field strengths, frequencies, and at different temperatures. For all systems, the magnetic susceptometer produces a magnetic field that induces and measures an infield magnetic moment in response to these three factors.

Due to cost, ease of use, and versatility, the Bartington MS2 is often the most widely available and used MS measurement

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system. Capable of parts per million (ppm) detection of magnetite, it can be coupled with a variety of sensors to allow initial characterization of the magnetic properties of a sample. These measurements are most commonly made at low field (80 Am\(^{-1}\)), low frequency (0.47 kHz), and room temperature.

For example, MS2C loop sensors coupled with measurements of \(g\)-ray porosity, p-wave velocity, and electrical resistivity on multisensor tracks are a standard shipboard measurement on both coring and drilling operations alike. Standard 1.5-m-long core sections can be characterized in as rapidly as 10 min in some configurations, providing real-time data to aid core correlation, site selection, and for initial sediment interpretation. Such measurements should account for the background MS before and after measurement to help correction for local environment and drift complication. This can be easily corrected for when measuring discrete samples as the background can be recorded after

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### Table 2


<table>
<thead>
<tr>
<th>Magnetic behavior</th>
<th>Crystal arrangement</th>
<th>Minerals</th>
<th>MS</th>
</tr>
</thead>
<tbody>
<tr>
<td>True ferromagnetism</td>
<td>Parallel coupling of all unpaired electrons to the applied field</td>
<td>Fe – iron</td>
<td>276,000</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Co – cobalt</td>
<td>204,000</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Ni – nickel</td>
<td>68,850</td>
</tr>
<tr>
<td>Antiferromagnetism</td>
<td>In oxides of Fe, Co, and Ni, electron coupling occurs via an intermediate oxygen atom with alternate crystal layers aligning opposite to each other resulting in no net magnetic moment</td>
<td>Fe(_2)O(_3) – magnetite</td>
<td>660</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(\gamma)Fe(_2)O(_3) – maghemite</td>
<td>440</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Fe(_3)S(_4) – gregite</td>
<td>170</td>
</tr>
<tr>
<td>Ferrimagnetism</td>
<td>If the magnetic moments of the alternating iron compounds are unequal, then a magnetic moment is created</td>
<td>Fe(_3)O(_4) – magnetite</td>
<td>660</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(\gamma)Fe(_3)O(_4) – hematite</td>
<td>(~0.7)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(\alpha)Fe(_2)O(_3) – goethite</td>
<td>(~0.4)</td>
</tr>
<tr>
<td>Canted antiferromagnetism</td>
<td>For various reasons, sublattices in otherwise antiferromagnetic arrangements may not be perfectly aligned and a small residual magnetization exists</td>
<td>(\alpha)Fe(_2)O(_3) – hematite</td>
<td>(~0.7)</td>
</tr>
<tr>
<td>Paramagnetism</td>
<td>Some unpaired electrons align to an applied field; this magnetization is lost on removal of the field</td>
<td>Fe(_3)O(_4) – magnetite</td>
<td>660</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Mg,Fe,Al silicate – biotite</td>
<td>(~0.5)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(\gamma)Fe(_2)OOH – lepidocrocite</td>
<td>0.7</td>
</tr>
<tr>
<td></td>
<td></td>
<td>FeCO(_3) – siderite</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Fe(_2)S(_2) – pyrite</td>
<td>0.3</td>
</tr>
<tr>
<td>Diamagnetism</td>
<td>Compounds with no unpaired electrons act to repel the applied field; this magnetization is lost on removal of the field</td>
<td>Si(_2)O(_3) – quartz</td>
<td>(~0.0058)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>CaCO(_3) – calcite</td>
<td>(~0.0048)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>H(_2)O – water</td>
<td>(~0.009)</td>
</tr>
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Units of mass-normalized magnetic susceptibility (MS) – \(10^{-6}\) m\(^3\) kg\(^{-1}\).

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**Figure 2** Domain-size boundaries and variation in MS as a function of magnetic grain size for magnetite. Domain abbreviations: SP = superparamagnetic, SD = single domain, PSD = pseudo-single domain, MD = multi domain. Data from Maher (1988), Ozdemir and Banerjee (1982) and Dankers (1978).
Compounds of iron comprise roughly 5% of the Earth’s crust and are ubiquitous in many environments in a variety of inorganic and bioavailable forms (Table 2); for example, iron oxides can constitute 2–6% of basalts (Thompson and Oldfield, 1986). Iron-bearing minerals are often eroded, transported, and deposited coevally with other minerals, nutrients, or pollutants. Thus, magnetic properties of environmental materials often correlate well with other independently determined properties. In many sedimentary environments, magnetite is the most common ferrimagnetic mineral. Given its strong response to a magnetic field (Table 2), low-field MS is often considered proportional to magnetite concentration (Dearing, 1999). However, although mineral concentration is often considered the principal factor driving MS, this is often not the case and these assumptions are not always valid. Natural samples are rarely unimodally composed and instead are aggregates of varying amounts of different minerals and grain sizes, complicating such a simple interpretation. As magnetic properties are not uniform across magnetic mineralogies or grain sizes, the interpretation of what drives MS variation is often not simply a reflection of ferrimagnetic concentration. Similar values can potentially be driven by different compositions, for example, the presence of a small amount of magnetite or larger concentrations of hematite. MS can be diluted by diamagnetic minerals (e.g., biogenic carbonates) that are often significant components of ocean sediments. The nature of the magnetic carriers also can affect the MS. For example, authigenic bacterial magnetosomes are composed of SP to SD magnetite, while terrigenous inputs to marine systems are often PSD to MD grain sizes. The interplay of all of these factors can sometimes produce different and unique MS values characteristic of a specific environment or process. However, it is rarely possible to determine the factors driving the MS signal from MS measurements alone; a greater suite of magnetic measurements are normally necessary (e.g., Table 1). Such an evaluation is almost always required, not only to identify the absolute drivers of change, but also to recognize the causes of differences between measurements and to help with the understanding and accounting for the potential processes involved in driving these changes.

However, given the abundance of Fe oxides in the environment, their ability to harmonize with natural processes, and the ability of magnetic measurements to detect subtle changes in concentration, mineralogy, and grain size, links are often drawn between measurements of MS and environmental change in many diverse settings. Lacustrine MS records often show strong linkages to other independent climate proxies, and compared to the deep-sea environment, lake records possess relatively high sedimentation rates, enabling high-resolution reconstructions of climate. However, many of the most climatically sensitive lakes are found in glacially proximal environments and are thus temporally restricted due to their postglacial formation. Similarly, loess records have shown to possess remarkable MS archives of climate changes; however, their formation is quasi-continuous and restricted to certain areas of the globe. In contrast, relatively slow but continuous sedimentation rates in marine records permit the potential for long climate change reconstructions from large areas of the globe.

**Environmental Records of Susceptibility**

Like all other environmental records of susceptibility, the MS records of marine sediments are an aggregated sum of the interplay of magnetic concentration, magnetic mineralogy, and magnetic grain size. Variations in these properties can be forced by, and often be characteristic of, several different depositional processes and postdepositional (diagenetic) transformations. Depositional processes generally reflect the balance between terrigenous flux, modulated by some combination of surface and bottom currents, eolian transport, ice-rafted and other glacial marine processes, volcanogenic inputs, tectonic and downslope transport, and biogenic preservation. Postdepositional alterations may result in the creation, destruction, and/or transformation of magnetic minerals. All these factors can influence an MS signal and their relative influence can vary both spatially and temporally. Unraveling these often codependent influences is a nontrivial matter, but with careful selection of both site location and time period, coupled with analysis and understanding of the factors driving the MS signal, MS records of ocean sediments can potentially reveal much
about the paleohistory of both oceans and climate. Here we will report some of the important ways MS has been used in the study of paleoceanographic and paleoclimatic records of ocean cores and some of the important factors to consider in their interpretation.

**MS as a Proxy for Terrigenous Flux**

The delivery and distribution of terrigenous sediments to the world’s oceans is subject to a number of complex processes. One of the major factors governing terrigenous sediment concentration in the ocean is the availability and transport of detrital grains. Transport and delivery-related processes can be, and have been, climatically mediated over a range of timescales in response to changes in climatic regimes. In environments with minimal diagenetic transformations, the MS records of ocean cores can simply reflect (and are often assumed to reflect) the composition of the terrigenous material delivered to a site. Given that the production of biogenic materials (e.g., carbonate and silica) can also be climatically dependent, the MS records of many marine sediments often reflect the balance between terrigenous flux and biogenic dilution. Common examples of this type of behavior are found in Pacific, Indian, and Atlantic Oceans.

The MS of ODP sites 721 and 722 in the Arabian Sea strongly correlate with the percentage of terrigenous materials ($r = 0.98$; Bloemendal and deMenocal, 1989). Delivery of this terrigenous fraction is linked to eolian transport and monsoon intensity. Spectral analysis of the Arabian Sea MS records and ODP site 661 from the eastern equatorial Atlantic showed strong influence of the 23 and 19 ka precession-related orbital periodicities prior to 2.4 Ma with a shift to 41 ka periodicities after 2.4 My, possibly reflecting ice sheet effects on monsoon climate or dust source regions (Bloemendal and deMenocal, 1989; Figure 3). Similarly for ODP site 882 in the North Pacific, Maslin et al. (1996) used MS as a proxy for ice-rafted debris (IRD), and showed synchronous changes between IRD, the $\delta^{18}O$ record, and the biogenic opal record (Figure 4). Despite the different processes responsible for delivering sediment to these two sites, both MS records identify a major climate reorganization 2.4–2.7 Ma, resulting in global changes in terrestrial flux from the continents. It is thought that these shifts are associated with the initiation of northern hemisphere glaciations and the start of the Quaternary. These long records linked to climatically mediated terrestrial fluxes helped establish the links between orbital and climatic forcings of MS records.

Greater sedimentation rates in the North Atlantic, coupled with greater carbonate preservation, have permitted higher resolution studies of individual events, with their chronologies constrained by more accurate dating. Surrounded by a terrestrial regime dominated by the growth and decay of continental-scale ice sheets, bedrock geology composed of both highly magnetic basalts and crystalline basement rocks and weakly magnetic carbonates, and a sediment transport regime dominated by dynamic surface and subsurface circulations, the North Atlantic has proven to be an exceptional archive for MS studies of paleoclimate and paleoceanography.

In the mid-latitude IRD belt (40° and 55° N) of the glacial North Atlantic (Ruddiman, 1977), MS has been widely used to correlate IRD events between cores, illustrating the basin-wide coherence of the IRD pattern. Through glacial and interglacial cycles, Robinson (1986) showed that MS peaks were driven by fluxes of IRD in the mid-latitude North Atlantic, linked to increased glacial calving. This signal is also mediated by productivity (Figure 5). During glacial periods, increased delivery of IRD is coupled with a reduction in the production of biogenic carbonates resulting in high MS values. During interglacials, decreased delivery of IRD is coupled with increased biogenic production resulting in increased diamagnetic carbonates resulting in low MS values. This flux/dilution relationship results in strongly contrasting MS values on glacial/interglacial timescales that are replicable across large spatial scales. The highest glacial susceptibility peaks are associated with sand layers (>150 μm) deposited during Heinrich events (Bond et al., 1992), providing a diagnostic signature of these canonical Laurentide Ice Sheet discharge events that prominently characterize the glacial North Atlantic (e.g., Bond et al., 1992). MS records south of the glacial polar front in the North Atlantic differ greatly from those within the IRD belt, reflecting the strong melting gradient and IRD-forced nature of MS. Mapping of these spatial MS signatures also permitted reconstruction of iceberg trajectory paths, paleosurface currents, and the migration of the glacial polar front.

From the Southern Ocean, Pugh et al. (2009) show MS records from three cores from the Scotia Sea to be almost identical to the European Project for Ice Coring in Antarctica (EPICA) dust record (Figure 6). Again, glacial periods are characterized by high susceptibility and interglacials with susceptibility minima. However, these records are apparently not driven by IRD, but by dust export from continental landmasses. The processes involved in driving ice rafting in the North Atlantic and dust flux in the Southern Ocean are very different; however, MS records show similar trends. This similarity is forced by shifts in dominant global climatic patterns over glacial/interglacial timescales, which in turn mediates the terrestrial flux and the attendant MS record.

While global climate changes can have a large impact on global sediment flux and the MS record, it is essential to interpret any MS record within the context of its location and processes acting in its formation. For example, Stoner et al. (1995) and Carlson et al. (2008) examined the magnetic and geochemical record of cores from the Eirik Drift in the NE Labrador Sea, southwest of Greenland. Unlike the majority of records from further south in the North Atlantic Basin, glacial terminations and interglacial periods are characterized by higher MS values than glacial periods. During the glacial–interglacial transitions of marine isotope stage (MIS) 2/1 and MIS 6/5, enhanced MS is accompanied by a coarsening of magnetic grain sizes and an increase in Ti/Fe concentrations. Here, minima in IRD (represented by the >125 μm fraction) and increases in the silt fraction are synchronous with increases in MS and magnetic grain sizes and are thought to be associated with sediment-laden melt water pulses associated with increased melting of the Greenland Ice Sheet during deglaciation (Carlson et al., 2008; Stoner et al., 1995). Thus, while these Greenland proximal records, like the North Atlantic IRD and Southern and Indian Ocean eolian records are driven by flux of terrigenous fragments responding to changing climate drivers, their MS signals produce very different temporal signatures.
Figure 3  Magnetic susceptibility time series for ODP Site 661 and ODP site 721, and the variance spectra (inset figures) for time intervals between 3.5 Ma and the present. Adapted by permission from Macmillan Publishers Ltd: [Nature] (Bloemendal J and deMenocal P (1989) Evidence for a change in the periodicity of tropical climate cycles at 2.4 Myr from whole-core magnetic susceptibility measurements. Nature 342: 897–900), copyright 1989.
MS records driven by the flux of terrestrial fragments can yield high-resolution information about changes in climate and its effect on the terrestrial landscape. The glacial/interglacial, flux/dilution relationship in the ocean remains a major driver behind many MS records and permits comparisons over local, regional, and to some degree even global scales. However, these records must first be interpreted within the context of the processes driving these records and other factors which may potentially modify this relationship must also be assessed.

**MS as a Proxy for Sediment Redistribution**

In the deep ocean, MS records often correlate well with IRD percentage and reflect surface currents and iceberg melt routes. However, Rasmussen et al. (1997) showed a remarkable similarity between the MS record of ENAM93-21 from the Faeroe–Shetland Channel in the North Atlantic and the $\delta^{18}$O of the Greenland Ice Core Project (GRIP) record (Figure 7). This MS record captured all 15 Dansgaard–Oeschger (D–O) cycles within the MIS 3 and MIS 2 intervals. In contrast to the North Atlantic IRD pattern, abrupt warming is characterized by sharp increases in MS, with gradual cooling associated with decreasing MS values, showing little relation to IRD inputs (Figure 7(c)) either in the model proposed by Robinson (1986) or regionally. MS lows are associated with colder stadial periods and Heinrich event intervals. Subsequent studies have found similar patterns along the path of dense overflow waters in the subarctic basins of the North Atlantic that contribute to North Atlantic Deep Water (NADW). Because these overflow waters have been implicated as drivers of the D–O cycles observed in the Greenland ice cores, these MS observations have implications for examination of the processes involved. MS patterns were shown to be strongly controlled by variations
Figure 5  Correlation of cores using MS through seven glacial/interglacial stages. Reprinted from *Physics of the Earth and Planetary Interiors*, 42, Robinson SG, The late Pleistocene paleoclimatic record of North Atlantic deep-sea sediments revealed by mineral-magnetic measurements, 22–47, Copyright (1986), with permission from Elsevier.

Figure 6  Correlation between EPICA dust concentration and three Scotia Sea cores. Note the high susceptibility in glacial phases and low MS during interglacials, similar to the IRD record. Reprinted from *Earth and Planetary Science Letters*, 284, Pugh R, McCave I, Hillenbrand C, and Kuhn G, Circum-Antarctic age modelling of Quaternary marine cores under the Antarctic Circumpolar Current: Ice-core dust-magnetic correlation, 113–123, Copyright 2009, with permission from Elsevier.
in titanomagnetite concentration (Kissel et al., 1997; Snowball and Moros, 2003) and similar patterns were shown to be spatially replicable over a transect of several cores (e.g., Kissel et al., 2009). It is suggested that during colder phases when NADW formation was sluggish, transport and delivery of ferromagnetic magnetite to drift sites was restricted resulting in lower MS values. Conversely, during warmer phases, NADW strength increased, as did basaltic fragment transport, increasing the MS values at these locations.

Peaks and minima of MS in oceanic sediments can often be interpreted in a paleoclimatic context. However, resuspension during downslope turbidite flows and subsequent density-related settling of relatively heavy magnetic minerals within an otherwise magnetically weak matrix can also produce characteristic signatures. These MS peaks can be important for identifying turbidite layers and for distinguishing them from paleosignals, as these events can have severe implications for the paleointerpretation of sedimentary records (e.g. core S8-79-6 in Figure 5). MS peaks generated by these events can often be process and event diagnostic. Goldfinger et al. (2007, 2008) used MS to study paleoseismology, correlating MS peaks between earthquake-event-triggered turbidite flows over several hundred kilometers along the Cascadia margin. Earthquake-generated turbidite deposits have not only been recorded by magnetic measurements on the Pacific margins, but also in Quebec and off the Portuguese Coast, illustrating the potential of MS for identifying these deposits.

MS and Source Variation

IRD, bottom current, and slope process records of MS often assume that magnetic concentration linked to flux/dilution relationships are the principal drivers behind the MS record. However, Pirrung et al. (2002a,b) showed that MS varied over three orders of magnitude in the surface sediments of both the Nordic Seas (Pirrung et al., 2002a) and the Atlantic sector of the Southern Ocean (Pirrung et al., 2002b), linked to the source of minerals from different provinces (Figure 8). Highest values were found around Scoresby Sund and along the Iceland–Faeroe–Shetland Ridge, linked to basaltic sources in the Nordic Seas; similarly, in the Southern Ocean, highest values are proximal to mafic sources. It is often assumed that all ferrimagnetic minerals are terrigenously derived, yet all terrigenous materials are not ferrimagnetic. In the absence of a strong ferrimagnetic source component, MS values may be influenced by weaker magnetic components; such areas are proximal to mafic sources. It is often assumed that all ferrimagnetic minerals are terrigenously derived, yet all terrigenous materials are not ferrimagnetic. MS peaks generated by these events can often be process and event diagnostic. Goldfinger et al. (2007, 2008) used MS to study paleoseismology, correlating MS peaks between earthquake-event-triggered turbidite flows over several hundred kilometers along the Cascadia margin. Earthquake-generated turbidite deposits have not only been recorded by magnetic measurements on the Pacific margins, but also in Quebec and off the Portuguese Coast, illustrating the potential of MS for identifying these deposits.
Figure 8  MS variation in surface sediments from the Nordic Seas and the Atlantic sector of the Southern Ocean. Reproduced with kind permission from Springer Science+Business Media: Geo-Marine Letters, Magnetic susceptibility and ice-rafted debris in surface sediments of the Nordic Seas: Implications for Isotope Stage 3 oscillations, 22, 2002a, 1–11, Pirrung M, Fütterer D, Grobe H, Matthiessen J, and Niessen F, Figure 1; Geo-Marine Letters, Magnetic susceptibility and ice-rafted debris in surface sediments of the Atlantic sector of the Southern Ocean, 22, 2002b, 170–180, Pirrung M, Hillenbrand C, Diekmann B, Fütterer D, Grobe H, and Kuhn G, Figure 2.
Heinrich layer MS values are also low in this region, illustrating MS dilution due to diamagnetic carbonate background terrigenous concentrations. Instantaneous tephra deposits can also possess distinct magnetic signatures and can be used as stratigraphic markers (e.g., Peters et al., 2010).

Watkins and Maher (2003) and Watkins et al. (2007) extended the work of Pirrung et al. (2002a) to include both potential sources and core sediments surrounding the whole North Atlantic Basin (Figure 9), both at the present day and during the Last Glacial Maximum (LGM). Their results neatly illustrated that even as sediment from the IRD belt (poleward of 40°N) during the LGM possessed higher susceptibility than present (Watkins et al., 2007), there exists significant regional variation in MS, both for the present day and at the LGM. For example, IRD originating from Icelandic basalts and the North American Shield possess characteristically high susceptibilities, and are reflected in the sedimentary properties of the Western North Atlantic. In contrast, the lower susceptibility areas identified by Pirrung et al. (2002a) are proximal to the gray and red sandstones from Spitsbergen which outcrop and surround the Norwegian and Barents Seas, and eolian dust sources are responsible for the relatively low susceptibility of the central Atlantic. With this variation defined, MS can potentially be used to map the timing and intensity of IRD events, and under optimal conditions can provide information on their sources through the last glacial–interglacial transition (cf. Watkins and Maher, 2003; Watkins et al., 2007). However, isolating the relative influences of sediment flux and sediment source is an almost impossible process using MS alone. Instead, a wider range of magnetic measurements are often needed (see Table 1), with combinations of these parameters permitting assessment independent of magnetic concentration. Source variation and the roles of IRD and bottom water currents also rarely act in isolation of each other or indeed several other processes which may deliver or restrict transport of terrigenous fractions. Site selection and magnetic measurement which consider these processes, and an interpretation that accounts for them, are often key to successfully relate the MS signal to a certain aspect of the climate system.

**Postdepositional Records of MS**

Magnetic minerals accumulating in sedimentary environments are often assumed by the paleocommunity to be conservative, maintaining properties that reflect their original source materials and mechanisms of deposition. In areas with high organic matter accumulation and/or in low oxygen environments, organic material may be buried before fully oxidizing, promoting...
bacterially mediated reduction diagenesis (Froelich et al., 1979). Fine ferrimagnetic grains (either of biogenic or terrestrial origin) are most susceptible to reduction in a profile undergoing diagenetic changes. This can result in partial loss of the deposited magnetic signal and is therefore a significant issue for interpretation of MS records. Deeper in these records, soluble iron can be reprecipitated as other magnetic phases, including pyrite and greigite, and these minerals have shown to be significant components of oceanic records. These minerals can nucleate around larger surviving ferrimagnetic grains protecting their core (and magnetic signal) from further reduction.

Oceanic areas producing strong diagenetic change are generally restricted to continental margins and/or where sedimentation and organic accumulation rates are high and in areas of upwelling and in partially landlocked basins (e.g., the Mediterranean and the Sea of Japan). Such areas are generally well known and while assessment of diagenetic change cannot be conclusive using MS alone, simple additional diagnostic magnetic measurements can help document evidence for diagenetic change. Distinctive signatures include steep decreases in concentration indicators (e.g., MS) with a parallel coarsening of parameters sensitive to magnetic grain size pointing to the selective loss of finer ferrimagnetic grains. As hematite and goethite are less affected by suboxic diagenesis, an increase in coercivity may also accompany selective loss of finer grains.

Diagenetic processes usually result in degradation of the depositional signal and the paleoclimatic information it contains. However, reduction-driven diagenetic transformations may not always result from steady-state conditions, but from changing boundary conditions providing additional opportunities for the generation of paleoinformation. For example, in an analysis of four sediment records from the Sea of Japan, light/dark rhythms in the sediment record suggested cyclic switching between times of oxic and anoxic conditions, related to glacio-eustatic sea level changes (Vigliotti, 1997). Here MS correlates with the \( \delta^{18}O \) record, showing sea level-lowered glacial periods are more favorable for production of anoxic basin conditions resulting in an increase in sulfuric compounds and a reduction in magnetic concentration and coarsening of the magnetic grain size. Interglacial periods contain deeper waters, greater basin oxygenation, and greater retention of the fine ferrimagnetic fraction (Figure 10).

Postdepositional transformations act not only to remove ferrimagnetic minerals from a profile, but in situ production of fine-grained authigenically derived bacterial magnetite can also potentially significantly alter the depositional record. Bacterial

![Figure 10](image-url)  
**Figure 10** Sulfur (%) and MS for ODP sites 794 (open symbols) and 795 (full symbols) from the Sea of Japan through 18 marine isotope stages. Reprinted from *Quaternary Science Reviews*, 16, Vigliotti L, Magnetic properties of light and dark sediment layers from the Japan sea: Diagenetic and paleoclimatic implications, 1093–1114, Copyright (1997), with permission from Elsevier.
magnetite chains have been identified as significant remanence carriers in both oxic and anoxic oceanic records (e.g., Kirschvink, 1982; Petersen et al., 1986; Stolz et al., 1990). Mechanisms controlling their formation can also possess certain boundary conditions. Lean and McCave (1998) examined a deep-sea core east of New Zealand, identifying bacterial magnetite phases and proposing climatic mediation of conditions favoring bacterial magnetite formation. However, this record is probably not reflective of all authigenic contributions, especially in areas susceptible to diagenesis. As with fine-grained ferrimagnetic constituents within terrigenous sediment, these phases are rapidly lost as they pass through the upwardly migrating Fe redox boundary by iron-reducing bacteria (Karlin et al., 1987; Robinson et al., 2000). Thus, in general, bacterial magnetosomes typically are only found in the upper few meters of sediment. These organisms produce extracellular SP magnetite, which can be readily distinguished from coarsely grained IRD or high-coercivity eolian dusts by additional magnetic measurements. However, if they go undetected in a sediment record, they have the potential to overprint the depositional magnetic signal which may result in an erroneous interpretation.

Conclusions

The MS record at any one site is the sum of the influences of many factors including flux and subsequent redistribution (by numerous processes, the source, and mineralogy of the terrigenous material, and dilution by diamagnetic materials). This signal can then be modified by processes of the magnetic fraction postdepositionally.

Correlations between MS and other climate proxies result from the ubiquitous nature of iron oxides in marine sedimentary environments and their sensitivity to climatically driven processes. The key is to derive which aspect of the climate system they are recording. Similar records can be produced by different processes, which can lead to erroneous interpretations. Therefore, independent constraints are required when interpreting MS record beyond the local scale. As similar delivery processes can often produce different MS records, postdepositional processes can affect records to different degrees. It is often necessary to couple MS measurements with other simple magnetic analyses to understand the drivers of the MS before attempts are made to extract any paleoinformation.

In summary, MS measurements have proven to be a highly effective tool for the interpretation of paleoenvironmental change, possessing sensitivity to aspects often unresolved by other methods. With good understanding of the reasons driving any MS changes, MS records have become part of a two-way iterative feedback process for the understanding of paleoclimate. We have to understand the climatic drivers before we can interpret the MS signal, but the MS record can provide excellent new insights into paleoclimate that may not be available using other proxies. Ongoing work to address issues of interlab calibration and recognition of the potential for increased resolution through particle-sized specific measurement will increase the robustness and understanding of these techniques.


References


Relevant Websites


www.icdp-online.org – International Continental Drilling Program (ICDP).
