Magnetic methods and the timing of geological processes

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Abstract: Magnetostratigraphy is best known as a technique that employs correlation among different stratigraphic sections using the magnetic directions that define geomagnetic polarity reversals as marker-horizons. The ages of the polarity reversals provide common tie points among the sections, allowing accurate time correlation. Recently, magnetostratigraphy has acquired a broader meaning, now referring to many types of magnetic measurements within a stratigraphic sequence. Many of these measurements provide correlation and age control not only for the older and younger boundaries of a polarity interval, but also within intervals. Thus, magnetostratigraphy no longer represents a dating tool based only on the geomagnetic polarity reversals, but comprises a set of techniques that includes measurements of all geomagnetic field parameters, environmental magnetism, rock magnetic and palaeoclimatic change recorded in sedimentary rocks, and key corrections to magnetic directions related to geodynamics, tectonics and diagenetic processes.

Discovery of geomagnetic reversals

Over the past century numerous methodologies have been developed to detect time variations of the geomagnetic field and environmentally significant magnetic properties in rocks. These methods comprise measurements of natural remanence, magnetic susceptibility, demagnetization and induced artificial magnetizations. Brunhes (1906) and Matuyama (1926) were among the first to recognize that old rocks have inclination values that are very different from today's values, and sometimes of opposite polarity to the present-day magnetic field. According to Matuyama (1926), these changes represent reversals in the polarity of the ancient geomagnetic field. Cox *et al.* (1963) recognized that these polarity reversals were global events, and that, by combining palaeomagnetic and geochronologic data, a sequence of geomagnetic field reversals could be constructed. This led directly to the development of the Geomagnetic Polarity Time Scale (GPTS).

Vine & Matthews (1963), using data acquired during marine cruises, recognized that magnetic anomalies had a symmetrical pattern with respect to the mid-ocean ridges, and that there was a relationship between geomagnetic field reversals and motions of the ocean floor. The oceanic magnetic anomalies are related to the magnetization of the oceanic basalts that cool down while spreading from mid-oceanic ridges, and can be used in combination with geomagnetic field polarity sequences from rocks found on land to further develop a globally extensive GPTS (Heirtzler *et al.* 1968). Opdyke (1972) first integrated magnetostratigraphy and biostratigraphy for Plio-Pleistocene marine sediments, and since that time biostratigraphic information has been increasingly used for correlation of the observed polarity sequences in sedimentary rocks with the appropriate part of the radioisotope-calibrated GPTS. Subsequently, Alvarez *et al.* (1977) recognized sets of magnetic polarity reversals within the Cenozoic Gubbio (Italy) sedimentary sequence. William Lowrie studied the geological and physical processes that permit pelagic sediments to keep magnetization and defined the magnetostratigraphy of those Italian sections (Lowrie *et al.* 1982), allowing the scientific community to build and further refine the GPTS (Cox *et al.* 1963; Heirtzler *et al.* 1968; LaBreque *et al.* 1977; Berggren *et al.* 1985; Cande & Kent 1992, 1995; Huestis & Acton 1997; Singer *et al.* 2002; Channell *et al.* 1995; Malinverno *et al.* 2012; Ogg 2012).

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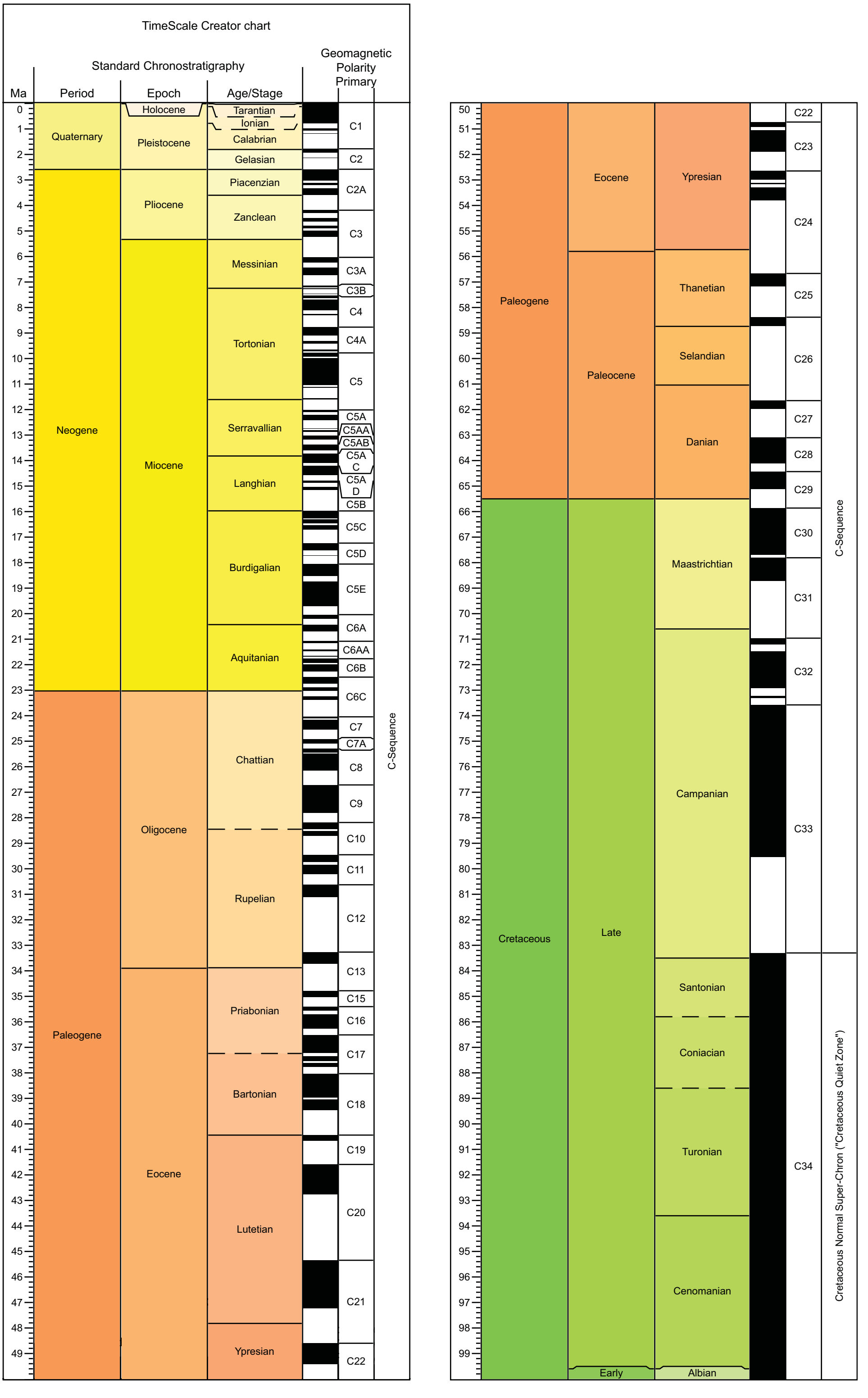
TimeScale Creator chart
Standard Chronostratigraphy
Geomagnetic Polarity Primary
Ma
Period
Epoch
Age/Stage
Quaternary
Neogene
Paleogene
Cretaceous
Holocene
Pleistocene
Pliocene
Miocene
Oligocene
Eocene
Paleocene
Late
Early
Tarantian
Ionian
Calabrian
Gelasian
Piacenzian
Zanclean
Messinian
Tortonian
Serravallian
Langhian
Burdigalian
Aquitanian
Chattian
Rupelian
Priabonian
Bartonian
Lutetian
Ypresian
Thanetian
Selandian
Danian
Maastrichtian
Campanian
Santonian
Coniacian
Turonian
Cenomanian
Albian
C-Sequence
Cretaceous Normal Super-Chron ("Cretaceous Quiet Zone")

Today, geomagnetic reversals are routinely recognized along stratigraphic sections composed of sedimentary (marine or continental) or volcanic materials. To determine a polarity stratigraphy, the sediment or rock must first contain a record of the geomagnetic field that is generally acquired at the time of emplacement. In order to measure the original magnetization that records the geomagnetic field polarity at the time of formation of the rock, sample direction must be measured in the field (i.e. *in situ*). The capacity of a rock to maintain its own magnetic field and resist demagnetization is related to the coercivity of its magnetic minerals. There are different ways by which rocks can record a natural remanent magnetization (NRM) in the presence of an external magnetic field (e.g. the geomagnetic field; Kodama 2012): (1) thermoremanent magnetization (TRM) is acquired when a rock cools down below the Curie temperature of its magnetic minerals; (2) chemical remanent magnetization (CRM) is acquired when a new magnetic mineral grows after the rock is formed and establishes its own magnetization; (3) viscous remanent magnetization (VRM) is attained in an ambient field for magnetic relaxation during time; (4) isothermal remanent magnetization (IRM) occurs in nature when rocks are struck by lightning and are submitted to a magnetic field larger that their coercivity; and, the most important in sediments, (5) detrital remanent magnetization (DRM) is acquired when depositional magnetic grains align themselves with the geomagnetic field as they are settling through the water column or are in unconsolidated sediment. Depositional magnetic grains deposited on the seafloor are then able to lock in and retain the original magnetization in the direction of the geomagnetic field during the initial consolidation of sediments (Tauxe *et al.* 2006; Tauxe & Yamazaki 2007). However, in some deep-sea sediments, a time delay of magnetization has been occasionally observed and is attributed to very low sedimentation rates and delayed lock-in below the sediment–water interface (Verosub 1977; Suganuma *et al.* 2011). This magnetization acquired near but not directly at the time of deposition is referred to as post-depositional remanent magnetization (pDRM). There are also some conditions in which magnetic crystals produced by magnetotactic bacteria may record the geomagnetic field (Petersen *et al.* 1986; Housen & Moskowitz 2006; Vasiliev *et al.* 2008; Yamazaki 2009; Roberts *et al.* 2011, 2012; Jovane *et al.* 2012).

The cause of the geomagnetic field reversals is still unknown, although geodynamical modelling demonstrates the occurrence of reversals (Glatzmaier & Roberts 1995; Kuang & Bloxham 1997). The main theory relates reversals to internal fluid instability of the Earth's outer core. In this region of the core, convective movements and complex vortices are created within tangent cylinders, that is, cylinders that are pretended coaxial movements in relation to the Earth's rotation axis, and tangent to the inner core/outer core boundary with their projection on Earth surface at 79.1° latitude. The thermal and compositional convection processes at the core-mantle boundary influence the geodynamo with long-term variations ($>10^5$ years) of intensity, inclination and declination (Olson *et al.* 2010). This organized motion evolves chaotically with the magnetic field produced by the electromagnetic dynamo growing, decaying and occasionally flipping in the opposite direction (Gubbins & Bloxham 1985, 1987; Bloxham & Jackson 1992; Olson & Aurnou 1999; Jackson *et al.* 2000; Hulot *et al.* 2002; Aurnou *et al.* 2003; Wardinski & Holme 2006). The geomagnetic reversals occur on the entire globe, also near the tangent cylinder and polar regions (Jovane *et al.* 2008). There are also other theories to explain reversals, for example, one in which geomagnetic reversals are linked to extraterrestrial impacts (Muller & Morris 1986).

Magnetostratigraphy and the geomagnetic polarity time scale

At its most fundamental level, magnetostratigraphy documents the geological record of polarity changes of the geomagnetic field. The individual normal (black) and reversed (white) (Fig. 1) polarity intervals are known as chrons and typically range in duration from 10 kyr to 10 Myr. The transition from a reversed-polarity chron to a normal-polarity chron and vice-versa is very short (± 5 kyr). This allows a numerical age to be assigned to each rock unit containing a polarity reversal within a stratigraphic succession. Since polarity reversals effectively occur simultaneously over the whole surface of the Earth, they can be used for global time correlation. Shorter periods of opposing polarity within a chron are called, depending on duration, 'subchrons,' 'microchrons' and 'cryptochrons' (Cande & Kent 1992). Longer periods with dominant single polarity are called 'superchrons' or 'megachrons,' also depending on duration (Opdyke & Channell 1996).

Chrons are conventionally labelled and named after the corresponding seafloor spreading magnetic

Fig. 1. Example of the GPTS for the past 100 million years, from *TSCreator 5.0* (www.tscreator.org). Courtesy of J. Ogg. (2012).

anomaly number (Cande & Kent 1992). Chron number is usually suffixed by the letter *n* or *r*, depending on whether the dominant magnetic polarity is normal or reversed, and is prefixed, for instance, by the letter C for Cenozoic, or M for Mesozoic (Gee & Kent 2007). The most recent reversal of the geomagnetic field occurred 781 kyr ago and is the boundary between Brunhes (C1n) and Matuyama (C1r) Chrons. Other normal and reversed chrons, Gauss (C2n) and Gilbert (C2r), contain distinctive subchrons, such as the Olduvai Subchron (C1r.2n) within the Matuyama Chron. The GPTS is continually updated to provide the most accurately known ages for the chron boundaries (e.g. Fig. 1).

Vector component diagrams (Zijderveld 1967) are used display stepwise demagnetization data. When a series of rock layers shows the same sign of inclination of the characteristic remanent magnetization (ChRM), it is called a magnetozone (or magnetostratigraphic unit), because within this interval a single polarity is constant. Magnetostratigraphy correlates magnetozones to the GPTS. Namely, inclination, declination and intensity of the ChRM for each sample are examined through a principal component analysis (PCA) of demagnetization steps (Kirschvink 1980). Demagnetization can be accomplished in different ways for different magnetic components. Stepwise thermal (TH) and alternating field (AF) demagnetizations are the most popular methods. Chemical and pressure demagnetization methods can also be applied to erase unwanted secondary components.

The calculated PCA inclination values sometimes do not represent the true inclination of the geomagnetic field at that time of the formation or deposition of the rock unit (Butler 1992; Tauxe 2010). This problem is related to geological factors that can be resolved using other magnetic methods. These factors may be related to diagenetic compaction or rotation of tectonic plates and tilt of structural blocks.

Beyond classical magnetostratigraphy

Dating with other recorded features of the geomagnetic field can also be undertaken, as follows.

Excursions and aborted reversals

It is very well known from palaeomagnetic records today that, in addition to polarity changes, the Earth's magnetic field has experienced changes from its regular near-axial configuration for brief periods of time without establishing, and perhaps not even approaching, a reversed state of the palaeofield. These types of behaviour are called geomagnetic excursions or 'aborted reversals'.

Such geomagnetic excursions have been reported in geological recorders such as lava flows of various ages in different parts of the world as well as from deep-sea, lake sediment and sedimentary rocks. This type of geomagnetic feature generally are observed to start with a sudden and often fairly smooth movement of the virtual geomagnetic poles (VGP) toward equatorial latitudes. The VGP may then return almost immediately, or it may cross the equator and move through latitudes in the opposite direction before travelling back again to resume a near-axial position. The term 'excursion' was defined to describe a VGP movement of more that 40° from the geographic pole for intermediate pole positions that end up with a return of the Earth's field to its pre-existing polarity (Barbetti & McElhinny 1976). During an excursion, the dynamo does not establish itself in the opposite polarity. Defined in this way, excursions are distinguished from secular variation (when the VGP colatitude is $10^\circ < \theta < 40^\circ$) and from short polarity episodes, which is a term applied when the opposite polarity (θ is $<40^\circ$ or $>140^\circ$) persists sufficiently long for at least one oscillation in the strength of the main dipole (about 104 years, Jacobs 1984).

Several short and almost complete changes in geomagnetic inclination have occurred within the present-day Brunhes Chron. These rapid and global geomagnetic events are called 'excursions' or 'aborted reversals' (Lund *et al.* 2006; Laj & Channell 2007). Among these excursions, the Laschamp (40–41 ka), Blake (*c.* 115–120 ka) and the Pringle Falls (*c.* 211–218 ka) are the most important geomagnetic events because they have been widely documented (e.g. Valet & Meynadier 1998; Guyodo & Valet 1999; Valet *et al.* 2008). Because excursion events have been recognized further back in time (e.g. Handschumacher *et al.* 1988; Sager *et al.* 1998; Tivey *et al.* 2006), we can infer that they occurred probably also during older chrons. These excursions, sometimes called tiny wiggles (e.g. Lanci & Lowrie 1997), cannot yet be used as time constrain.

Relative palaeointensity

The Earth's magnetic field can be simplified as a big dipole with the poles at the geographical poles, which is called the geocentric axial dipole (GAD). The behaviour of the geomagnetic field is complex and is related not only to the GAD but also to other components that change at different timescales: years (secular variation), millennia (excursions) and millions of years (reversals) (Gee & Kent 2007).

Several short and almost complete changes in geomagnetic inclination have occurred within the present-day Brunhes Chron. These rapid and global geomagnetic events are called ‘excursions’ or ‘aborted reversals’ (Lund *et al*. 2006; Laj & Channell 2007). An important long-term (10^5 year) variation of spherical parameters (inclination, declination and intensity) of the geomagnetic field is the secular variation (Valet *et al.* 2008), which includes geomagnetic jerks, westward drift and palaeointensity. Geomagnetic jerks are abrupt changes in one of the geomagnetic components related to inner core flow patterns or major earthquakes (Mandea *et al.* 2000; Florindo *et al.* 2005). The secular variation, which is not constant and uniform on the Earth, is mainly related today to a westward drift of about 0.2° per year since 1400 AD, and an eastward drift from 1000 to 1400 AD (Dumberry & Finlay 2007). The variation in intensity of the geomagnetic field through geological time is known as palaeointensity. Measurements of palaeointensity in different geological sequences (e.g. lava flows, or marine or lake sediments) show a consistent pattern, allowing the reconstruction of a relative palaeointensity curve for the past 2 myr (Fig. 2; Sint-2000; Valet *et al.* 2005). Sint-2000 is a stack curve of independent palaeointensity records from various latitudes for the last 2 myr. It is possible to date a geological sequence by correlating the reference Sint-2000 intensity record to the palaeointensity record of the studied sequence. This is possible only when the magnetic minerals along the section are relatively uniform (King *et al.* 1983; Valet & Meynadier 1998), and when the magnetic intensity has been previously normalized for the concentration and grain-size of magnetic material (thus ‘relative’), which may be climatically driven

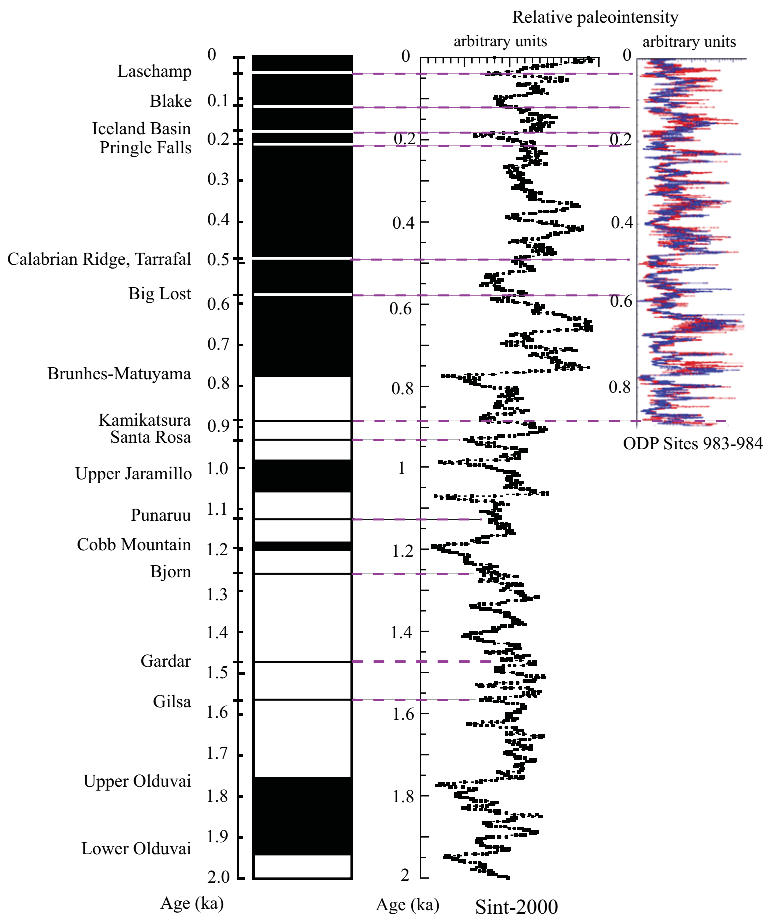

Fig. 2. Polarity column showing the position of the successive excursions and reversals with respect to the fluctuations of dipole field intensity derived from the composite Sint-2000 record and from ODP sites 983 (red line) and 984 (blue line) in the northeastern Atlantic Ocean (arbitrary units increasing toward right) (Guyodo & Valet 1999; Valet *et al.* 2008).

(e.g. Valet & Herrero-Barvera 2000; Channell *et al.* 2002; Yamazaki 2008). It is worth noting that severe declines in relative intensity coincide with geomagnetic reversals and excursions (Fig. 2; Valet *et al.* 2008).

Another application is the scatter of inclination and declination on a sphere, or angular dispersion (S) (e.g. Jovane *et al.* 2008). S is related to latitude, with higher values near the poles, and may indicate the geomagnetic stability of core vortices (Olson & Aurnou 1999; Aurnou *et al.* 2003).

Tectonics and deformation

While palaeomagnetic data have been used for a long time to examine plate-scale to regional/local tectonic and structural problems (e.g. Irving 1963, Beck 1981; Van der Voo 1988), a number of recent studies have combined magnetostratigraphic and palaeomagnetic analyses to examine timing and temporal variations in structural processes. This approach is made possible by the collection of more samples per site (or single geological bed) and more sites than is typical for standard magnetostratigraphic studies (e.g. Zhao *et al.* 2001; Liu *et al.* 2003; Titus *et al.* 2011). This technique has the potential to test velocity models of plate-margin and fault-zone deformation that are based primarily on Global Positioning System (GPS) data (e.g. McCaffrey 2005). GPS deformation studies utilize short (decades at most) time-series of GPS displacement data and their predictions of vertical-axis rotation can be compared with observations obtained from palaeomagnetic data (e.g. Titus *et al.* 2011).

Magnetic inclination and declination track the position of palaeopoles through geological time (Irving 1956). A reconstructed sequence of palaeopoles for a plate is known as an Apparent Polar Wander Path (APWP). We define the APWP through study of ChRM, which points to the VGP of the original position of a rock at the time it was formed. APWPs of the major tectonic plates have been reconstructed through much of Phanerozoic time (e.g. Besse & Courtillot 2002; Kearey *et al.* 2009), but there are still large uncertainties in the APWPs of minor plates and in reference poles (McElhinny & McFadden 2000).

Magnetic anisotropy studies can help define shallowing effects related to diagenetic compaction, flow direction during deposition or cooling and deformation of rocks owing to tectonic stresses. The larger datasets generated from these studies can be utilized to test and resolve other important aspects of how sedimentary rocks record the geomagnetic field. The inclination, declination and intensity of the DRM of sedimentary magnetic grains, and the TRM of lava flows, can be affected by geomagnetic components at the time of deposition or solidification (Tauxe 2010). Consequently, magnetic susceptibility may not be isotropic, giving different values when measured in different directions. In such cases, the anisotropy of magnetic susceptibility (AMS) is calculated as a tensor by comparing the magnetic susceptibility values in three perpendicular directions, which produce a matrix representing the ellipsoid of the magnetic susceptibility (e.g. Hrouda 1982). This ellipsoid of the magnetic fabric can be spherical, oblate (flattened) or prolate (cigar-shaped) providing information, for example, on the direction of a palaeocurrent (e.g. Ellwood & Whitney 1980) or lava flow (e.g. Stacey 1960; Ellwood 1978), strain patterns (e.g. Goldstein 1980), fabric/structure of granites (e.g. Ellwood & Whitney 1980) and the degree of compaction of strata studying the relation between inclination shallowing and oblation of the fabric (e.g. Tan & Kodama 2002).

Environmental magnetism

Environmental magnetism is used to identify and characterize magnetic mineralogy, and provides constraints for interpreting palaeomagnetic results and in understanding the factors that control environmental change. Environmental magnetic measurements are inexpensive, rapid and non-destructive, and provide fundamental information on the size, abundance and composition of magnetic minerals (Verosub & Roberts 1995; Kodama 2012; Liu *et al.* 2013). These measurements can be performed along a sedimentary sequence at high-resolution and include (1) magnetic susceptibility, and artificial remanences; (2) anhysteretic remanent magnetization (ARM); (3) isothermal remanent magnetization (IRM) and back-field isothermal remanent magnetization (BIRM); and (4) coercive-dependent parameters (S-ratio and 'hard' IRM referred to as HIRM).

Low-field magnetic susceptibility ($\chi =$ normalized for weight, $\kappa =$ normalized for volume) represents how much magnetization (m) a material retains when a magnetic field (H) is applied (Table 1). Thus it is a complex parameter reflecting the sum of magnetic materials, such as ferromagnetic *sensu lato* (e.g. magnetite), antiferrimagnetic (e.g. hematite) and non-magnetic materials, such as paramagnetic (e.g. silicates or clays) and diamagnetic (e.g. quartz or carbonate). Results from low-field magnetic susceptibility must be interpreted with caution since it can be related to numerous processes.

Anhysteretic remanent magnetization activates only the finest magnetic minerals, frequently single domain (SD) grains, which do not have a domain wall and are uniformly magnetized. It is,

Table 1. *Units and conversions in the Système Internationale d'unités (SI) and the old CGS*

	CGS	SI
Energy	10^7 erg	1 joule (J)
Force (F)	10^5 dyne	1 Newton (N)
Current (I)	0.1 adamp	1 ampere (A)
Magentic Inducion (B)	10^4 gauss (G)	1 Tesla (T)
Magnetic Field (H)	$4\pi \times 10^{-3}$ oersted (Oe)	1 A m^{-1}
Magnetic Moment (M)	10^3 emu	1 A m^2
Magnetization/volume (m)	10^{-3} emu (=G cm^{-3})	1 A m^{-1}
Magnetization/mass (J)	1 emu g^{-1} or G cm^{-3} g^{-1}	1 A m^2 kg^{-1}
Magnetic susceptibility/volume (κ)	Dimensionless	Dimensionless
Magnetic susceptibility/mass (χ)	10^{-3} cm^3 g^{-1}	1 m^3 kg^{-1}

From Butler (1992) and Collinston (1983).

consequently, a proxy for the concentration of fine magnetite of eolian, biogenic or impactoclastic origin, or from other environmental processes. Isothermal remanent magnetization activates all magnetic grains up to saturation (also called SIRM or M_{rs}) that is usually at 1 T or 900 mT (Table 1). IRM, therefore, is the main proxy for magnetic concentration. The ratio between ARM and IRM (ARM/IRM_{900}) provides a proxy for magnetic grain size. Then, smaller IRMs can be applied backwards to IRM, which are called back-field IRM (BIRM), to produce IRM ratios that supply information on magnetic composition. These are the so-called *S*-ratios (King & Channell 1991) and HIRMs (imparted at 300 and 100 mT, respectively). They are calculated as, for example, $S\text{-ratio}_{300} = BIRM_{300}/IRM_{900}$, $S\text{-ratio}_{100} = BIRM_{100}/IRM_{900}$, $HIRM_{300} = (IRM_{900} + BIRM_{300})/2$ or $HIRM_{100} = (IRM_{900} + BIRM_{100})/2$ to indicate the different responses of high-coercivity magnetic minerals. The presence of hematite can be determined with the proxy IRM_{900}@AF120 mT for the concentration of hematite by AF demagnetizing the IRM_{900} step at 120 mT (Larrasoaña *et al.* 2003). Liu *et al.* (2007) also proposed the *L*-ratio $(IRM_{900} + BIRM_{300})/(IRM_{900} + BIRM_{100})$ to define how the hardness of hematite can affect HIRM and the *S*-ratio.

Environmental magnetic records obtained by finely sampled measurements along sedimentary sequences can be evaluated as time series and related to palaeoclimatic and palaeoenvironmental change. In particular, palaeoclimate change at 10^5–10^6 year scales – often resolved in the sedimentary record – has been strongly modulated by astronomically forced insolation (e.g. Berger 1988; Laskar *et al.* 2004, 2011; Table 2). The record of astronomical forcing frequencies in palaeomagnetic proxies can be used as a tool to perform astronomical tuning and develop continuous timescales along stratigraphic sequences (Hinnov & Hilgen 2012).

Organization of this special volume

Following three conference sessions entitled 'Magnetostratigraphy: not only a dating tool' and presented at the AGU Meeting of Americas (Foz de Iguaçu, Brazil) in 2010 and AGU Fall Meetings 2010, 2011 (San Francisco, USA), we decided to assemble the presentations into a Special Publication entitled *Magnetic Methods and the Timing of Geological Processes*. In this volume, we present research that includes innovations in classical magnetostratigraphy and emerging techniques that use sequential rock magnetic measurements to define chronology in stratigraphy. We introduce applications that use magnetic direction and geomagnetic polarity reversals to infer geodynamo processes, tectonics, diagenesis and climate change.

Table 2. *Periods and relative amplitudes of the main astronomical forcing cycles for 0–10 Ma (Laskar* et al. *2004; Hinnov & Hilgen 2012). The frequencies and relative amplitudes change over different time intervals, and relative amplitudes change additionally as the result of climatic filtering*

Astronomical parameter	Period in kiloyears (relative amplitude)				
Orbital eccentricity	405 (1.0)	131 (0.43)	124 (0.56)	99 (0.48)	95 (0.74)
Obliquity (tilt)	53.04 (0.34)	40.80 (1.0)	40.11 (0.33)	39.45 (0.42)	29.75 (0.16)
Precession index	23.61 (1.0)	22.33 (0.87)	19.07 (0.45)	19.12 (0.69)	16.44 (0.08)

Finally, we present examples of astronomical forcing of environmental magnetism in the sedimentary record, and their utility in defining high-resolution timescales.

Part 1: integrated magnetostratigraphy

When integrated with biostratigraphy, cyclostratigraphy, chemostratigraphy and geochronology, magnetostratigraphy provides valuable timescale information. In this section, modern integrated magnetostratigraphy is demonstrated with marine Cenozoic studies. New Oligocene–Miocene magnetostratigraphy from the equatorial Pacific pinpoints the Oligocene–Miocene transition at the base of C6Cn.2n, and identifies three new excursions (**Guidry *et al.* 2012**). New work is presented on the Eocene climatic optimum interval (50–55 Ma) and Late Eocene–Oligocene transition into the Cenozoic icehouse (*c.* 33.5 Ma) from Integrated Ocean Drilling Progam sites and Italian sections (**Firth *et al.* 2012; Savian *et al.* 2013; Jovane *et al.* 2013**), and a composite integrated magnetostratigraphic sequence from Italy is provided for the entire Palaeogene Period (**Coccioni *et al.* 2012**).

Part 2: dating tectonic processes with magnetic methods

The evolution of palaeomagnetic declination in Chinese continental stratigraphy reveals block rotation, uplift and basin infilling related to the Cenozoic India–Asia collision, as discussed by three papers in this section (**Yan *et al.* 2012*a*, *b*; Fang *et al.* 2012**). The timing of these declination changes is constrained by magnetic reversal stratigraphy. A fourth paper (**Zhao *et al.* 2013**) analyses tectonic-driven sedimentation change in the Nankai Trough, Japan, with the assistance of integrated magnetostratigraphy.

Part 3: relative palaeointensity for dating geological sequences

In this section, the palaeomagnetic secular variation recovered from a Holocene Lake in Indonesia is compared with a recent geomagnetic model for 0–3 ka, and used to evaluate radiocarbon-based chronologies (**Haberzettl *et al.* 2012**). A new evaluation of the Pringle Falls lacustrine sequence confirms the clockwise loop of the VGP around the globe associated with the Pringle Falls Excursion (**Herrero-Bervera & Cañón-Tapia 2012**). The Laschamp, Skalamaelifell and Blake excursions, detected in sediments from offshore Queensland, Australia, are applied as chronostratigraphic constraints (**Herrero-Bervara & Jovane, this volume, in press**). Finally, VGP trajectories measured on the Lower Cretaceous Serra Geral volcanic sequence are interpreted to indicate anisotropy in the Earth's interior (**Caminha-Maciel & Ernesto 2012**).

Part 4: palaeoclimatic change from rock magnetic proxies

The studies in this section resolve astronomical timescales of palaeoclimatic change using rock magnetism stratigraphy evaluated from a wide variety of marine environments: Plio-Pleistocene marginal marine, Cretaceous carbonate platform, Permian lower slope and basinal carbonates, Permo-Carboniferous glaciogenic rhythmites, and Ordovician shallow shelf limestone/mudstone. The Plio-Pleistocene study astronomically tunes a magnetic susceptibility series from a sedimentary section in northern Italy that otherwise exhibits no cyclicity (**Gunderson *et al.* 2012**). The Cretaceous study (**Hinnov *et al.*, this volume, in press**) finds that ARM cyclicity is independent from host carbonate platform cycles in northeastern Mexico; the former is linked to a precession-forced eolian dust flux and the latter to much lower frequency sea-level fluctuations. **Ellwood *et al.* (2012*a*)** develop obliquity-tuned 'floating timescales' for magnetic susceptibility variations along Middle Permian sections, Texas, to assess sedimentation rates and durations of geological events. **Franco & Hinnov (2012)** evaluate anisotropy of magnetic susceptibility in Brazilian Permocarboniferous glacial rhythmites as a palaeoclimate indicator. Finally, the cyclic Ordovician Kope Formation, USA, is evaluated for astronomical frequencies through analysis of a composite high-resolution magnetic susceptibility series (**Ellwood *et al.* 2012*b***).

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