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Magnetostratigraphy – concepts, definitions, and applications

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With 12 figures

Abstract. The most characteristic feature of the Earth's magnetic field is that it reverses polarity at irregular intervals, producing a 'bar code' of alternating normal (north directed) and reverse (south directed) polarity chrons with characteristic durations. Magnetostratigraphy refers to the application of the well-known principles of stratigraphy to the pattern of polarity reversals registered in a rock succession by means of natural magnetic acquisition processes. This requires that the rock faithfully recorded the ancient magnetic field at the time of its formation, a prerequisite that must be verified in the laboratory by means of palaeomagnetic and rock magnetic techniques.

A sequence of intervals of alternatively normal or reverse polarity characterized by irregular (non-periodic) duration constitutes a distinctive pattern functional for correlations. Over the last 35 Myr, polarity intervals show a mean duration of ~ 300,000 years, but large variations occur from 20,000 yr to several Myr and even up to tens of Myr. By correlating the polarity reversal pattern retrieved in a rock succession to a reference geomagnetic polarity time scale (GPTS), calibrated by radioisotopic methods and/or orbital tuning, the age of the rock succession can be derived. Magnetostratigraphy and correlation to the GPTS constitute a standard dating tool in Earth sciences, applicable to a wide variety of sedimentary (but also volcanic) rock types formed under different environmental conditions (continental, lacustrine, marine). It is therefore the stratigraphic tool of choice to perform correlations between continental and marine realms. Finally, we emphasise that magnetostratigraphy, as any other stratigraphic tool, works at best when integrated with other dating tools, as illustrated by the case studies discussed in this paper.

Part one – Concepts, definitions, and applications

Introduction

Dating and time control are essential in all disciplines of the Earth Sciences, since they allow to correlate rock sequences from distant localities and different (marine and continental) realms. Moreover, accurate time control is the sine qua non to understand rates of change of natural processes and thus to determine the underlying mechanisms that explain our observations. Biostratigraphy of different faunal and floral systems has been used since the 1,840 s to erect the relative geological age of sedimentary rocks and hence to perform correlations among them. Radioisotopic dating, originally applied mostly to igneous rocks, has become increasingly sophisticated and can now count on a wide variety of isotopic decay systems capable to provide numerical ages also in sedimentary rocks formed under

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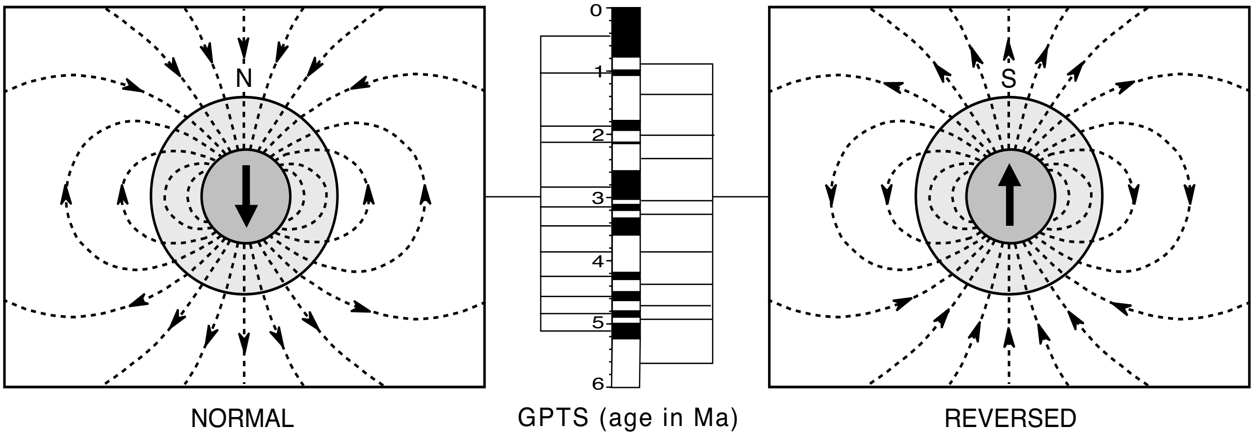


Fig. 1. Schematic representation of the geomagnetic field of a geocentric axial dipole ('bar magnet'). During normal polarity of the field the average magnetic north pole is at the geographic north pole, and a compass aligns along magnetic field lines. Historically, the north pole is referred to as the pole attracting the 'north seeking' needle of a compass, but physically it is a south pole. During normal polarity, the inclination is positive (downward directed) in the northern hemisphere and negative (upward directed) in the southern hemisphere. Conversely, during reversed polarity, the compass needle points south, and the inclination is negative in the northern and positive in the southern hemisphere. In the geomagnetic polarity time scale (column in the middle), periods of normal polarity are conventionally represented by black intervals, reversed intervals by white intervals.

favorable environmental conditions. In this paper, we describe the principles of magnetostratigraphy defined as a dating tool that uses the record of polarity (normal or reverse) of the ancient geomagnetic field registered in igneous or sedimentary rock sequences by natural magnetic acquisition processes.

The most distinctive property of the Earth's magnetic field is that it can reverse polarity (Fig. 1). During the siege of Lucera (Apulia, Italy) by the armies of Charles d'Anjou, an army engineer called Pierre Pèlerin de Maricourt, better known as Petrus Peregrinus ('Peter the Pilgrim'), made a remarkable series of observations. He described the dipolar nature of a (magnetised) spherical lodestone, and showed that the magnetic force of this natural dipole is strongest and vertical at the poles. In 1269, he reported his findings in the Epistola de Magnete – regarded as the first scientific treatise ever written – and became the first to formulate the law by which poles with same magnetic charge repel whereas poles with opposite charge attract. In 1600, William Gilbert published the results of his experimental studies on magnetism entitled De Magnete, Magneticisque Corporibus, et de Magno Magnete Tellure. He investigated the variation in inclination over the surface of a spherical lodestone and concluded for the first time that 'magnus magnes ipse est globus terrestris' (the Earth's globe itself is a great magnet). Apart from the spherical form of the Earth, magnetism was the first physical property attributed to the body of the Earth as a whole. Newton's theory of gravitation came 87 years later with the publication of his 'Philosophiæ Naturalis Principia Mathematica. A pleasantly readable article on a more elaborate history of magnetism is by Stern (2002).

Palaeomagnetic studies of igneous rocks provided the first reliable information on polarity reversals. In 1906, Bernard Brunhes, at that time director of the Puy-de-Dôme Observatory (France), observed lava flows magnetised in a direction approximately antiparallel to the present geomagnetic field, and suggested that this was caused by a reversal of the field itself, rather than by a self-reversal mechanism of the rock. In 1929, Motonori Matuyama demonstrated that young Quaternary lavas were magnetised in the same direction as the present field (normal polarity), whereas older lavas were magnetised in the opposite direction. But it was only in the early 1,950 s that Jan Hospers with his study of Icelandic basalts succeeded to convince many in the geophysics community that reversed magnetism in rocks was not caused by a self-reversal processes but was a record of times when the Earth's magnetic field had reversed polarity state (Hospers 1951). He was the first to realise that the polarity of lava flows could be a powerful stratigraphic correlation tool (cf. Irving 1988; Stern 2002), while Khramov (1958) realised that by correlating volcanic and sedimentary rock successions worldwide, it could be possible to develop a single geochronological palaeomagnetic time scale valid for the whole Earth. The first geomagnetic polarity time scales (GPTS) began to take shape when the combined use of radioisotopic dating and magnetostratigraphy was adopted on lava flows (Fig. 2; Cox et al. 1963, 1964). Improved techniques and the undertaking of extensive palaeomagnetic investigations in many parts of the world have significantly increased the amount of palaeomagnetic information and have now provided a very detailed Geomagnetic Polarity Time Scale (e. g. Opdyke and Channell 1996), of which the Neogene part is fully astronomically calibrated (Lourens et al., 2004) through the extensive use of cyclostratigraphy (e. g. Strasser et al. 2006).

More than one century after the first reversal of the Earth's magnetic field was discovered by Brunhes (1906), more than half a century after the start of the modern era of magnetostratigraphy (Hospers 1951), and almost half a century after the first development of the modern GPTS (Cox 1963), it can be concluded that magnetostratigraphy has evolved into an indispensable stratigraphic tool for Earth Sciences.

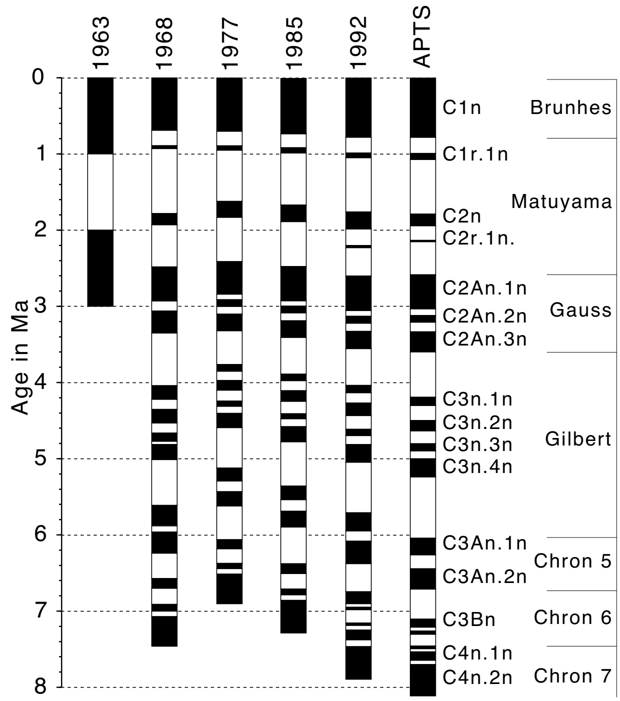


Fig. 2. Development of the geomagnetic polarity time scale (GPTS) over the last half century of study. The initial assumption of periodic behaviour (Cox, 1963) was soon abandoned as new data became available. The first modern GPTS based on marine magnetic anomaly patterns was established by Heirtzler et al. (1968). Subsequent revisions by Labreque et al. (1977), Berggren et al. (1985), Cande, Kent (1992) show improved age control and increased resolution. A major breakthrough came with the astronomical polarity time scale (APTS) in which every individual reversal is accurately dated (e. g. Hilgen et al., 1997). Black (white) intervals denote normal (reversed) polarity; for chron nomenclature, see text.

The palaeomagnetic signal

The Earth's magnetic field is generated in the liquid outer core through a dynamo process that is maintained by convective fluid motion. At the surface of the Earth, the field can be conveniently described as a dipole field, which is equivalent to having a bar magnet at the centre of the Earth. Such a dipole accounts for approximately 90% of the observed field. The remaining 10% derives from higher order terms: the non-dipole field. At any one time, the best fitting geocentric dipole axis does not coincide exactly with the rotation axis of the Earth, but averaged over several thousand years, we may assume the dipole to be both geocentric and axial (Merrill et al. 1996). Initially, it was believed that the field reversed periodically, but as more results on lava flows became available, it became clear that geomagnetic reversals occur randomly. It is precisely this random character that confers stratigraphic value to a measured polarity reversal sequence. A polarity reversal typically takes several thousands of years to occur, fast enough to be considered globally synchronous on geological time scales. The field itself is sign invariant whereby the same configuration of the geodynamo can produce either normal or reverse polarity. What causes the field to reverse is still debated, but recent hypotheses suggest that lateral changes in heat flow at the core-mantle boundary play an important role (e. g. Gubbins 1999). Although polarity reversals occur at irregular times, the reversal frequency can change considerably over geological time spans. For instance, the polarity reversal frequency has increased since 80 Ma from approximately 1 rev/Myr to 5 rev/My in more recent times. In the Cretaceous, from 124.5 to 84 Ma, no reversals occurred and the field maintained stable normal polarity for some 40 Myr. This Cretaceous Superchron is known in ocean-floor magnetic anomaly profiles as the Cretaceous Normal Quiet Zone.

The ancient geomagnetic field can be registered in rocks at the time of their formation. Rocks commonly contain magnetic minerals, usually iron (hydr)oxydes or iron sulphides. During rock-forming processes, these magnetic minerals (or more accurately, their magnetic domains) statistically align with the then ambient field, and are subsequently 'locked in' the rock system, thus preserving the direction of the field as a natural remanent magnetisation (NRM): the palaeomagnetic signal.

We distinguish three basic types of NRM, depending on the mechanism of palaeomagnetic signal acquisition: TRM, CRM and DRM. A thermoremanent magnetisation (TRM) is the magnetisation acquired when a rock cools below the Curie temperature of its magnetic minerals, thereby 'locking' the magnetic domains along positions statistically aligned with the ambient field and producing a magnetic remanence that at room temperature may be stable for billions of years. A chemical remanent magnetisation (CRM) is the magnetisation acquired when a magnetic mineral grows through a critical 'blocking volume' or grain size at which the field is locked in and the acquired remanence may again be stable over billions of years. A detrital remanent magnetisation (DRM) is the magnetisation acquired when magnetic grains of detrital origin are deposited. Magnetic grains responsible for a DRM can also form directly in the water column as magnetosomes: intra-cellular chains of magnetic minerals made by magnetotactic bacteria (e. g. Vasiliev et al. 2008). Detrital magnetic grains statistically align with the ambient field as long as they are in the water column or in the soft water-saturated topmost layer of the sediment. Upon compaction and dewatering, the grains are mechanically fixated in a 'lock-in depth zone' and will preserve the direction of the ambient field. Within the sediment, authigenic or diagenetic formation of magnetic minerals may take place which also record the field. This may occur in an early stage, but also well after deposition, deeper within the sediment. In the latter case, this may cause an apparent delay of the NRM acquisition which can distort the magnetic record.

Demagnetisation, laboratory tests, and field tests

Frequently, the total NRM is the vector sum of different magnetic components (Fig. 3). This is because the primary NRM, i. e., the magnetisation originated at the time of rock formation, may be overprinted by magnetic components acquired later in geologic history through weathering reactions at room temperature or thermochemical reactions associated with tectonic or burial processes. An overprint component can be removed through 'magnetic cleaning' techniques. These are primarily the thermal demagnetisation technique and the alternating magnetic field (AF) demagnetisation technique (e. g. Zijderveld 1967; Langereis et al. 1989; Butler 1992; Tauxe 1998). During demagnetisation experiments, samples are subjected to stepwise increasing values of temperature or alternating field in a zero magnetic field (field-free) space. The residual magnetisation is measured after each demagnetisation step and the resultant changes in direction and intensity are displayed and analysed in order to reconstruct the complete component structure of the NRM. Some palaeomagnetic laboratories have recently invested in automatic measurement systems, which enable continuous measurements of large amounts of samples. Automated and repeating measurement schemes yield previously unattainable amounts of information, and may give results even from low-intensity limestones (e. g. Gong et al. 2008) provided such systems are designed to maintain low noise levels during measurements.

The results of stepwise demagnetisation are commonly visualized and analyzed using the so-called Zijderveld diagrams (after Zijderveld 1967), also known as vector end-point demagnetisation diagrams (Fig. 3). In these diagrams, both the intensity and directional changes of the NRM occurring during demagnetisation are displayed at the same time. Magnetic components are then extracted from the Zijderveld diagrams using least-square analysis (Kirschvink 1980), and the most stable and consistent component that can be isolated is referred to as the characteristic remanent magnetisation (ChRM). This ChRM is further investigated to establish if it represents a record of the geomagnetic field at, or close to, the time of rock formation, or a secondary magnetisation acquired later in geologic history by post-depositional processes. To assess the primary nature of the ChRM, and hence its suitability for magnetostratigraphic studies, rock magnetic experiments and reliability tests are usually carried out. Rock magnetic experiments are aimed at determining the fundamental characteristics of the minerals bearing the magnetic remanence (e. g., type, grain size, etc.). A review of these methods is beyond the scope of this paper but can be found in appropriate text books (e. g. Butler 1992) or review papers. The three most important reliability tests for magnetostratigraphy are:

Consistency test. A natural remanent magnetisation component is considered primary in origin when it defines a sequence of polarity reversals that is laterally traceable by independent means (e. g. lithostratigraphy) between distant sections from different parts of the basin.

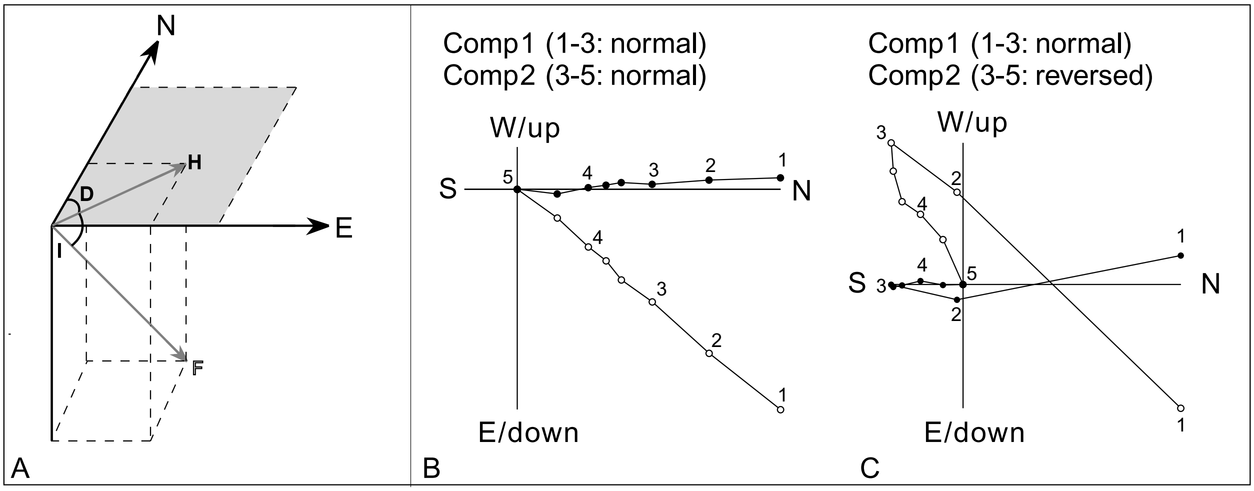


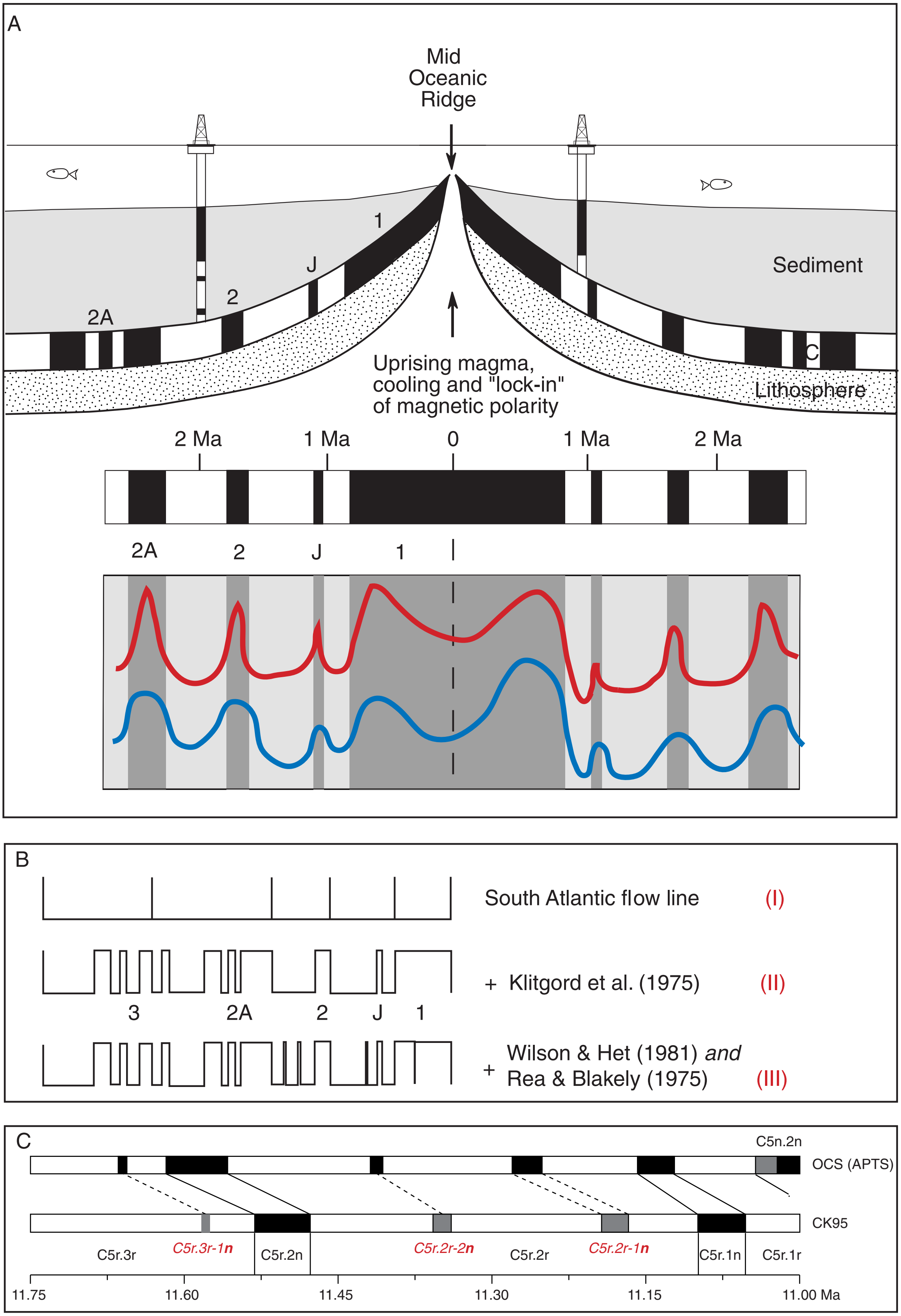
Fig. 3. (A) The magnetic field on any point on the Earth's surface is a vector (F) which possesses a component in the horizontal plane called the horizontal component (H) which makes an angle (D) with the geographical meridian. The declination (D) is an angle from north measured eastward ranging from 0° to 360°. The inclination (I) is the angle made by the magnetic vector with the horizontal. By convention, it is positive if the north-seeking vector points downward and negative if it points upward. (B) and (C) To resolve the different magnetic components that can be acquired in a rock during its geological history, rock samples are subjected to a process of stepwise demagnetisation. The standard method for presentation and analysis of the results is called Zijderveld diagrams (after Zijderveld, 1967). Changes in the magnetisation vector during demagnetisation involve both its direction and its intensity; orthogonal vector Zijderveld diagrams show the changes in both. The endpoint of the vector measured after each demagnetisation step is projected both onto the horizontal plane (closed symbols) and onto the vertical plane (open symbols). Difference vectors (lines between end points) then show the behaviour of the total vector upon stepwise removal of the magnetisation. Conventionally, she solid points are these endpoints when projected onto the horizontal plane containing axes NS and EW, whereas the open points are these endpoints when projected onto the vertical plane containing axes NS (or EW), and up-down. Although many variations exist in literature, the only sensible projected axes combinations are W/up vs. NS and N/up vs. EW.

Reversal test. The observation of characteristic remanence directions with different polarity and, in particular, the occurrence of antiparallel (within statistical error) directions is taken as a strong indication for the primary origin of that ChRM. This test is greatly enhanced if a polarity zonation can be established, and if this zonation is independent of possible changes in the composition of the rock.

Fold test. If the ChRM directions from differently tilted beds converge after correction for the dip of the strata, this remanence was acquired before tilting. Strictly speaking, this fold test does not directly prove a primary origin of this component, but only that it dates from before tilting.

The Geomagnetic Polarity Time Scale

For the construction of the 'bar code' pattern of magnetic polarity intervals, geophysicists rely on two fundamentally different records of geomagnetic polarity history: the marine magnetic anomaly record and the magnetostratigraphic record. Surveys over the ocean basins carried out from the 1950's onward found linear magnetic anomalies, parallel to mid-oceanic ridges, using magnetometers towed behind research vessels (Cox et al. 1963; Heirtzler et al. 1968). During the early 1960s, it was suggested, and soon after confirmed, that these anomalies resulted from the remanent magnetisation of the oceanic crust. This remanence is acquired during the process of sea-floor spreading, when uprising magma beneath the axis of the oceanic ridges cools through the Curie temperatures of its constituent ferromagnetic minerals in presence of the ambient geomagnetic field, thus acquiring its direction and polarity. The continuous process of rising and cooling of magma at the ridge results in magnetisaed crust of alternating normal and reverse polarity, which produces a slight increase or decrease of the measured field: the marine magnetic anomalies (Fig. 4a). It was also found that the magnetic anomaly pattern is generally symmetric on both sides of the ridge, and, most importantly, that it provides a remarkably continuous record of


A
Mid
Oceanic
Ridge
1
J
2
2A
Sediment
C
Lithosphere
Uprising magma,
cooling and "lock-in"
of magnetic polarity
2 Ma
1 Ma
0
1 Ma
2 Ma
2A
2
J
1
B
South Atlantic flow line
(I)
+ Klitgord et al. (1975)
(II)
3
2A
2
J
1
+ Wilson & Het (1981) *and*
Rea & Blakely (1975)
(III)
C
C5n.2n
OCS (APTS)
CK95
C5r.3r
C5r.3r-1n
C5r.2n
C5r.2r-2n
C5r.2r
C5r.2r-1n
C5r.1n
C5r.1r
11.75
11.60
11.45
11.30
11.15
11.00 Ma

the geomagnetic reversal sequence. The template of magnetic anomaly patterns from the ocean floor has remained central for constructing the GPTS from magnetic polarity chron M0r in the Early Cretaceous onward (~ 124.5–0 Ma; He et al. 2008). Combined magnetostratigraphic, biostratigraphic, and radioisotopic results of deep-sea sediments and land-based sections have confirmed and refined the general validity and accuracy of the GPTS (e. g. LaBreque et al. 1977; Berggren et al. 1985). The development of the GPTS reflects increasing detail and gradually improved age control.

The latest development in constructing a GPTS comes from orbital tuning of the sediment record, the so called Astronomically calibrated Polarity Time Scale (APTS) (Hilgen et al. 1997). The APTS is now almost complete for the Neogene (Lourens et al. 2004) and has been developed for the Late Triassic (Kent and Olsen 1999; Olsen and Kent 1999). It differs essentially from the conventional GPTS, in the sense that each reversal boundary – or any other geological boundary, e. g. biostratigraphic datum levels or stage and epoch boundaries – is dated individually. This time scale has the inherent promise of increasingly advancing our understanding of the climate system, because cyclostratigraphy and orbital tuning rely on deciphering environmental changes driven by climate change, which in turn is orbitally forced. The main feature of an APTS is that the age of each reversal is directly determined, rather than interpolated between radioisotopic calibration points. This has important consequences for changes in spreading rates of plate pairs, as sea floor spreading rates can now be more accurately determined. Indeed, Wilson (1993) found that the use of astronomical ages resulted in very small and physically realistic spreading rate variations. As a result, the discrepancy in plate-motion rates from the global plate tectonic model (NUVEL-1; DeMets et al. 1990) with respect to those derived from geodesy has become much smaller, and NUVEL-1 has been updated (to NUVEL-1A; DeMets et al. 1994) to incorporate the new astronomical ages.

Periods of a predominant (normal or reverse) polarity are called chrons, and the four youngest ones are named after leading scientists in the field of palaeomagnetism or geomagnetism, i. e. Brunhes, who suggested the existence of field reversals, Matuyama, who proved this, Gauss, who mathematically described the field, and Gilbert, who discovered that the Earth itself is a big magnet. The Brunhes to Gauss chron sequence contain short intervals of opposite polarity called subchrons, which are named after the locality where they were discovered, e. g. the Olduvai normal polarity subchron within the Matuyama reverse polarity chron is named after the Olduvai Gorge (Tanzania), or the Kaena reverse polarity subchron within the Gauss normal polarity chron named after Kaena Point (Hawaii).

Older chrons were not named but numbered according to the anomaly numbers originally given by Heirtzler et al. (1968). Cande and Kent (1992) developed a consistent (sub)chron nomenclature that is now used as the standard (details in their Appendix: Nomenclature). A more recent revision of the Cande and Kent (1992) time scale was made by Cande and Kent (1995) in which they adopted the orbitally tuned timescale (Shackleton et al. 1990; Hilgen 1991a, b) for the last 5.3 Myr.

Tiny wiggles, cryptochrons and subchrons

The magnetic anomaly template over the last 84 Myr was thoroughly revised by Cande and Kent (1992), up to the Cretaceous Normal Superchron (Cretaceous

Fig. 4. A) Formation of marine magnetic anomalies during seafloor spreading. The oceanic crust is formed at the ridge crest, and while spreading away from the ridge it is covered by an increasing thickness of oceanic sediments. The black (white) blocks of oceanic crust represent the original normal (reversed) polarity of the thermoremanent magnetisation (TRM) acquired upon cooling at the ridge. The black and white blocks in the drill holes represent normal and reversed polarity depositional remanent magnetisation (DRM) acquired during deposition of the marine sediments. Normal polarity anomalies are given numbers and refer to anomaly 1 (Brunhes Chron), 2 (Olduvai subchron) and 2A (Gauss Chron); J = Jaramillo subchron. B) Stacks of marine magnetic anomaly profiles (red line) are used to model (blue line) the polarity sequence. B) Categories used by Cande and Kent (1992): category I = major anomalies along a synthetic reference flow line in the South Atlantic, II = filling in details from the best South Atlantic profiles (Klitgord et al. 1975), III = further details from (often deeptow) anomaly profiles from fast-spreading ridges (Wilson and Hey, 1981, Rea and Blakely, 1975). C) Example of Chron-Subchron-Cryptochron nomenclature: reversed Chron C5r, with normal subchrons C5r.1n and C5r.2n. Cryptochrons C5r.2r-1, C5r.2r-2 and C5r.3r-1 elevated to the status of subchrons C5r.2r-1n, C5r.2r-2n and C5r.3r-1n (after Abdul Aziz and Langereis 2004).

Quiet Zone in magnetic anomaly profiles). They constructed a synthetic flow line in the South Atlantic, with first order distances built up from a combination of finite rotation poles, designated as category I intervals. On these intervals, they projected the best quality profiles surveyed in this ocean basin, providing category II intervals (Fig. 4b). Since spreading rates in the Atlantic are relatively slow, they subsequently filled in the category II intervals with high-resolution profiles from fast spreading ridges (their category III). This enabled them to include much more detail on short polarity intervals (or subchrons), for instance around 7 Ma (Fig. 2). In total, they used 9 calibration points to construct their GPTS, including for the first time as youngest tie point an astronomically calibrated age for the Gauss/Matuyama boundary.

The reliability and completeness of the GPTS is crucial for geochronology but also for understanding the long-term statistical properties of the geomagnetic field. The shortest polarity intervals in the GPTS are typically on the order of 30 kyr in duration, but the magnetic anomaly patterns of fast spreading oceanic plates indicate that smaller-scale variations exist as well. They have been referred to as what they look like: tiny wiggles (LaBrecque et al. 1977); these very short and low intensity anomalies have an uncertain origin. Tiny wiggles may represent very short subchrons of the field, as has been proven for some of them (e. g. the Cobb Mt. subchron at 1.21 Ma, or the Réunion subchron at 2.13–2.15 Ma), or just represent intensity fluctuations of the geomagnetic field, causing the oceanic crust to be less (or more) strongly magnetised. Because of their uncertain or unverified nature, these were called cryptochrons (Cande and Kent, 1992). The cryptochrons have a designation (–1, –2, etc.) following the primary (sub)chron nomenclature. For example, three new cryptochrons were recently discovered in the reversed Chron C5r in Middle-Late Miocene continental deposits of Spain (Abdul Aziz et al. 2004; Abdul Aziz and Langereis 2004) (Fig. 4C). Chron C5r contains two normal polarity subchrons (C5r.1n and C5r.2n) and is consequently divided in three reversed polarity intervals (C5r.1r, C5r.2r and C5r.3r). Since two of the cryptochrons were found in C5r.2r, they are denoted C5r.2r-1 and C5r.2r-2, from young to old. The third cryptochron in C5r.3r then must become C5r.3r-1. Cande and Kent (1992) proposed that a cryptochron can be elevated to the status of subchron if it corresponds to a magnetostratigraphically documented pair of reversals. Since this was the case in the Spanish deposits, the cryptochrons therein found were designated as subchrons and acquired a polarity suffix: C5r.2r-1n – C5r.2r-2n and C5r.3r-1n.

The origin of many cryptochrons has not yet been confirmed by magnetostratigraphic studies. On the other hand, there is firm evidence for excursions and reversal excursions. The term excursion is used for virtual geomagnetic poles (VGP) deviating more than 45° from geographical north (Verosub and Banerjee 1977), while it is termed a reversal excursion for VGPs that deviate in excess of 90° from geographical north (Merrill and McFadden 1994) and possibly reaching (near) opposite polarity. Invariably, reversal excursions are associated with low palaeointensities of the geomagnetic field, both in the Brunhes and in the Matuyama Chrons (Guyodo and Valet, 1999; Channell et al. 2002). Reversal excursions have a typical duration of 3–6 kyr (Langereis et al. 1997), which is usually much too short to be detected in marine magnetic anomaly profiles, and explains why so few excursions have been detected, even as tiny wiggles. Excursions are most frequently observed within the Brunhes normal polarity Chron (0–781 ka), but this is in part because of the chronostratigraphic coverage of DSDP and ODP holes, which decreases more or less exponentially with depth, and in part because of our higher confidence in assuming that excursions of reverse polarity are primary in origin and not due to e. g. overprinting by the normal polarity present-day field. In any case, reversal excursions or even short subchrons are as a rule not suitable for magnetostratigraphic correlation because of their elusive nature inherent to their very short duration (Roberts and Winkelhofer 2004).

Part two – Case studies

We present three case studies to illustrate the use of magnetostratigraphy as a dating and correlating tool. Each case study comes from a different area with a different geological age and depositional setting. The first study deals with the Neogene part of the GPTS, which is based on the direct dating of polarity reversals and biostratigraphic datums using astronomical curves in the Monte dei Corvi section, Italy (Hüsing et al. 2007, 2009a). This case study provides a state-of-the-art example of integrated stratigraphy with orbitally controlled resolution. The second case study is from the middle Triassic Seceda core, Italy (Muttoni et al. 2004a), which provides an example of a multi-disciplinary and integrated stratigraphic approach. The third case study deals with the Carboniferous-Permian Reversed Superchron (PCRS), and discusses the state-of-the-art stratigraphy of its lower and upper boundaries. In addition, it provides an updated stratigraphic chart for the late Permian, and a newly interpreted record of the Illawarra reversal that marks the end of the PCRS.

Cenozoic case study: The Middle Miocene Monte dei Corvi section, Italy

Sedimentary cycles reflect climatic oscillations that are ultimately controlled by the Earth's orbital cycles as described by the Milankovitch theory, and are accordingly known also as astrocycles (Strasser et al. 2006). Perturbations in the Earth's orbit and rotation axis are climatically important because they affect the global, seasonal, and latitudinal distribution of the incoming solar insolation. They are held responsible for the Pleistocene ice ages but also affect low-latitude climatic systems such as monsoons. Orbitally forced climatic oscillations are recorded in sedimentary archives through changes in sediment properties, fossil communities, and chemical characteristics (Strasser et al. 2006). While Earth scientists can read the geological archives to reconstruct palaeoclimate, astronomers have formulated astronomical solutions that include both the solar-planetary system and the Earth-Moon system. With these astronomical solutions, they compute the past variations in precession, obliquity, and eccentricity (Varadi et al. 2003; Laskar et al. 2004). As a logical next step, sedimentary archives can be dated by matching patterns of palaeoclimatic variability with patterns in the computed astronomical curves. This astronomical tuning of the sedimentary record results in timescales that are largely independent of radioisotopic dating (Lourens et al. 2004; Kuiper et al. 2008).

Initially, research focused mostly on the Pliocene-Pleistocene – using palaeoclimatic records from Ocean Drilling Project sites in the eastern equatorial Pacific and North Atlantic (Shackleton et al. 1990) and sedimentary cycle patterns in Pliocene-Miocene marine successions exposed on land in the Mediterranean area (Hilgen 1991a, b; Hilgen et al. 1995; Krijgsman et al. 1995, 1999; Hüsing et al. 2007, 2009a). In the Mediterranean area, well-known sections such as Capo Rossello, Eraclea Minoa, Singa, and Vrica were used to construct detailed magnetostratigraphic records (Langereis and Hilgen 1991). These sections consist of carbonates and sapropels (brownish-colored layers enriched in organic carbon) arranged in a remarkably clear cyclic pattern that is controlled by precession and eccentricity. The next goal was to extend the Mediterranean-based APTS into the Miocene. Suitable upper Miocene sections were identified on land on the islands of Crete, Gavdos, and Sicily, and provided a straightforward calibration of their cyclic sedimentary pattern and magnetostratigraphy to the astronomical curves (Hilgen et al. 1995). The resulting time scale embraces in continuity the interval between 9.7 and 6.8 Ma, whereas a "Messinian Gap" is present from 6.8 to 5.3 Ma. This gap is explained by the less favourable sediments (e. g., diatomites, evaporites) deposited during the so-called Messinian Salinity Crisis and the notoriously complex depositional history of the Mediterranean during this time interval (Krijgsman et al. 1999). The classic Messinian sediments, however, also displayed distinct sedimentary cyclicities, holding great promise for astronomical dating. Cyclostratigraphic and detailed palaeoclimatic studies revealed that the sedimentary cycles of the Messinian pre-evaporites and evaporites are dominantly controlled by precession-induced changes in circum-Mediterranean climate (Krijgsman et al. 1999, 2001). Closing the Messinian gap resulted in an APTS for the last ~ 10 Myr, which comprised a precisely dated record of the palaeoceanographic and palaeoclimatologic changes occurring in the Mediterranean region at that time. Magneto-cyclostratigraphic dating consequently allowed marine-continental (Abdul Aziz et al. 2004) and Mediterranean-Paratethys (Vasiliev et al. 2004) correlations of unprecedented high resolution, in which the many fossil and palaeoclimate data from continental, lacustrine, and brackish-water settings could be correlated to the global marine proxy records.

Another application of the astronomical polarity timescale has been the dating of Neogene stage boundaries via their Global boundary Stratotype Section and Point (GSSP), many of which have recently been defined in the Mediterranean, like the base of the Zanclean Stage and base of the Pliocene Series (Van Couvering et al. 2000). The availability of a good astrochronology has effectively become a prerequisite for the definition of a GSSP. Another example is shown by the Tortonian GSSP, which has recently been placed at the mid-point of the sapropel in basic cycle 76 in the Monte dei Corvi Beach section near Ancona, Italy (Hilgen et al. 2005). Here, a detailed and integrated stratigraphy of calcareous plankton biostratigraphy, magnetostratigraphy, and cyclostratigraphy has been established (Hilgen et al. 2003).

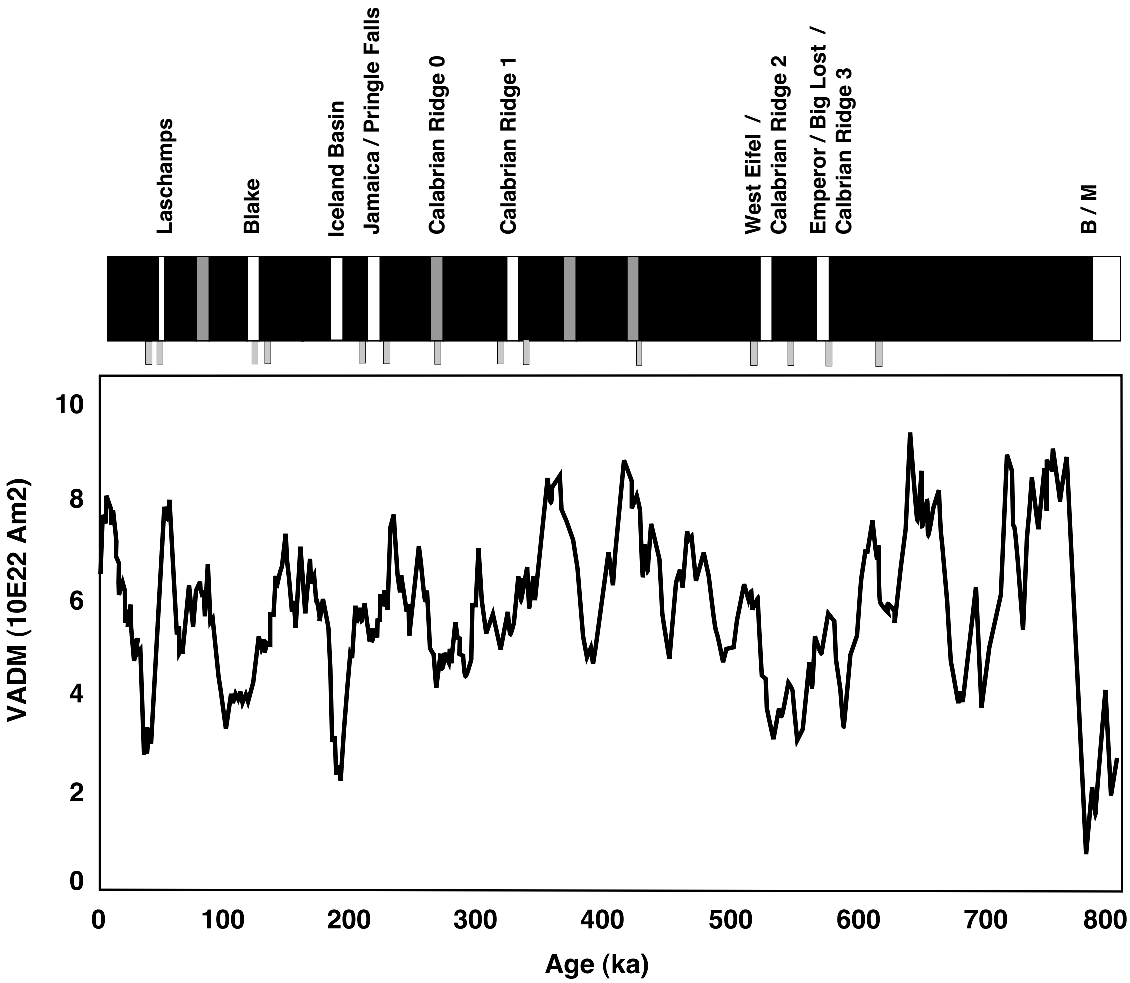


Fig. 5. Virtual Axial Dipole Moment (VADM) of the field for the past 800 kyr (Guyodo and Valet 1999). Low intensities are typically correlated with the occurrence of reversal excursions, short periods where the virtual geomagnetic pole deviates more than 90° from the north geographic pole (white intervals are well confirmed reversal excursions, grey intervals require confirmation; Langereis et al., 1997). Small grey bars outside the column are excursions from ODP cores found by Lund et al. (1998). B/M is Brunhes-Matuyama boundary, showing very low intensities – down to 10% of the stable polarity field – during the reversal.

This section has now been extended upwards to include the Tortonian-Messinian boundary and therefore the entire Tortonian Stage (Hüsing et al. 2009a). Palaeomagnetic results revealed a characteristic low-temperature component characterized by dual polarity, mostly carried by a Fe-sulphide (greigite), but in the younger interval by greigite and magnetite (Hüsing et al. 2009b). The different magnetic minerals carrying the NRM reflect the palaeoenvironment of deposition, whereby oxic, suboxic to anoxic, and anoxic conditions determined the occurrence of, respectively, magnetite, magnetite and greigite, and greigite as recorders of the Earth's magnetic field. The resultant magnetostratigraphy, calibrated to the Astronomically Tuned Neogene Time Scale, shows that the section ranges at least from Chron C5An.2n up to C3Br.2n. The Monte dei Corvi section is the only continuous Tortonian section in the Mediterranean area and is therefore suggested as the Tortonian Reference Section (Fig. 6) (Hüsing et al. 2009a). The correlation of the Tortonian GSSP to the middle part of Chron C5r.2n guarantees global correlation potentials, and the astronomical tuning performed on a precessional scale using the La2004(1,1) solution (Laskar et al. 2004) yielded astronomical ages of each basic cycle and, hence, of the calcareous plankton events and magnetic reversal boundaries with uncertainties on the order of a few thousand years. As a result, the age of the Tortonian GSSP is now 11.625 Ma,

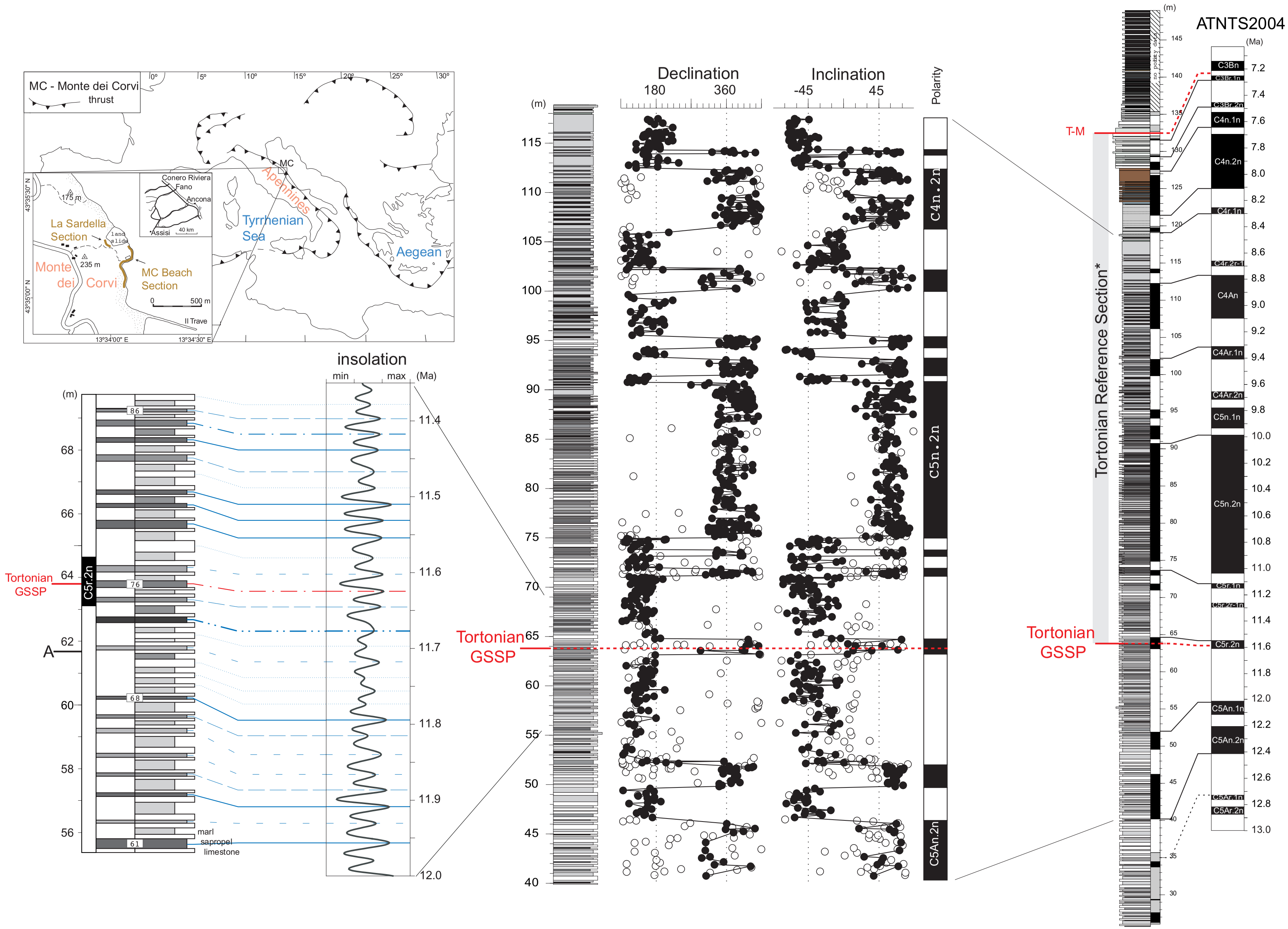


Fig. 6. Magnetostratigraphy for the composite Monte dei Corvi Beach section and calibration to the Neogene astronomical time scale (ATNTS2004, Lourens et al. 2004). Black (white) zones indicate normal (reversed) polarity intervals; grey shading denotes uncertain polarity; filled (open) cicles denote reliable (unreliable) ChRM directions. The Tortonian GSSP, located in the mid-point of the sapropel in cycle 76, is indicated by the red dashed line. The Monte dei Corvi section is suggested as the Tortonian reference section. Sedimentary cycles are calibrated to the astronomical solution La2004(1,1), based on sapropels corresponding to insolation maxima (precession minima). A = Ancona ash (after Hüsing et al. 2007, 2009a).

superseding the previously published estimate of 11.608 Ma (Hilgen et al. 2005). This clearly illustrates that continuous refinements of the geological timescale is a forever ongoing research effort.

Mesozoic case study: The Middle Triassic Seceda section, an integrated stratigraphic study from the Dolomites, Italy

Middle Triassic magnetostratigraphy and biostratigraphy in both Tethyan marine and continental sequences received considerable attention during the last decade. In the compilation of Muttoni et al. (2000), a total of ~ 42 superposed and biostratigraphically calibrated polarity zones were recognised in 15 partially overlapping Tethyan marine sections spanning a late Early Triassic to late Middle Triassic interval of perhaps 10–15 Myr. This compilation was a preliminary attempt to compile a standard Middle Triassic GPTS.

Research by different groups continued since then in both Tethyan and Boreal marine realms (Nawrocki and Szulc 2000; Hounslow and McIntosh 2003; Muttoni et al. 2004a; Gradinaru et al. 2007; Hounslow et al. 2007) as well as in the Germanic (Central European) Basin (Szurlies et al. 2003; Szurlies 2007). As a result, the sequence of polarity reversals for some Middle Triassic time intervals is relatively well established. For example, excellent matching of magnetostratigraphies has been found for the Olenekian-Anisian boundary interval between Albania (Muttoni et al. 1996) and Romania (Gradinaru et al. 2007). Another illustrative example of a laterally reproducible magnetostratigraphy comes from Anisian-Ladinian boundary sections from the Dolomites, Italy (Muttoni et al. 2004a), and is briefly summarised hereafter.

Magnetostratigraphic investigations on biostratigraphically dated Tethyan limestones and radioisotopically dated tuff intervals of Middle Triassic age from the Dolomites started in the late 1,990 s (Muttoni et al. 1997). A ~ 110 m long core was drilled at Mount Seceda in the northwestern Dolomites (Brack et al. 2000). With over 90 % recovery, the core offered a unique opportunity to reconstruct a consistent portion of the Middle Triassic time scale in stratigraphic continuity (Muttoni et al. 2004a). The conodont biostratigraphy of the laterally equivalent outcrop section could be integrated by means of magneto- and lithostratigraphic correlations (Fig. 7). A logical next step was the correlation with data from previously studied sections from the Dolomites (Frötschbach, Pedraces, Belvedere), but also from Trentino (Margon-Val Gola), and from the Brescian Alps (Bagolino), as is illustrated in figure 8.

The Seceda core spans a complete succession of Buchenstein Beds limestone members and associated Pietra Verde volcaniclastic layers. Two ash layers located in these intervals yielded accurate U-Pb age data (Mundil et al. 1996, Brack et al. 1996), indicating an average sediment accumulation rate of ~ 1 cm/kyr (Fig. 7). Palaeomagnetic analyses were performed on numerous samples from the oriented core. A characteristic remanent magnetisation component could be established and rock magnetic experiments supported its primary origin (Muttoni et al. 2004a). Moreover, the palaeomagnetic mean directions from the Seceda core and from the Frötschbach, Pedraces and Belvedere sections (Brack and Muttoni 2000) pass the fold test, giving even more confidence in the origin of the palaeomagnetic signal. These observations support the successful magneto- and lithostratigraphic correlations between distant sections and suggest that the Buchenstein Beds carry an original Triassic magnetisation, acquired well before the Cenozoic Alpine deformation.

According to the general palaeogeographic evolution of Pangea during the Permian and Triassic, the Dolomites as part of the African promontory of Adria were located in the northern hemisphere during Middle Triassic times, an assessment that is essential for interpreting magnetic polarity. A sequence of 24 polarity zones was recognized at Seceda (Fig. 7) and was successfully correlated to magnetostratigraphic sequences developed in other sections from the Dolomites and Trentino by using also laterally traceable lithostratigraphic marker beds such as tuff levels (Tc, Td, Te) and limestone beds (Fig. 8). This integrated stratigraphy allowed piecing together sections across virtually the entire eastern Southern Alps, and allowed the construction of a reference magnetostratigraphy. Merging the U-Pb dates from Seceda and the faunal associations from all sections notably augmented the numerical and biostratigraphic definition of the Anisian-Ladinian boundary interval. The resulting composite sequence was found to cover an Anisian-Ladinian time span of ~ 4 Myr, where geochronological and magnetostratigraphic control indicate a frequency of 4 reversals per Myr.

The integrated Anisian-Ladinian chronology contributed to resolve the “Latemar controversy”: a debate on the duration of deposition of the Latemar carbonate

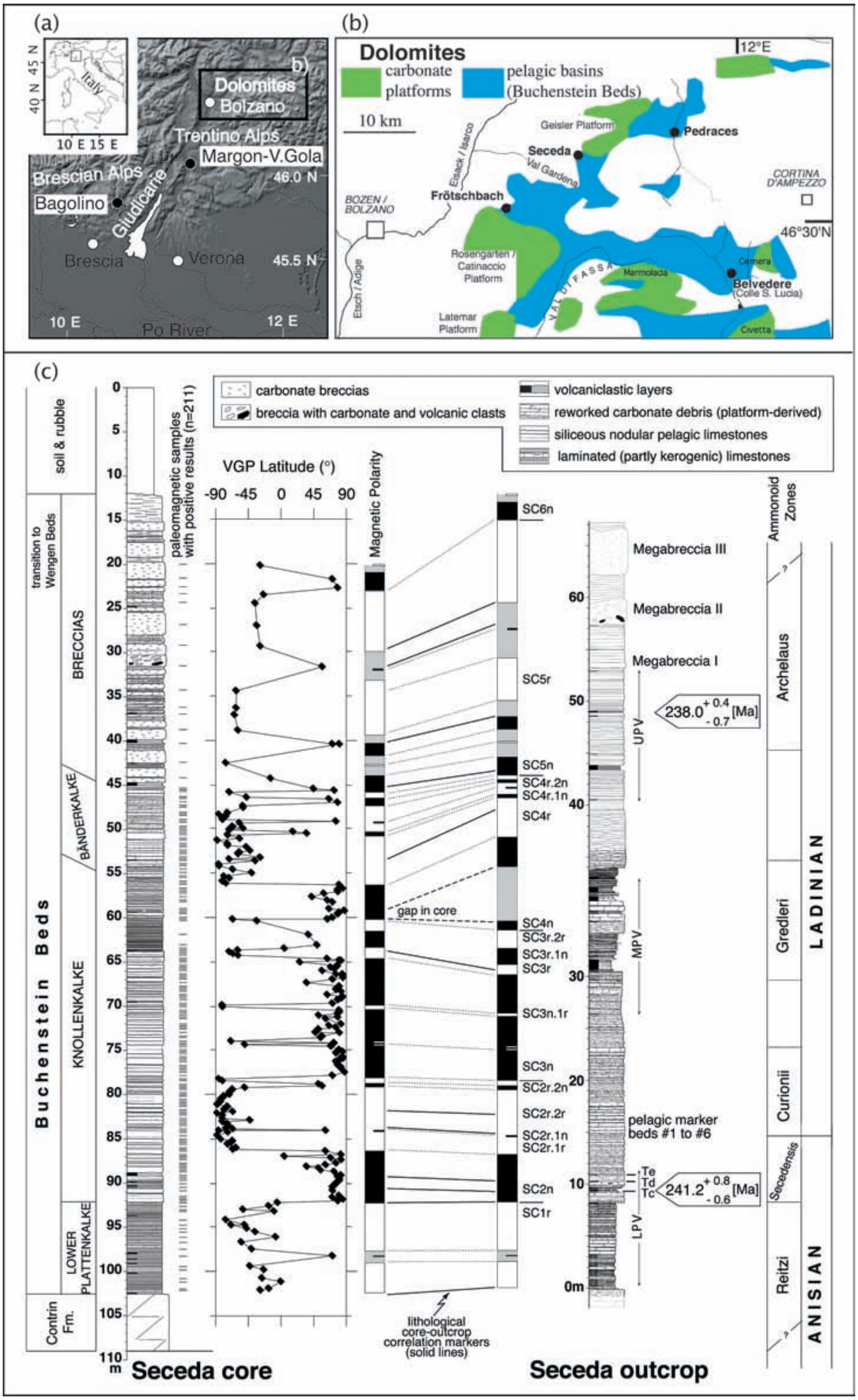


Fig. 7. A) Alpine region with location of the stratigraphic sections: Seceda, Frötschbach, Pedraces, Belvedere, and Rosengarten (Dolomites); Margon-Val Gola (Trentino); Bagolino (Brescian Alps). B) Sections in the Dolomites are placed with respect to the distribution of Ladinian carbonate platforms and pelagic basins. C) Lithology and magnetostratigraphy of the Seceda core and outcrop (Muttoni et al. 2004a). VGP latitudes are derived from the characteristic magnetic component; black (white) is normal (reverse) polarity, grey represents intervals with no data. U-Pb single zircon age data are from Mundil et al. (1996).

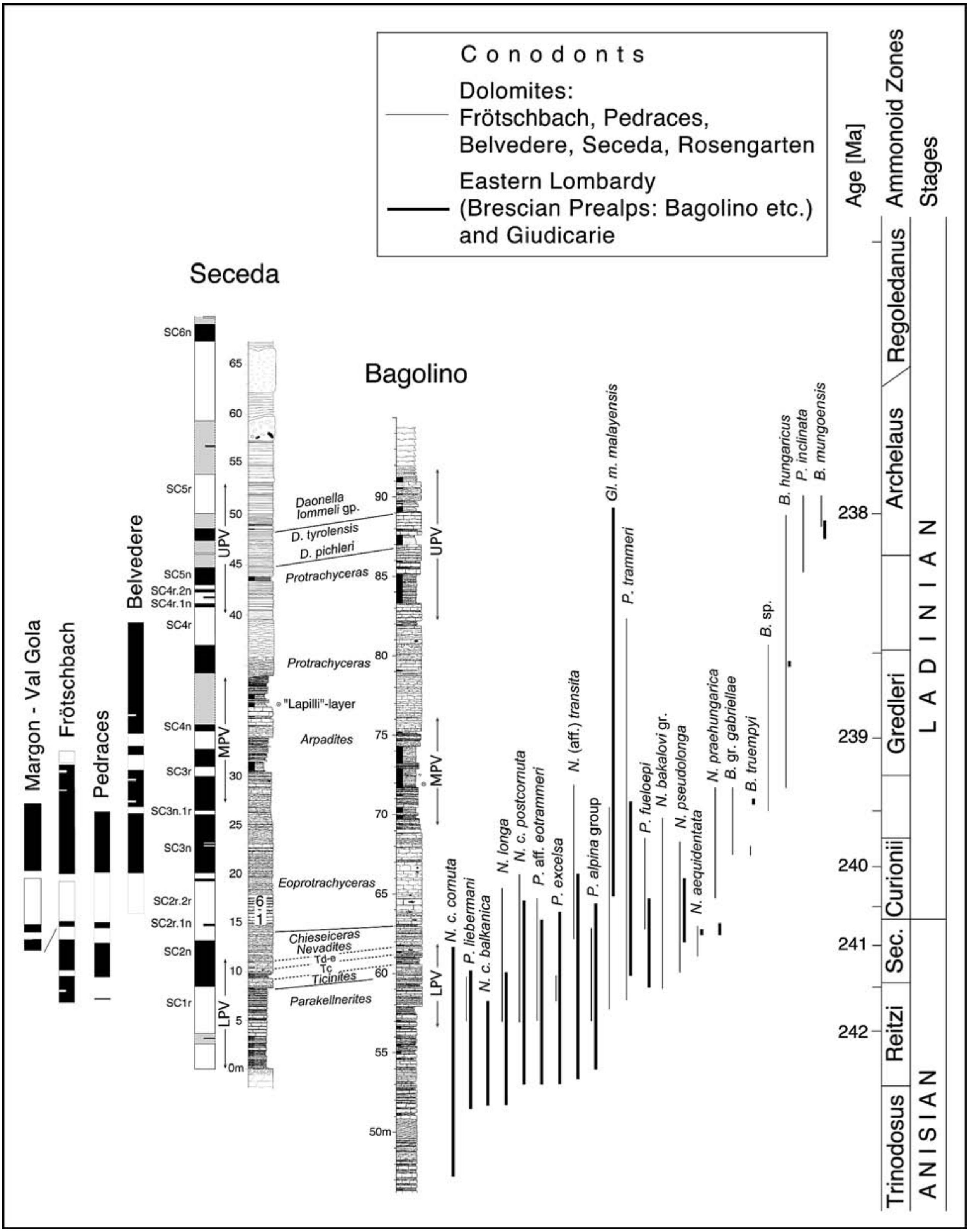


Fig. 8. Integrated Anisian-Ladinian boundary stratigraphy and biochronology. Biostratigraphic data from the Dolomites are projected onto the Seceda outcrop reference stratigraphy, while those from the Brescian Alps and Giudicarie are projected onto the Bagolino reference stratigraphy. Numerical ages are derived from interpolation of dates from Mundil et al. (1996). For details, see Muttoni et al. (2004a) and references therein.

platform from the Dolomites (Fig. 7b). The platform interior – a 470 m thick lagoonal succession of ~ 600 shallowing-upward cycles – was attributed a 9–12 Myr record of precessional forcing of sea level change (Hinnov and Goldhammer 1991; Preto et al. 2001). However, the U-Pb dating of volcanoclastic layers suggests instead that the ~ 600 Latemar cycles cover only a few million years (Brack et al. 1996; Mundil et al. 1996, 2003). An independent astrochronological interpretation of sedimentary cycles in the Muschelkalk of Central Europe coupled with biostratigraphic correlations with Tethyan successions suggest that the Latemar sequence was deposited in maximum 2.6 Myr (Menning et al. 2005). Interestingly, two sophisticated modern techniques to measure geologic time – astrochronology and U-Pb single-crystal zircon dating – lead to age estimates of the duration of the Latemar carbonate platform that differ by almost one order of magnitude: a significant discrepancy.

Kent et al. (2004) performed a magnetostratigraphic analysis of the entire Latemar lagoonal succession, which indicated that most of the succession is of normal magnetic polarity. Although the effects of lightnings and possible thermochemical overprints complicate the picture, Kent et al. (2004) regarded the Latemar results to represent the original polarity of the geomagnetic field. The predominant normal polarity together with biostratigraphic and lithostratigraphic correlations with the adjacent Buchenstein Beds basin led them to consider the Latemar deposition as coeval with Chron SC2n at Seceda (Fig. 7). Considering that no significant normal polarity bias or low reversal frequency have been documented for the Middle Triassic in global palaeomagnetic compilations, Kent et al. (2004) concluded that Chron SC2n at Seceda and the time-equivalent Latemar deposits cannot be anywhere near as long as 9–12 Myr as was implied by the original cyclostratigraphic interpretation.

A duration of ~ 1 Myr for Chron SC2n derives from a straightforward interpretation of the U-Pb age model for Buchenstein deposition. This age model is compatible with all other age constraints, except those of the Latemar. Based on this and other arguments discussed in Kent et al. (2004), and a follow-up comment by Hinnov (2006), Kent et al. (2006) argued that the long duration of Latemar deposition suggested by conventional cycle counting is in error. The recent Concise GTS2008 time scale (Ogg et al. 2008), which supersedes the GTS2004 time scale (Gradstein et al. 2004), now uses the U-Pb data from the Dolomites as age constraints for the Middle Triassic.

Palaeozoic case study: The Carboniferous-Permian Reversed Superchron and Late Permian magneto-cyclostratigraphy of the Central European Basin

Late Carboniferous and Permian rocks were already investigated in various places around the world in the 1950's and 1960's because many of them have a stable and strong magnetisation. These early results indicated that North America, Europe, and several parts of Asia were positioned significantly more to the south than nowadays, and that opposite rotations had occurred in North America and Europe caused by the opening of the North Atlantic. These observations significantly contributed to the development of the theory of plate tectonics. In addition, the dominance of reverse polarity in Permo-Carboniferous rocks led to propose the existence of the Kiaman Magnetic Interval (Irving and Parry 1963) or Carboniferous-Permian Reversed Superchron (CPRS).

The lower boundary of the CPRS was originally named Patterson Reversal (Irving and Parry 1963), until Opdyke et al. (2000) detected normal polarity in younger rocks of the same area. A global analysis and synthesis of 27 magnetostratigraphic records of Carboniferous age has shown that there is at present no consensus on where to place the base of the CPRS: it has been observed in early Bashkirian (SE Australia), late Bashkirian (eastern Canada), early Moscovian (Central Europe), and middle Moscovian rocks (Ukraine) (see Figs. 9, 10). In the GTS2004, the boundary is rather arbitrarily placed close to the Moscovian-Kasimovian boundary at 306.7 Ma, within a reverse polarity interval (Davydov et al. 2004). It would instead be more logical to place the base of the CPRS at the youngest level bearing normal polarity in the Moscovian (middle Kashirian of eastern Europe; Khramov et al. 1974), which would result in an age of ~ 310 Ma according to GTS 2004, Alternatively, the base of the CPRS could be placed at the top of the youngest normal polarity chron in the late Baskirian around 312 Ma. The most recent update of the geological time scale (Ogg et al. 2008), accessible through Time Scale Creator 4.0 (www.tscreator.com), assumes that nearly all short normal subchrons reported between 314 and 299 Ma are questionable, and a result, it places the base of the CPRS at 313.6 Ma within the late Bashkirian (Fig. 10). Approximately five short normal polarity zones could be pres-

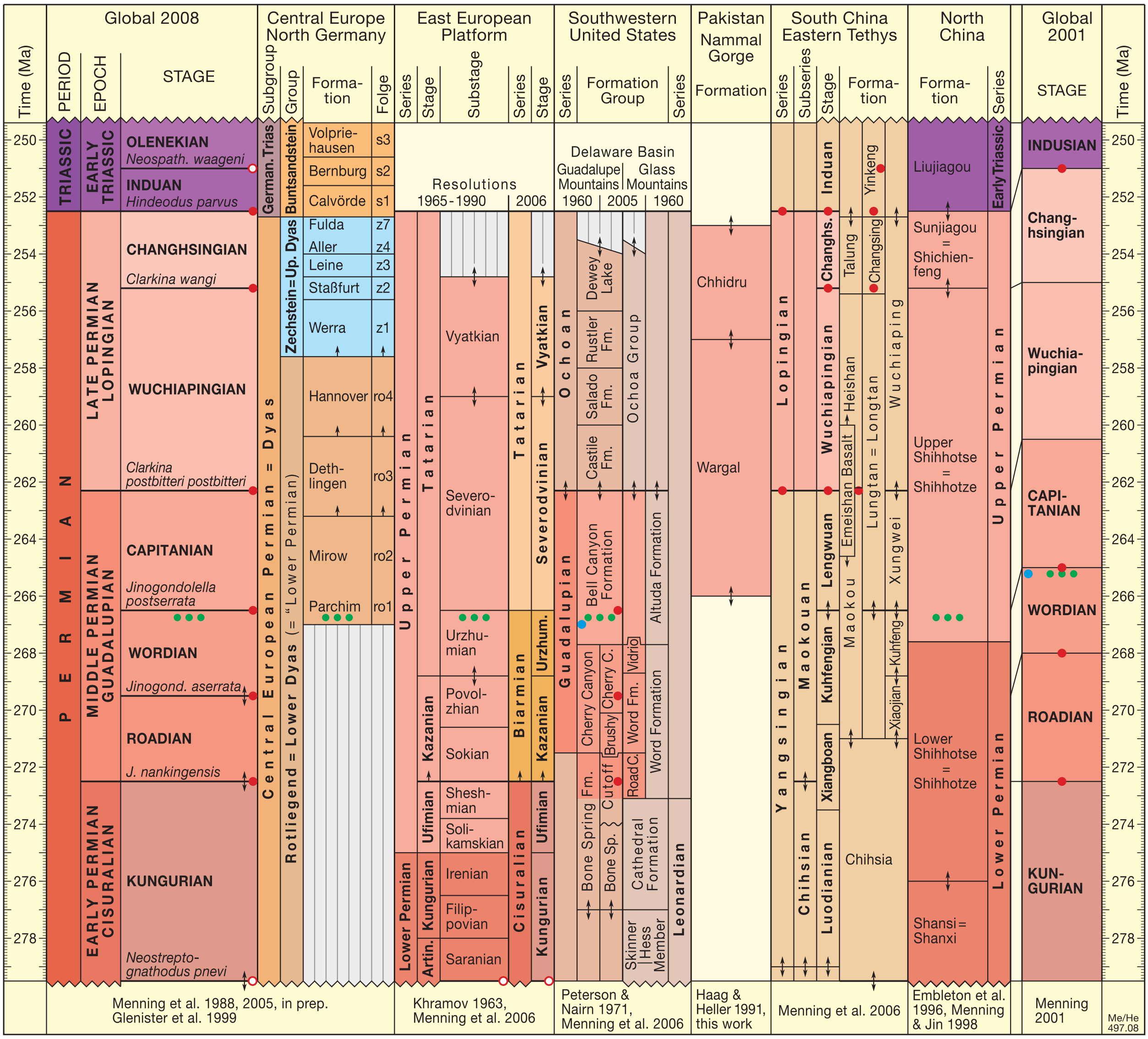


Fig. 9. Permian Global Stratigraphic Scale as used in the Devonian-Carboniferous-Permian Correlation Chart 2006 (DCP 2006) and Regional Stratigraphic Scales. Full red dots: GSSP; open red dots: GSSP proposed; green dots: Illawarra Reversal; blue dot: Pb-U age of Bowring et al. 1988; arrows: uncertain position of a boundary according to the numerical time scale or a global stage boundary. In the global column the index fossils for the stage boundaries are presented as adopted respectively proposed. The terms of the Global Stratigraphic Scale (GSS) are in capital letters to distinguish them from regional terms from which they were derived: unfortunately, commonly they cover different time spans.

ent within the CPRS (Menning 1995), but these are now reduced to two reliable zones in Time Scale Creator 4.0, although no clear argumentation or references are given in support of this interpretation. Clearly, the lower boundary of the CPRS requires verification in multiple, globally distributed sections.

The upper boundary of the CPRS was named Illawarra Reversal (Irving and Parry 1963). Initially, it seemed to occur near the Permian-Triassic boundary, and this led to speculations on a connection with the late Permian mass extinction (Peterson and Nairn 1971). This inference ignored the well-documented position of the Illawarra Reversal in the lower Tatarian of eastern Europe (Khramov 1963), being significantly older than the Permian-Triassic boundary. Later, varying positions have been suggested: Kungurian, Ufimian, Kazanian of eastern Europe, early Wordian, late Wordian, early Capitanian (see Fig. 10). These positions are not based on new field and laboratory work but mainly on re-interpretations of existing data combined with successive geological time scales that have significantly different numerical ages. The Illawarra Reversal is a first class time marker when detected reliably and it is the anchor point for the middle and late Permian part of the Devonian-Carboniferous-Permian Correlation Chart (DCP 2003; Menning et al. 2006; DCP 2006, Fig. 9). For most of the Permian, however, the absence of reversals excludes the use of magnetostratigraphy, and other stratigraphic techniques are required. Successions of different environments (e. g. marine, limnic, fluvial, sabkha, volcanic) can then for example be correlated through climatic change. Recently, a first attempt at such a correlation between European Permian sections was carried out based on climatic and environmental information (Schneider et al. 2006).

The relative age of the Illawarra Reversal was estimated to be 10 ± 4 Myr older than the Permian-Triassic boundary (Menning 1986), which has an U-Pb SHRIMP age of 251.2 ± 3.4 Ma (Claoué-Long et al. 1991). Meanwhile, the age of the Permian-Triassic boundary was lowered to ~ 252.5 Ma based on a U/Pb age of 252.6 ± 0.2 Ma from a tuff layer located slightly below the boundary (Mundil et al. 2004). A volcanic tuff from just below the Illawarra Reversal at Nipple Hill (Texas) provided an U-Pb age of 265.3 ± 0.2 Ma (Bowring et al. 1998). Using the ages of ~ 310 (or 314) Ma for the base and ~ 265 Ma for the top, results in a duration of 45 (or 50) Myr for the CPRS. The positions of the boundaries of the PCRS compare well with those of Irving (1971), but he arrived at a longer duration of 60 Myr (290–230 Ma) based on the time scales of his time. The CPRS is thus the longest time span of the Phanerozoic Eon in which one polarity dominates. This duration is some 5–10 Myr longer that of its normal polarity pendant, the Cretaceous Normal Superchron, ranging 124.5–84 Ma (Ogg et al. 2004)

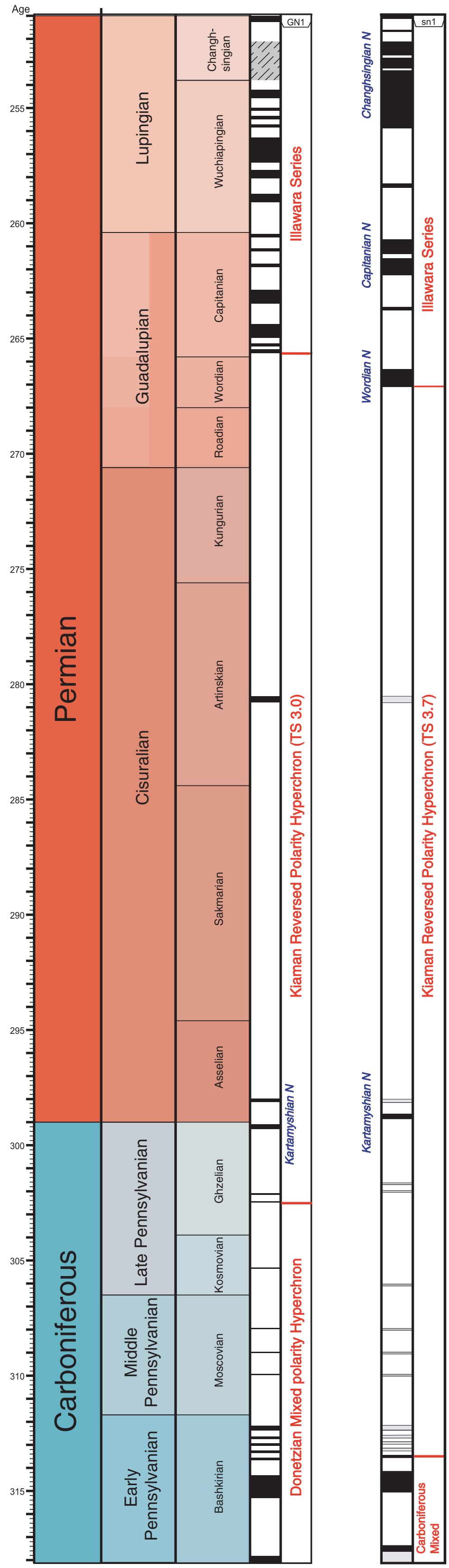


Fig. 10. Two versions of the Carboniferous-Permian Reversed Superchron, on the left the polarity time scale according to GTS2004 (Gradstein et al. 2004), made with Time Scale Creator 3.0, i. e. according to the 2004 version. On the right according to Time Scale Creator 4.0 (www.tscreator.org), providing an update of GTS2004, i. e. the Concise GTS2008 (Ogg et al. 2008). It appears that in the latest version the subchrons preceeding the lower boundary CPRS of GTS2004 are considered uncertain. Also the Illawarra series postdating the PCRS has significantly changed in both timing and number of reversals, according to a new compilation of Steiner (2006).

A reliable polarity time scale for post-Illawarra times spanning 265–252.5 Ma is not yet available, but in any case, it may contain as much as 15 polarity zones (Menning 1995). In the continental record, 10 zones have been detected in the upper Rotliegend of Central Europe (Menning et al. 1988) and 4 zones in the upper Zechstein (Szurlies et al. 2003). The lower Zechstein, however, has not been investigated in detail. In the marine record, the most significant results come from Nammal Gorge in Pakistan where 10 polarity zones have been detected in the late Permian (Haag and Heller 1991). A recent compilation by Steiner (2006) of approximately 27 published middle to late Permian magnetostratigraphies suggests some 10 polarity zones in the late Permian. Unfortunately, no short-living fossils are available to integrate magnetostratigraphic results from marine and continental successions with any acceptable level of accuracy. Reliable climatic/cyclostratigraphic records are needed to resolve these correlation uncertainties (e. g. Schneider et al. 2006), together with more magnetostratigraphic studies.

Carboniferous to Triassic rock sequences in the Central European Basin are up to 6,000 m thick. They comprise a unique and complete late Permian succession without gaps longer than 100 kyr. In northeast Germany, borehole Mirow 1/1a/74 cored 1,600 m of the 3,300 m Rotliegend succession and bottomed in volcanics of the Rotliegend Group (latest Carboniferous-Permian, 300–258 Ma). Here, we reinterpret the magnetostratigraphic results from the upper 1290 m thick part of the Rotliegend sediments. The sedimentary cycles observed in the perennial Rotliegend salt lake in northwest Germany by Gast (1995) can be correlated to northeast Germany (Gebhardt 1995) and thus to the magnetostratigraphy of borehole Mirow 1/1a/74. This allows calibration of the regional GPTS for the Late Permian based on cycle scaling.

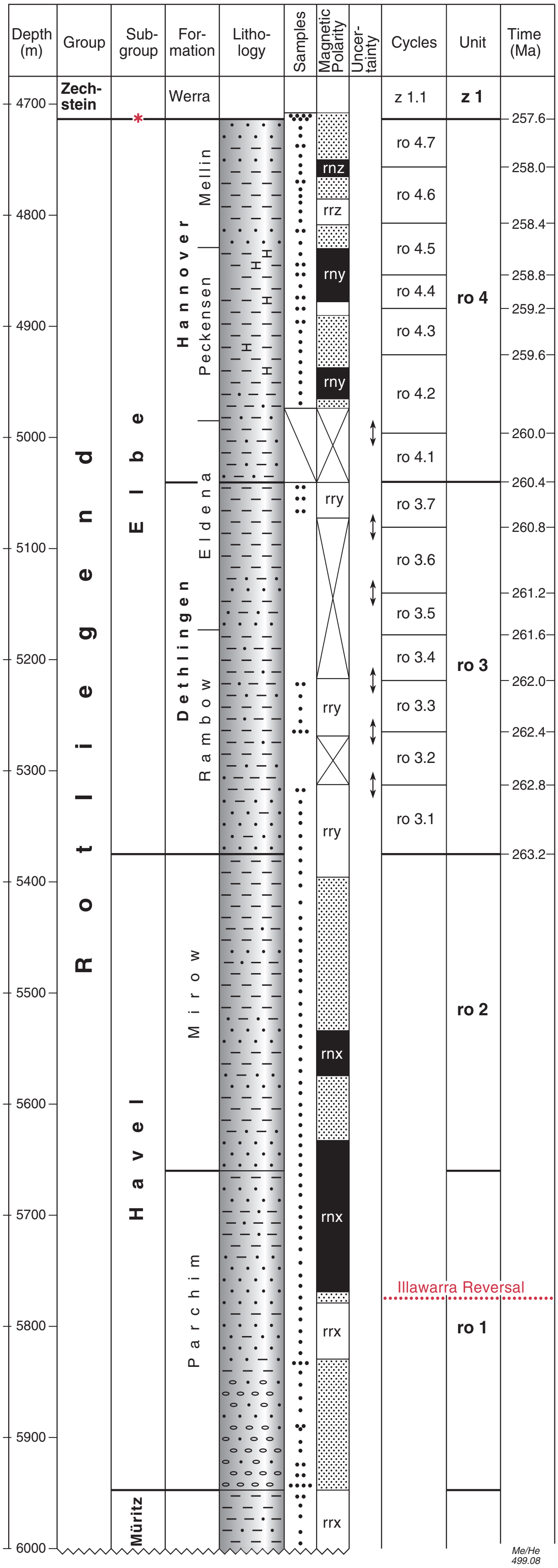
The cycles of the Elbe Subgroup (latest Rotliegend) are interpreted as orbitally forced cycles (Gast 1995). They can only correspond to the long eccentricity cycle of ~ 400 kyr, because ~ 100 kyr eccentricity cycles cannot fill in the available time span. In the Elbe Subgroup, 14 such cycles were detected, corresponding to a total duration of ~ 5.6 Myr (Fig. 11). Unfortunately, the duration of the post-Illawarra magnetic zones can only be estimated roughly, because of intervals with uncertain polarity and without core material. Until now, the longest zone is the upper part of zone rny (cycles ro4.3 to ro4.5), which is estimated to have a minimum duration of 700 kyr (Fig. 11). Future work should be directed to study the 14 cycles of the Elbe Subgroups in detail to better determine the exact number of magnetic zones and their duration.

Part three – Discussion

Integrated stratigraphy and marine-continental correlations

Several examples are provided in the literature of good correlation of land-derived magnetostratigraphies with the polarity record retrieved from marine magnetic anomalies, which prove that geomagnetic polarity reversals are synchronous and global. This synchrony distinguishes magnetostratigraphy from biostratigraphy, bearing in mind that polarity reversals are fundamentally of a binary nature while biostratigrapic zones have a distinctive character. The first magnetic stratigraphies in sedimentary rocks (Creer et al. 1954; Irving and Runcorn 1957) documented normal and reverse polarities in the Proterozoic Torridonian Sandstones in Scotland, as well as in rocks of Devonian and Triassic age. These studies were conducted on poorly fossiliferous sediments that possessed however a magnetisation intensity measurable by the magnetometers available at that time, and as a consequence, correlations of polarity zones therein retrieved could not be supported by biostratigraphic data.

The modern era of magnetostratigraphy integrated with biostratigraphy started with the early studies on Plio-Pleistocene marine sediments and deep-sea cores (e. g., Opdyke 1972) and continued over the ensuing years with classic studies on e. g. the Cretaceous-Paleocene successions of the Central Apennines at Gubbio, Italy (e. g. Alvarez et al. 1977). This multidisciplinary approach has since been significantly refined and has extended the GPTS originally derived from palaeomagnetic studies on basaltic outcrops and marine magnetic anomalies. *Vice versa*, magnetostratigraphy may fundamentally help dating biostratigraphic zonations that have at best regional significance, like in the Para-



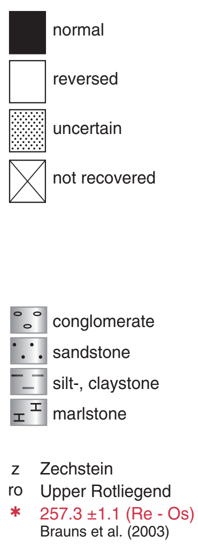


Fig. 11. Succession of the late Permian in northeast Germany, borehole Mirow 1/1a/74 cored 1,600 m of the 3,300 m Rotliegend succession and bottoms in volcanics of the Rotliegend Group. The Rotliegend succession was investigated with the purpose to detect the Illawarra reversal (Menning et al. 1988). Here, we reinterpret the magnetostratigraphic results from the upper 1,290 m thick part of the Rotliegend sediments; rnx, rrx, etc. refer to informally labeled polarity zones. The sedimentary cycles observed in the perennial Rotliegend salt lake in northwest Germany by Gast (1995) can be correlated to northeast Germany (Gebhardt 1995) using gamma-ray logs and thus to the magnetostratigraphy of bore hole Mirow 1/1a/74. This allows calibration of the regional GPTS for the Late Permian based on cycle scaling.

tethys where zonations based on ostracods or molluscs are primarily of environmental significance rather than being chronostratigraphic tools (Vasiliev et al. 2004).

In addition to the advent of astrochronology, another important stratigraphic tool comes from chemostratigraphy or isotope stratigraphy, in particular from stable isotopes (see Weissert et al. 2008). The integrated use of magnetostratigraphy, biostratigraphy, cyclostratigraphy, isotope stratigraphy, and radioisotopic dating from selected sections or cores contributes to refine the GPTS by continuously providing accurate age determinations of the constituent polarity reversals.

A major problem in correlating marine sequences with their time-equivalent continental sequences is their fundamentally different fossil content. Marine transgressions within a continental sequence may give a first-order constraint, but such sequences are relatively rare. Here, magnetostratigraphy plays a crucial role, preferably aided by cyclostratigraphic constraints. A first example of a truly bed-to-bed correlation – at the precessional level – between the marine and continental realm was given by Van Vugt et al. (1998), who correlated early Pliocene lacustrine deposits in NW Greece with the marine Trubi formation in the central Mediterranean. Also the Miocene continental deposits of northern Spain have now been successfully correlated to time-equivalent marine successions in Italy (Abdul Aziz et al. 2004; Hüsing et al. 2007), and the ages of individual polarity reversals younger than 11 Ma are virtually indistinguishable within error resolution.

It must be noted that such detailed correlations are not – and can never be – based on magnetostratigraphic tie-points alone: positions of reversals may differ due to differential/delayed lock-in depth/time because of different environments and recording quality. Hence, although magnetostratigaphy may well serve as a first-order correlation, the final correlation is based on astronomical fine-tuning. This concept has also been applied on the Chinese loess sequences (Heslop et al. 2000), showing that recorded polarity reversals showed considerable delay because of a large lock-in depth in such aeolian/palaeosol deposits. This concept is still not widely accepted, however, and some studies have used magnetostratigraphic tie-points, which in some cases may cause the same reversal to occur both in a warm, interglacial stage in one locality, and in a cool stage in another place. This would make cyclostratigraphy a useless exercise and argues once again for an integrated stratigraphic approach in order to avoid such pitfalls.

Resolution: subchrons, cryptochrons, reversals, excursions, hiatuses

The International Commission on Stratigraphy has guided the use of magnetostratigraphic units (polarity zones), their time equivalents (polarity chrons), and chronostratigraphic units (polarity chronozones) (Anonymous 1977; Opdyke and Channell 1996). During the last two decades, the recognition of numerous short-lived polarity intervals or excursions (Langereis et al. 1997; Krijgsman and Kent 2004) as part of the Earth's magnetic field record requires an extension of the current terminology. In this respect, Laj and Channell (2007) argued to include excursions and brief polarity microchrons (Table 1) in the terminology.

Although the resolution of a magnetostratigraphic study largely depends on sampling resolution, there are some clear limitations regarding the (temporal) resolution that can be achieved. For a 'simple' first-order magnetostratigraphy, this may not be a serious obstacle since even in slowly accumulating sediments (< 1 cm/kyr), the required first order time resolution – say, a minimum of 15–20 kyr sampling resolution, but preferably better than that – can be usually achieved. This minimum sampling resolution derives from the typical cut-off of 30 kyr for tiny wiggles/cryptochrons in marine magnetic anomaly studies.

Establishing a correct terminology of polarity events of very short duration is a challenging problem since short records are frequently compromised by limitations of the recording medium, while also estimates of duration strongly depend on the available chronological tools. For magnetostratigraphy, Krijgsman and Kent (2004) advocated a duration cutoff for separating 'subchrons' and 'excursions', at 9–15 kyr, while Abdul Aziz and Langereis (2004) argued for a minimum duration of approximately 10 kyr for the entire interval of excursional (or transitional) directions and opposite polarity. In practice, however, such a definition will meet some serious problems, since in most cases the chronostratigraphic precision will be insufficient. The limitations inherent to sedimentary NRM acquisition processes require the use of records with high sedimentation rates, since delayed lock-in of the NRM and the resulting filtering process will often prevent these short phenomena to be recorded (Roberts and Winkelhofer 2004). An additional complication of delayed lock-in of the NRM is that the position of polarity reversals may be displaced to older levels (Van

Hoof and Langereis 1991). Indeed, detailed studies of time equivalent sections in the Mediterranean area have often shown small discrepancies in the (astrochronologically tuned) position of polarity reversals, on the order of 1–2 precessional cycle(s). In general, however, these discrepancies do not seriously disturb the chronostratigraphy of a succession.

Another complicating factor may be the occurrence of hiatuses or breaks in sedimentation. Recognition of hiatuses requires carefully assessing all available data from integrated stratigraphy and existing proxy

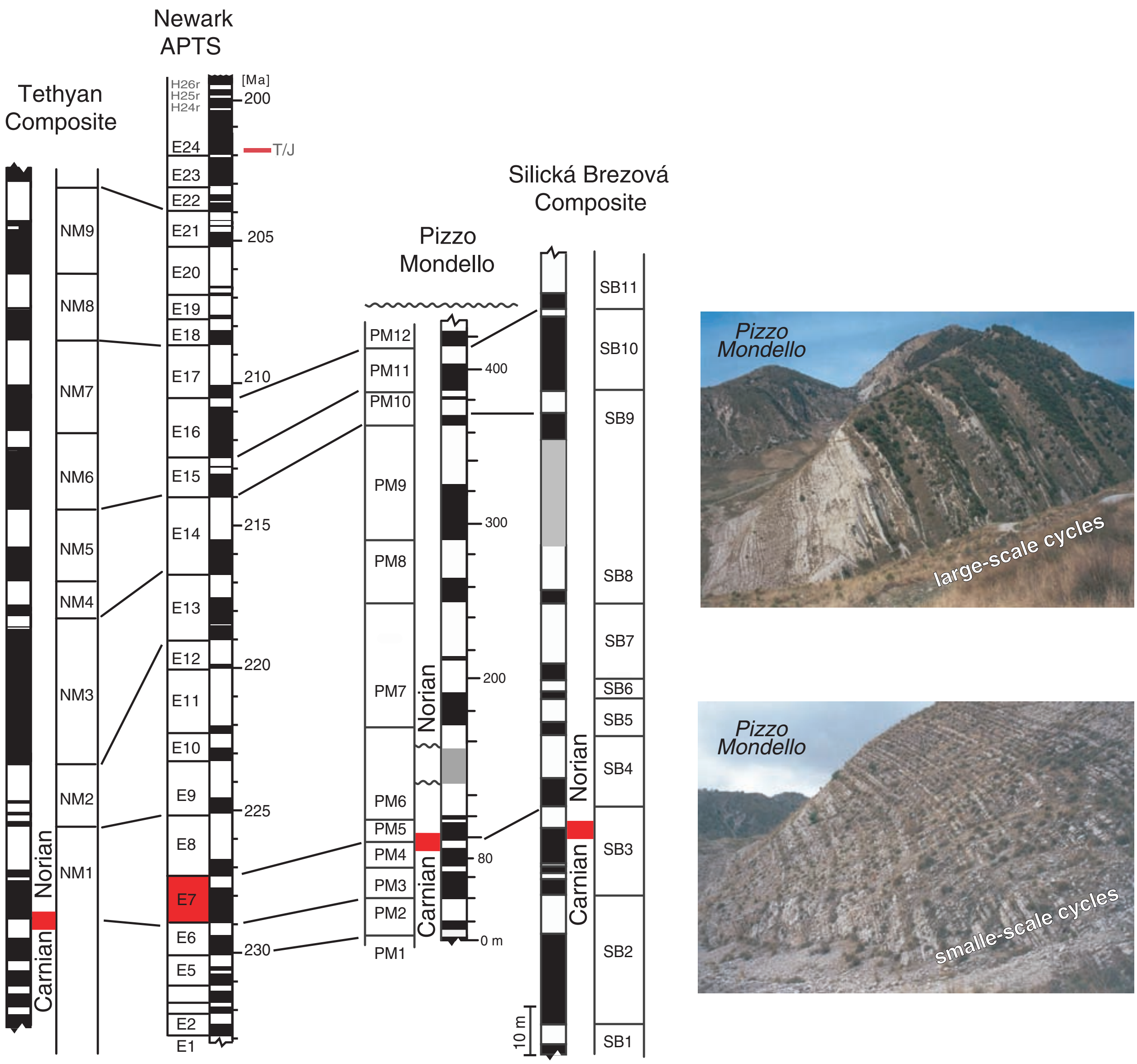


Fig. 12. Magnetostratigraphic correlation of the marine European sections of the Tethys region (Muttoni et al. 2001, 2004b; Krystyn et al. 2002; Gallet et al. 2003; Channell et al. 2003) to the astronomically tuned polarity time scale (APTS) of the Newark basin (Kent and Olsen 1999). Red intervals are the Carnian-Norian boundary intervals. Consensus places the Carnian-Norian boundary broadly within chronozone E7 implying an age between 227 and 229 Ma. The results from the Hartford basin (H-chronozones) from Kent and Olsen (2008) are shown, and the position of the palynological Triassic-Jurassic boundary (T/J) is indicated.
Photos: the pattern of both small and large scale cycles in the Sicilian Pizzo Mondello section seems promising to use cyclostratigraphy and obtain a more detailed correlation to the Newark reference section.

records. A polarity reversal that coincides with a visible change in sedimentary environment (lithology) is usually suspect, and the same applies to any sudden and notable change, e. g. derived from proxy data. Biostratigraphy may help to recognise hiatuses provided they are large enough relative to zonation resolution. Also cyclostratigraphy may be of help to detect hiatuses: although small-scale cycles (e. g. precession) may seem continuous and undisrupted across a hiatus, the logic of cylostratigraphy dictates that all or most Milankovitch cycles (the eccentricity, obliquity, precession periods) can be recognised in their appropriate ratios. Recognition of hiatuses will likely remain an unavoidable obstacle in geology, which may be addressed by correlation of multiple sections over large areas. Because of its binary nature, magnetostratigraphy alone can hardly recognise the presence of hiatuses, but when embedded in an integrated stratigraphic approach, it may contribute to resolve them. Also strongly varying sedimentation rates, usually associated with lithology variations, may distort the observed polarity pattern; hence, providing a detailed lithostratigraphic log of the sampled sequence is an important requirement in any magnetostratigraphic study.

Back to the future?

Future efforts should focus on extending the geomagnetic polarity time scale back in geological time to periods older than the marine magnetic anomaly record, i. e. older than ~ 160 Ma. This is critical not only for correlation and dating purposes, but also to learn more about the characteristics of the field and the geodynamo that generates it. For example, long-term reversal frequency variations – long periods of frequent field reversals versus equally long periods of stable polarity (superchrons) – tell us about the influence of mantle processes on the generation of the field in the liquid outer core, and possibly about the influence of (nucleation and growth of) the solid inner core.

It is evident that by going back in time, the record becomes increasingly sparse while it also deteriorates: the older the rock, the less the chance to be a pristine carrier of the original magnetisation acquired during its formation. Nevertheless, very ancient but reliable records – although scarce – of Archean and Proterozoic age have been acquired. The oldest known reversal that is reliably documented – by a positive reversal test, fold test, and an intra-formational conglomerate test – has an age of ~ 2.8 Ga (Strik et al. 2003). There are reports of reversals as old as ~ 3.2 Ga (Layer et al. 1998), but their reliability still requires substantiation by a positive field test. Recently, it has been suggested that there is a very long trend – since the Archean – of reduced stability of the field, possibly related to a growing inner core, leading to a decrease in reversal frequency back in time (Biggin et al. 2008). Clearly, these rare observations of the ancient field are still far from being fit for correlation or dating purposes.

So, how far can we go back in time? One of the greatest achievements of early Mesozoic magnetostratigraphy is represented by the Late Triassic Newark astrochronological polarity time scale (APTS; Kent et al. 1995; Kent and Olsen 1999; Olsen and Kent 1999), which was constructed by anchoring magnetostratigraphy and Milankovitch chronostratigraphy to the radioisotopically dated Orange Mountain basalt (Fig. 12), and which was recently extended up to the Early Jurassic by including data from the Hartford Basin (Kent and Olsen 2008). The correlation of the Newark APTS to the Triassic sections of the Tethyan marine realm is still far from straightforward, as evidenced by the continuously changing interpretations as new results are acquired, although there is now good consensus on the age of the Carnian-Norian boundary around 227–228 Ma (Fig. 12). Many Tethyan Triassic sections still depend on biostratigraphic dating often assuming equal duration of biozones through lack of other dating methods. New radioisotopic dates of Triassic sections are forthcoming, however, and aid in providing better constraints on duration of ammonite zones (e. g. Ovtcharova et al. 2006). In addition, there are sections, like the late Triassic Pizzo Mondello section in Sicily (Muttoni et al. 2004b), that seems to promise obtaining a cyclostratigraphic framework and hence a correlation with the Newark basin sequence. Much work in this respect has still to be done.

The often discontinuous nature of Permian deposits and the difficulty to correlate marine and continental sections have as yet limited the construction of a complete Permian polarity time scale. As we have seen in our case study, the polarity patterns predating and postdating the Carboniferous-Permian Superchron still require verification.

Older records documenting a coherent pattern of reversals may occasionally go back as far in time as the Cambrian to Ordovician (e. g. Kirschvink 1978; Kirschvink and Rozanov 1984; Gallet and Pavlov 1996), but are of little use for magnetostratigraphic correlation purposes. New prospects in magnetostratigraphy must rely on ‘automatic’ integration with biostratigraphy and cyclostratigraphy for every sedimentary sequence under investigation, augmented by other palaeoclimatic and palaeoenvironmental proxy records.

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