that, at least against fully converged, adequately resolved simulations of a local or regional character. If, within the acceptable parameter range identified, it is found that the original model no longer gives adequate large-scale predictions, then there are more basic problems to be addressed.

Summary

From the numerical side, no computer improvements that can be seen on the horizon seem likely to make reasonably ambitious GCMs accessible to rigorous and extensive parametric and numerical exploration, a prerequisite to their complete understanding. From the mathematical side, it seems to be our fundamental ignorance about turbulence that most severely restricts the range of our grasp, leaving us with an often painfully narrow range of computations to which theoretical remarks can be significantly addressed. For these structural reasons, the gulf between theory and much numerical modeling will probably continue to widen for the foreseeable future, and thus there may grow to be—indeed some would say it already exists—a division akin to C.P. Snow's 'Two Cultures'.

All the cautions about GCMs notwithstanding, they have become an integral part of the study of physical oceanography. With due regard for the novel capacities and limitations of numerical models, such scientific progress as we do make will more and more often hinge upon judicious computation.

See also

Deep Convection. Double-diffusive Convection. Forward Problem in Numerical Models. Thermohaline Circulation. Wind Driven Circulation.

Further Reading

The literature on ocean modeling is not yet productive of definitive treatises, in large measure because the field is yet young and rapidly evolving. Thus in lieu of textbooks or similar references, the reader is directed to the following series of articles.

For some predictions on the perennially intriguing issue of what improvements in large-scale modeling may be driven by plausible increases in computing speed with massively parallel machines see

Semtner A (2000) Ocean and climate modeling. *Communications of the ACM* 43 (4): 81–89.

For a look back at the history of one of the single most influential models in physical oceanography, see A.J. Semtner's Introduction to 'A numerical method for the study of the circulation of the World Ocean', which accompanies the reprinting of Kirk Bryan's now classic 1969 article of the title indicated. This pair appears back-to-back, beginning on page 149, in *Journal of Computational Physics*, (1997) 135 (2).

General readers may wish to consult the following succinct review, accessible to a broad audience:

Semtner AJ (1995) Modeling ocean circulation. *Science* 269 (5229): 1379–1385.

Finally, for those readers desiring a more in depth appreciation of modeling issues and their implications for specific features of the large scale circulation, consult the careful review

McWilliams JC (1996) Modeling the oceanic general circulation. In: Lumley JL, Van Dyke M, Read HL (eds) *Annual Review of Fluid Mechanics*, Vol. 28, pp. 215–248, Palo Alto, CA: Annual Reviews.

GEOMAGNETIC POLARITY TIMESCALE

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doi:10.1006/rwos.2001.0244

Introduction

Dating and time control are essential in all geoscientific disciplines, since they allow us to date, and hence correlate, rock sequences from widely different geographical localities and from different (marine and continental) realms. Moreover, accurate time control allows to understand rates of change and thus helps in determining the underlying processes and mechanisms that explain our observations. Biostratigraphy of different faunal and floral systems has been used since the 1840s as a powerful correlation tool giving the geological age of sedimentary rocks. Radiometric dating, originally applied mostly to igneous rocks, has provided numerical ages; this method has become increasingly sophisticated and can now—in favorable environments—also be used on various isotopic decay systems in sediments. We are concerned with the application of magnetostratigraphy: the recording of

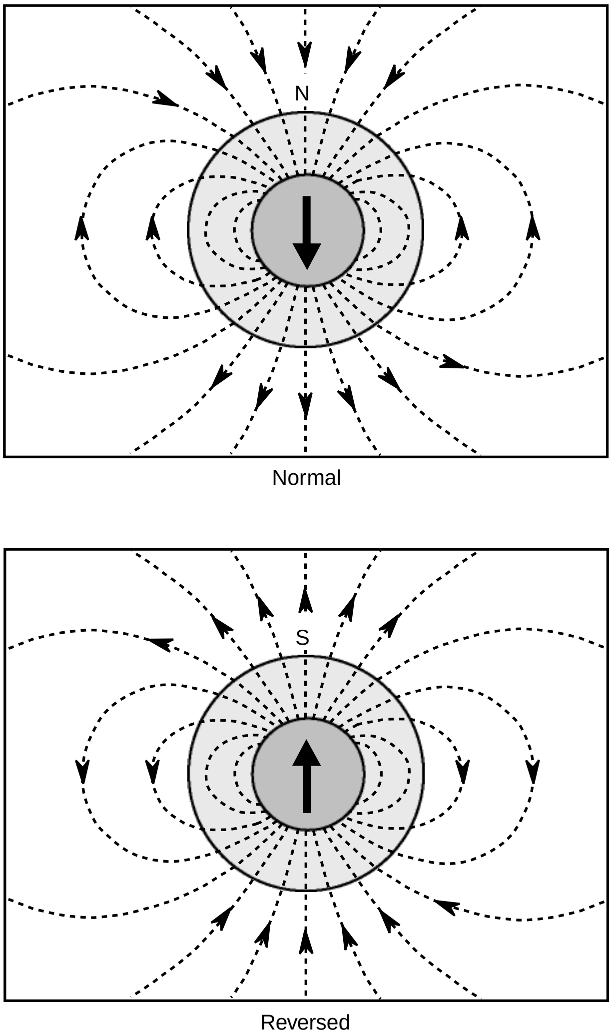

Figure 1 Schematic representation of the geomagnetic field of a geocentric axial dipole ('bar magnet'). During normal polarity of the field the average magnetic north pole is at the geographic north pole, and a compass aligns along magnetic field lines. Historically, we refer to the north pole as the 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.

the ancient geomagnetic field that reveals, in lava piles and sedimentary sequences, intervals with different polarity. This polarity can either be normal, that is parallel to the present-day magnetic field (north-directed), or reversed (south-directed) (**Figure 1**). As a rule, it appears that these successive intervals of different polarity show an irregular thickness pattern, caused by the irregular duration of the successive periods of either normal or reversed polarity of the field. This produces a 'bar code' in the rock record that often is distinctive. Polarity intervals have a mean duration of some 300 000 y during the last 35 My, but large variations occur, from 20 000 y to several million years. If one can construct a calibrated 'standard' or a so-called 'geomagnetic polarity timescale' (GPTS), dated by radiometric methods and/or by orbital tuning, one can match the observed pattern with this standard and hence derive the age of the sediments. Magnetostratigraphy with correlation to the GPTS has become a standard tool in ocean sciences.

The Paleomagnetic 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 conveniently be described as a dipole field, which is equivalent to having a bar magnet at the center of the Earth. Such a dipole accounts for approximately 90% of the observed field; the remaining 10% derives from higher-order terms: the nondipole field. At any one time, the best-fitting geocentric dipole axis does not coincide exactly with the rotational axis of the Earth, but averaged over a few thousand years we may treat the dipole as both geocentric and axial.

The most distinctive property of the Earth's magnetic field is that it can reverse its polarity: the north and south poles interchange. Paleomagnetic studies of igneous rocks provided the first reliable information on reversals. In 1906, Brunhes observed lava flows magnetized 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, Matuyama demonstrated that young Quaternary lavas were magnetized in the same direction as the present field (normal polarity), whereas older lavas were magnetized in the opposite direction. His study must be regarded as the first magnetostratigraphic investigation. Initially, it was believed that the field reversed periodically, but as more (K/Ar dating plus paleomagnetic) results of lava flows became available, it became clear that geomagnetic reversals occur randomly. It is this random character that fortuitously provides the distinctive 'fingerprints' and gives measured polarity sequences their correlative value.

A polarity reversal typically takes several thousands of years, which on geological timescales is short and can be taken as globally synchronous. The field itself is sign invariant: the same configuration of the geodynamo can produce either a normal or reversed polarity. What causes the field to reverse is still the subject of debate, but recent hypotheses suggests that lateral changes in heat flow at the core–mantle boundary play an important role. Although polarity reversals occur at irregular times, over geological time spans the reversal frequency can change considerably. For instance, the polarity reversal frequency has increased from approximately 1 My^{-1} some 80 My ago to 5 My^{-1} in more recent times. During part of the Cretaceous, no reversals occurred at all from 110 to 80 Ma: the field showed a stable normal polarity during some 30 My. Such long periods of stable polarity are called Superchrons, and the one in the Cretaceous is also recognized as the Cretaceous Normal Quiet Zone in ocean-floor magnetic anomalies.

The ancient geomagnetic field can be reconstructed from its recording in rocks during the geological past. Almost every type of rock contains magnetic minerals, usually iron oxide/hydroxides or iron sulfides. During the formation of rocks, these magnetic minerals (or more accurately: their magnetic domains) statistically align with the then ambient field, and will subsequently be 'locked in,' preserving the direction of the field as natural remanent magnetization (NRM): the paleomagnetic signal. The type of NRM depends on the mechanism of recording the geomagnetic signal, and we distinguish three basic types: TRM, CRM, and DRM.

TRM, or thermoremanent magnetization, is the magnetization acquired when a rock cools through the Curie temperature of its magnetic minerals. Curie temperatures of the most common magnetic minerals are typically in the range 350–700°C. Above this temperature, the magnetic domains align instantaneously with the ambient field. Upon cooling, they are locked: the magnetic minerals carry a remanence that usually is very stable over geological time spans. Any subsequent change of the direction of the ambient field cannot change this remanence. Typically, TRM is acquired in igneous rocks.

CRM, or chemical remanent magnetization, is the magnetization acquired when a magnetic mineral grows through a critical 'blocking diameter' or grain size. Below this critical grain size, the magnetic domains can still align with the ambient field; above it, the field will be locked and the acquired remanence may again be stable over billions of years. CRM may be acquired under widely different circumstances, e.g., during slow cooling of intrusive rocks, during metamorphosis or (hydrothermal) fluid migration, but also during late diagenetic processes such as weathering processes through formation of new magnetic minerals. A particularly important mechanism of CRM acquisition occurs in marine sediments during early diagenesis: depending on the redox conditions, iron-bearing minerals may dissolve, and iron may become mobile. If the mobilized iron subsequently encounters oxic conditions, it may precipitate again and form new magnetic minerals that then acquire a CRM.

DRM, or detrital remanent magnetization, is the magnetization acquired when magnetic grains of detrital origin already carrying TRM or CRM are deposited. The 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 (**Figure 2**). Upon compaction and dewatering, the grains are mechanically locked — somewhere in a 'lock-in depth zone' — and will preserve the direction of the ambient field.

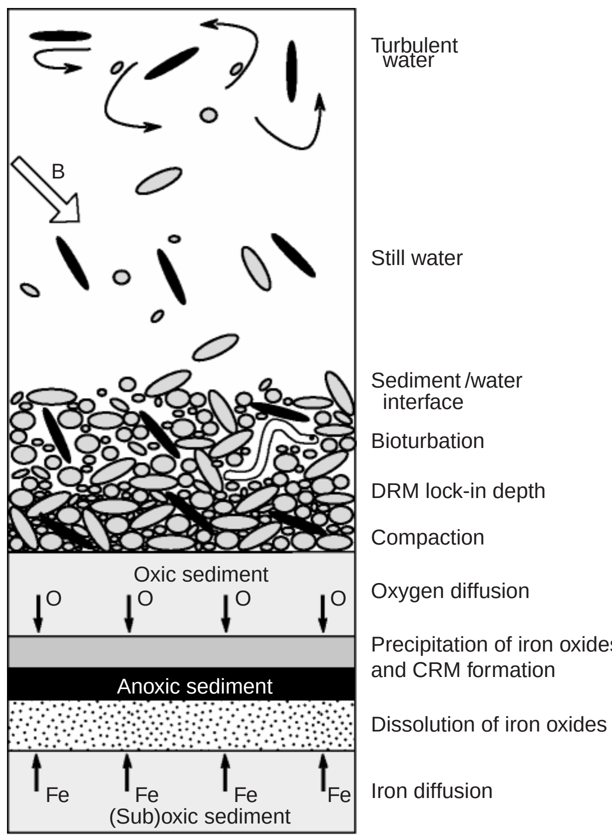

Figure 2 Depositional remanent magnetization (DRM) acquired in sediments involves a continuum of physical and chemcial processes. Detrital magnetic minerals (black) will be statistically aligned along the ambient geomagnetic field (**B**) in still water and in the soft and bioturbated water-saturated sediment just below the sediment–water interface. Upon compaction and dewatering, the grains are mechanically locked, preserving the direction of the field. Early diagenetic processes such as sulfate reduction may dissolve iron-bearing minerals. Upon encountering a more oxic environment, iron may precipitate as iron oxides, which will acquire a chemical remanent magnetisation (CRM). The thus-acquired CRM in this layer may have a much later age than the depositional age of this layer.

The Magnetic Signal in Marine Sediments

In sediments, it is often assumed that the natural remanent magnetization is due to DRM. In the past, there has been some debate on the mechanisms of DRM acquisition, and the term pDRM or postdepositional DRM has often been used, because the lock-in zone has a certain depth. Sediment at this depth is slightly older than that the sediment–water interface. Hence, the acquisition of NRM is always slightly delayed with respect to sediment age. For practical purposes (in magnetostratigraphy) this usually has no serious consequences. Nevertheless, the concept of a purely detrital remanent magnetization is an oversimplification of the real world. We therefore prefer to use the term depositional remanent magnetization (DRM) to refer to a continuum of physical and chemical processes that occur during and shortly after deposition. The acquisition of a DRM thus depends both on detrital input of magnetic minerals and the new formation of such minerals in the sediment through early diagenetic processes (**Figure 2**). Early diagenesis is widespread and occurs in virtually every sedimentary environment. Depending on the role of organic matter, existing magnetic minerals may dissolve and iron may be mobilized and precipitate elsewhere. These geochemical processes may results in the (partial) removal of the original paleomagnetic signal and in the acquisition of CRM in a particular zone (much) later than the deposition of the sediment in this zone. Clearly, such processes may severely damage the fidelity of the paleomagnetic record, and may offset the position of a reversal boundary by a distance in sediment that can correspond to a time of up to tens of thousands of years. However, in 'suitable' sediments the damage is usually restricted, and the paleomagnetic signal of such sediments may be considered as reliable and near-depositional. Suitable environments are generally those with a sufficiently high sedimentation rate, a significant detrital input, and a predominantly oxic environment. Therefore, it is often necessary to check the origin of the NRM using rock magnetic and geochemical methods.

As a rule, the total NRM is composed of different components. Ideally, the primary NRM — that originats from near the time of deposition — has been conserved, but often this original signal is contaminated with 'viscous' remanence components, referred to as VRM. Such a VRM may result from partial realignment of 'soft' magnetic domains in the present-day field or from low-temperature oxidation of magnetic minerals. It is generally easily removed through magnetic 'cleaning,' which consists of a routine laboratory treatment called demagnetization; details can be found in any standard textbook on paleomagnetism. Despite these pitfalls in sedimentary paleomagnetic records, sediments — in contrast to igneous rocks — have in principle the distinctive advantage of providing a continuous record of the geomagnetic field, including the history of geomagnetic polarity reversals. Paleomagnetic studies of the sediments will then reveal the pattern of geomagnetic reversals during deposition.

The Geomagnetic Polarity Timescale (GPTS)

Surveys over the ocean basins carried out from the 1950s onward found linear magnetic anomalies, parallel to mid-ocean ridges, using magnetometers towed behind research vessels. During the early 1960s, it was suggested, and soon confirmed, that these anomalies resulted from the remanent magnetization of the oceanic crust. This remanence is acquired during the process of seafloor spreading, when uprising magma beneath the axis of the midocean ridges cools through its Curie temperature (**Figure 3**) in the ambient geomagnetic field, thus acquiring its direction and polarity. The continuous process of rising and cooling of magma at the ridge results in magnetized crust of alternating normal and reversed polarity that produces a slight increase or decrease of the measured field — the marine magnetic anomalies. 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 wonderfully continuous 'tape recording' of the geomagnetic reversal sequence.

A major step in constructing a time series of polarity reversals was taken in 1968 by Heirtzler and co-workers. They used a long profile from the southern Atlantic Ocean and, extrapolating from a known age for the lower boundary of the Gauss Chron (**Figure 4**), they constructed a geomagnetic polarity timescale under the assumption of constant spreading. Subsequent revisions mostly used the anomaly profile of Heirtzler, often adding more detail from other ocean basins, and appending additional calibration points. These calibration points are derived from sections on land, which, first, must contain a clear fingerprint of magnetic reversals that can be correlated to the anomaly profile, and second, contain rocks that can be reliably dated by means of radiometric methods. The GPTS is then derived by linear interpolation of the anomaly pattern between these calibration points, again under the assumption of constant spreading rate between those points.

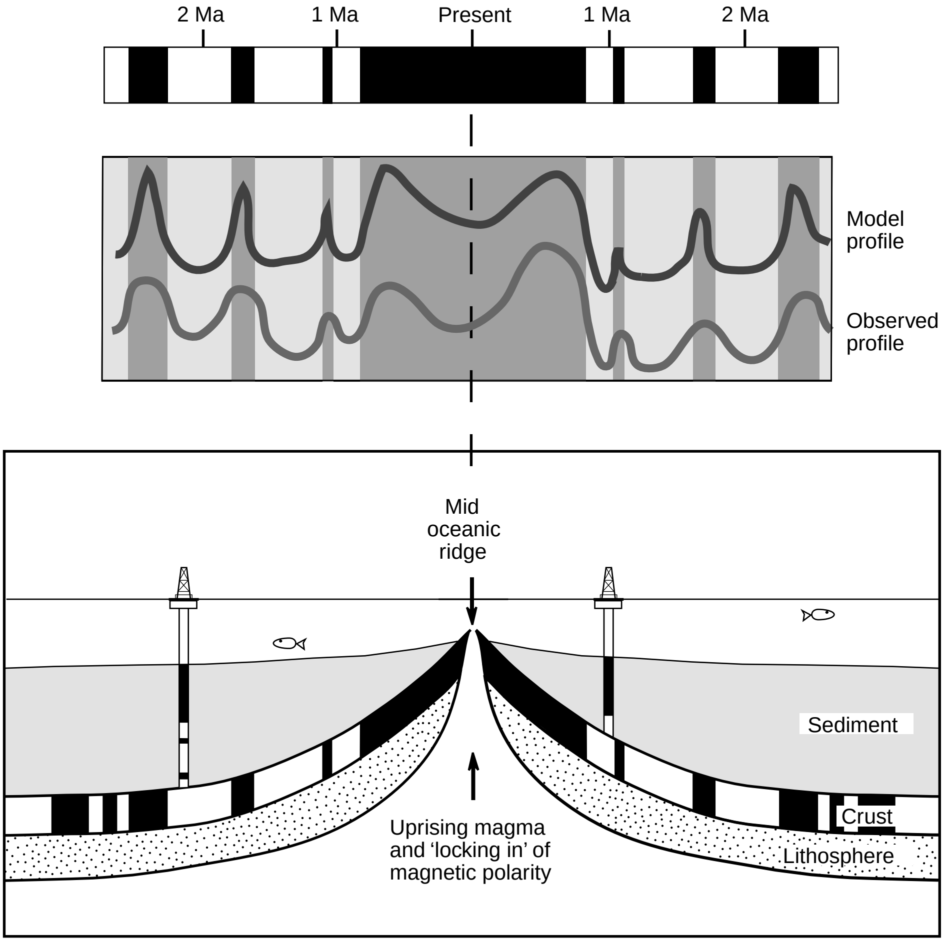

Figure 3 Formation of marine magnetic anomalies during seafloor spreading. The oceanic crust is formed at the ridge crest, and while spreading away from the ridge it is covered by an increasing thickness of oceanic sediments. The black (white) blocks of oceanic crust represent the original normal (reversed) polarity thermoremanent magnetization (TRM) acquired upon cooling at the ridge. The black and white blocks in the drill holes represent normal and reversed polarity DRM acquired during deposition of the marine sediments. The model profile (grey) represents computed magnetic anomalies produced by the block model of TRM polarity (top); the observed profile (dark) is the observed sea-level magnetic anomaly profile due to the magnetized oceanic crust.

The development of the GPTS (**Figure 4**) shows increasing detail and gradually improved age control. Periods of a predominant (normal or reversed) polarity are called chrons, and the four youngest ones are named after individuals: Brunhes (normal), who suggested field reversal; Matuyama (reversed), who proved this; Gauss (normal), who mathematically described the field; and Gilbert (reversed), who discovered that the Earth itself is a huge magnet. Chrons may contain short intervals of opposite polarity called subchrons, which are named after the locality where they were discovered; for example, the normal Olduvai subchron within the Matuyama reversed chron is named after Olduvai Gorge in Tanzania, and the Kaena reversed subchron within the Gauss normal chron after Kaena Point on Hawaii. Older chrons were not named but numbered, according to the anomaly numbers earlier given by Heirtzler, which has led to a confusing nomenclature of chrons and subchrons.

A major step forward was taken by Cande and Kent in 1992, who thoroughly revised the magnetic anomaly template over the last 110 My. They constructed a synthetic flow line in the South Atlantic, using a set of chosen anomalies that were taken as tie points. The intervals between these tie points were designated as category I intervals. On these intervals they projected (stacks of) the best-quality profiles surveyed in this ocean basin, providing category II intervals. Since the spreading of the Atlantic is 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); see, for instance, the increase of detail around 7 Ma in **Figure 4**. Very short

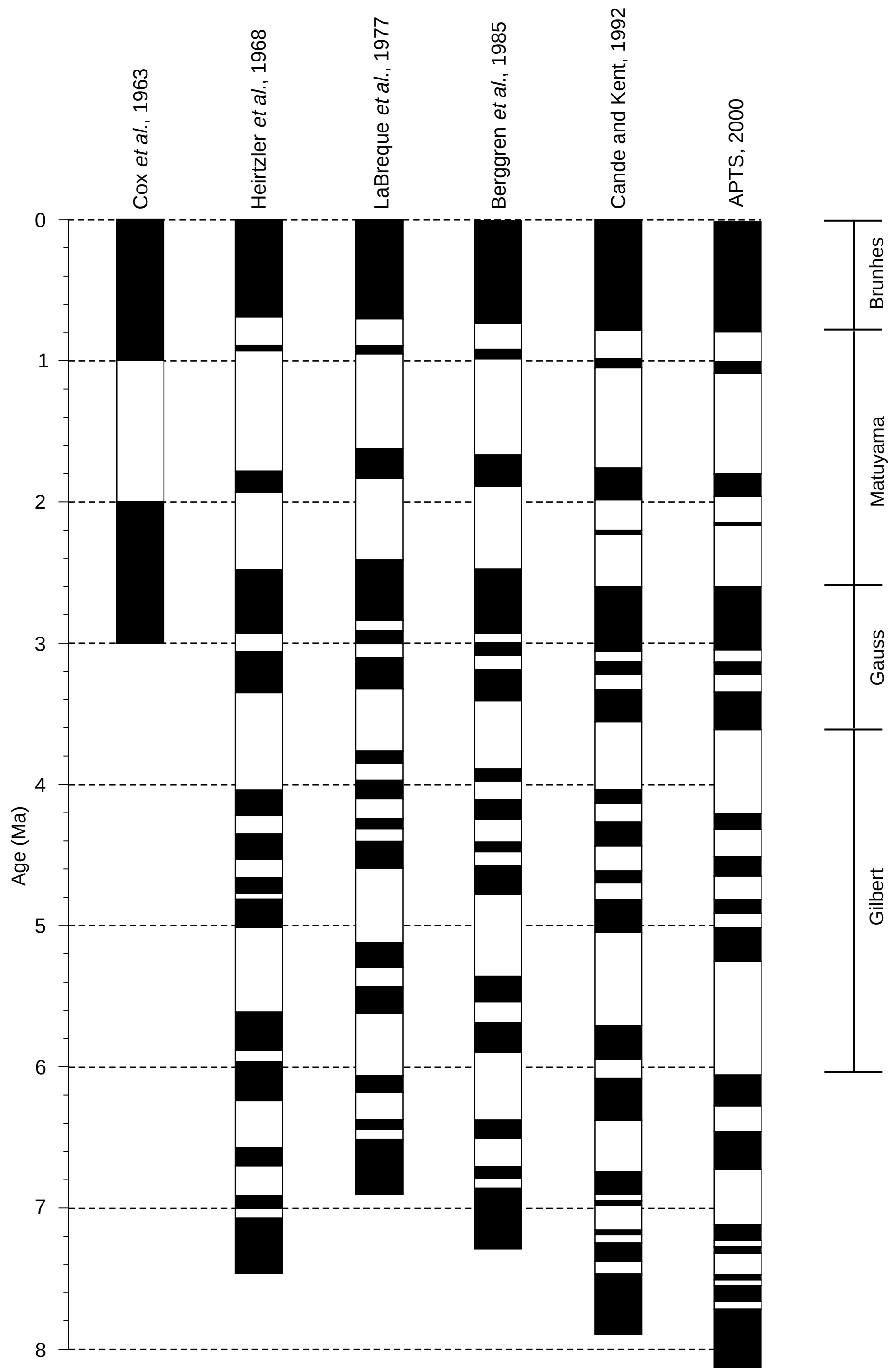

Figure 4 Development of the geomagnetic polarity timescale (GPTS) through time. The initial assumption of periodic behavior (in 1963) was soon abandoned as new data became available. The first modern GPTS based on marine magnetic anomaly patterns was established in 1968 by Heirtzler and co-workers. Subsequent revisions show improved age control and increased resolution. A major breakthrough came with the astronomical polarity timescale (APTS), in which each individual reversal is accurately dated.

and low-intensity anomalies still have an uncertain origin. They 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 may just represent intensity fluctuations of the geomagnetic field causing the oceanic crust to be less (or more) strongly magnetized. Because of their uncertain or unverified nature, these were called cryptochrons. In addition, Cande and Kent developed a consistent (sub)chron nomenclature that is now used as the standard. In total, they used nine calibration points, but they made a break with tradition by using, for the first time, an astronomically dated age tie point for the youngest one: the Gauss/Matuyama boundary. The correlation of the GPTS to global biostratigraphic zonations is covered extensively by Berggren *et al.* (1995).

The template of magnetic anomaly patterns from the ocean floor has remained central for constructing the GPTS from the late Cretaceous onward (110–0 Ma). Only recently, the younger part of the GPTS has been based on direct dating of each individual reversal through the use of orbitally tuned timescales. In their most recent version of the GPTS, Cande and Kent included the astronomical ages for all reversal boundaries for the past 5.3 My. The CK95 or Cande and Kent (1995) geomagnetic polarity timescale is at present the most widely used standard. Polarity timescales for the Mesozoic rely on some of the oldest magnetic anomaly profiles, down to the late Jurassic, and on dated magnetostratigraphies of sections on land.

The Astronomical Polarity Timescale (APTS)

The latest development in constructing a GPTS comes from orbital tuning of the sediment record; for details see the article Orbitally Tuned Timescales. It differs essentially from the conventional GPTS in the

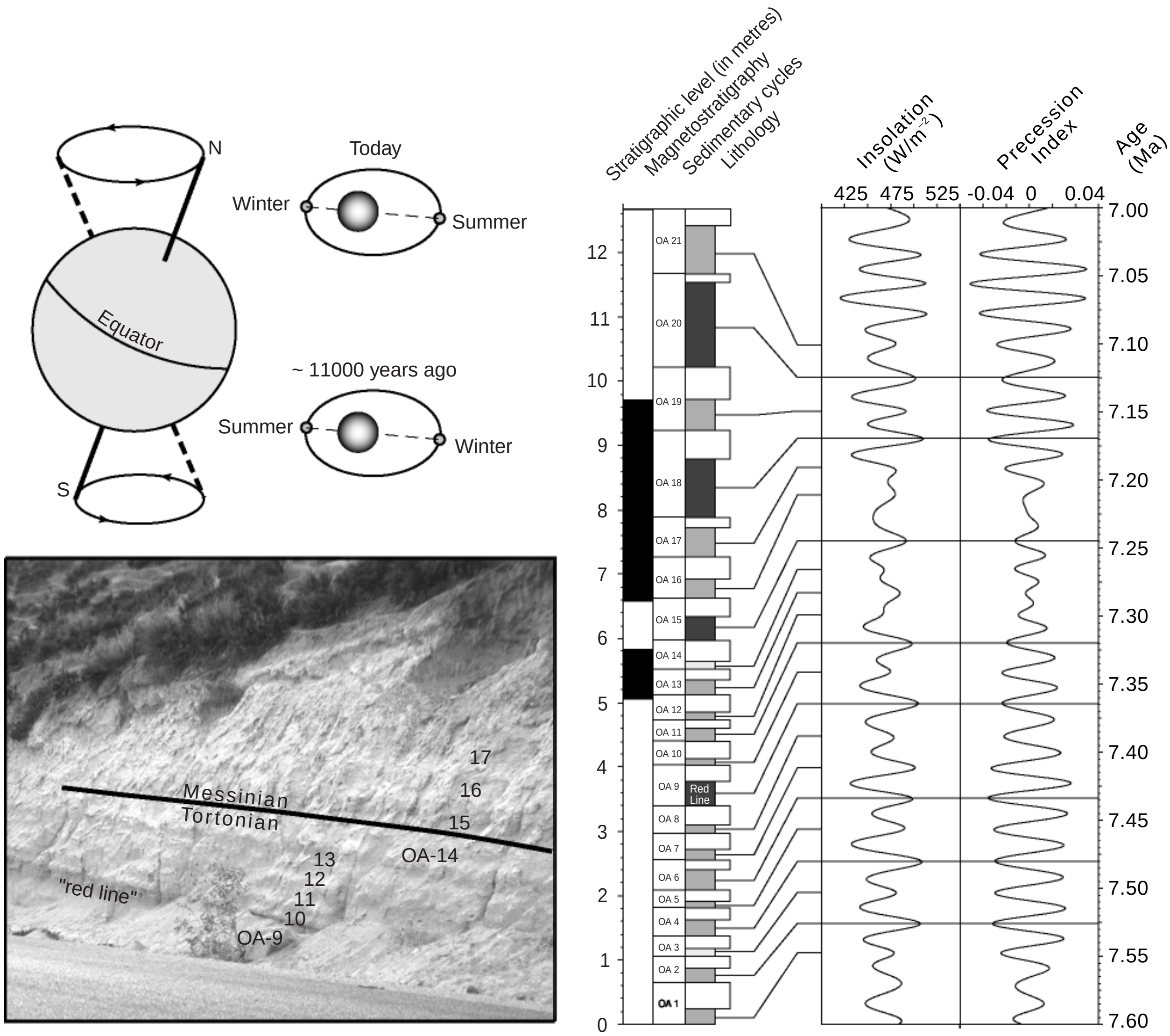

Figure 5 Magnetostratigraphy, cyclostratigraphy, and astrochronology of the marine Oued Akrech section from the Atlantic margin of Morocco, which defines the Tortonian–Messinian Global Stratotype Section and Point (GSSP). The sedimentary cycles represent a cyclically changing environment that correlates with variations in insolation. Insolation is strongly related to climatic precession (upper left panel), which induces cyclic changes in seasonal contrast, reflected one-on-one in the sedimentary cycles. (After Hilgen FJ *et al.* (2000) *Episodes* 23(3): 172–178.)

sense that each reversal boundary — or any other geological boundary for that matter, e.g. biostratigraphic datum levels or stage and epoch boundaries — is dated individually. This has provided a breakthrough in dating of the geological record and has the inherent promise of increasing understanding of the climate system, since cyclostratigraphy and subsequent orbital tuning rely on deciphering and understanding environmental changes driven by climate change, which in turn is orbitally forced.

The fact that the age of each reversal is directly determined, rather than interpolated between calibration points, has important consequences for (changes in) spreading rates of plate pairs. Rather than having to assume constant spreading rates between calibration points, one can now accurately determine these rates, and small changes therein. Indeed, Wilson found that the use of astronomical ages resulted in very small and physically realistic spreading rate variations. As a result, the discrepancy between plate motion rates from the global plate tectonic model (NUVEL-1) and those derived from geodesy has become much smaller. Meanwhile NUVEL-1 has been updated (to NUVEL-1A) to incorporate the new astronomical ages.

Another application is the dating of Pleistocene, Pliocene, and Miocene, and older stage boundaries, many of which have been defined in the Mediterranean. The availability of a good astrochronology has effectively become a condition for the definition of a Global Boundary Stratotype Section and Point (GSSP). An example is shown in **Figure 5**, showing the Tortonian–Messinian GSSP that has recently been defined in the Atlantic margin basin of western Morocco.

Perhaps one of the most promising areas of the application of astrochronology is in the bed-to-bed correlation of the two different realms of oceans and continent. Climate forcing may be expected to have a different expression in the different realms because of the different nature of their sedimentary environments. A recently established and refined orbital timescale for the loess sequences of northern China relies upon the correlation of detailed monsoon records to the astronomical solutions and the oceanic oxygen isotope records. An important finding was that the straightforward use of magnetostratigraphy and correlation to the GPTS cannot provide a sufficiently accurate age model for comparison with the ocean record, since the analysis of the astrochronological framework demonstrates considerable downward displacement of reversal boundaries because of delayed lock-in of the NRM. With the new chronology and its direct correlation to the oceanic record, it is now possible to analyse terrestrial paleomonsoon behavior for the past 2.6 My and compare it to climate proxies from the marine realm. This may give important information, for instance, on leads and lags of various systems in response to climate change, on phase relations with insolation, or on the relation between global ice volume and monsoonal climate.

See also

Aeolian Inputs. Magnetics. Monsoons, History of. Orbitally Tuned Timescales. Paleoceanography, Climate Models in.

Further Reading

Berggren WA, Kent DV, Aubry MP and Hardenbol J (1995) *Geochronology, Time Scales and Global Stratigraphic Correlations: A Unified Temporal Framework for an Historical Geology*. Society of Economic Paleontologists and Mineralogist, Special Volume 54.

Butler RF (1992) *Paleomagnetism, Magnetic Domains to Geologic Terranes*. Boston: Blackwell Scientific.

Cande SC and Kent DV (1992) A new Geomagnetic Polarity Time Scale for the Late Cretaceous and Cenozoic. *Journal of Geophysical Research* 97: 13917–13951.

Cande SC and Kent DV (1995) Revised calibration of the Geomagnetic Polarity Time Scale for the Late Cretaceous and Cenozoic. *Journal of Geophysical Research* 100: 6093–6095.

Gradstein FM, Agterberg FP, Ogg JG, *et al*. (1995) A Triassic, Jurassic and Cretaceous time scale. In: Berggren WA, Kent DV, Aubry M and Hardenbol J (eds) *Geochronology, Time Scales and Stratigraphic Correlation*. Society of Economic Paleontologists and Mineralogist, Special Volume 54: 95–128.

Hilgen FJ, Krijgsman W, Langereis CG and Lourens LJ (1997) Breakthrough made in dating of the geological record. *EOS, Transactions of the AGU* 78: 285, 288.

Heirtzler JR, Dickson GO, Herron EM, Pitman WC III and Le Pichon X (1968) Marine magnetic anomalies, geomagnetic field reversals, and motions of the ocean floor and continents. *Journal of Geophysical Research* 73: 2119–2136.

Heslop D, Langereis CG and Dekkers MJ (2000) A new astronomical time scale for the loess deposits of Northern China. *Earth and Planetary Science Letters* 184: 125–139.

Maher BA and Thompson R (1999) *Quaternary Climates, Environments and Magnetism*. Cambridge: Cambridge University Press.

Opdyke ND and Channell JET (1996) *Magnetic Stratigraphy*. San Diego: Academic Press.

Shackleton NJ, Crowhurst SJ, Weedon GP and Laskar J (1999) Astronomical calibration of Oligocene–Miocene time. *Philosophical Transactions of the Royal Society of London*, *A* 357: 1907–1929.

Wilson DS (1993) Confirmation of the astronomical calibration of the magnetic polarity time scale from seafloor spreading rates. *Nature* 364: 788–790.