PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

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Phil. Trans. R. Soc. Lond. A 1999 **357**, 1995-2007
doi: 10.1098/rsta.1999.0411

THE ROYAL SOCIETY

Orbital tuning of geomagnetic polarity time-scales

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Milankovitch climate cyclicity and magnetic polarity stratigraphy are being successfully combined as a powerful geochronometer in the astronomical polarity time-scale (APTS). The APTS for 0–5.23 Ma has been rapidly accepted as the definitive chronology for the Pliocene and Pleistocene against which even high precision radiometric dating is now calibrated. Extensions of astronomical calibration to the late Miocene (5.23 Ma to *ca.* 10 Ma) in Mediterranean and Pacific marine sections agree to within about a 100 ka eccentricity, which amounts to an uncertainty of only *ca.* 1% of the age. Orbital eccentricity periods are thought to remain stable over very long times and thus provide the possibility of precise relative age control in the pre-Neogene. An APTS, for example, has been developed in a thick lacustrine section of Late Triassic age (*ca.* 202–233 Ma) on the basis of the 404 ka orbital eccentricity cycle modulating the expression of the precession climate cyclicity. Finally, there has been renewed speculation about an obliquity-modulated precessional geodynamo based on periodicities in relative palaeointensity data from Ocean Drilling Program sediment cores.

Keywords: polarity; magnetostratigraphy; Milankovitch; time-scales; Neogene; Triassic age

1. Introduction

When averaged over a few thousand years, the Earth's magnetic field is closely approximated by a geocentric axial dipole field (figure 1). The resulting systematic change in field parameters as a function of latitude makes palaeomagnetism a powerful navigation tool for the geologic past. The main magnetic field is due to dynamo action in the outer core. The predominant axial dipole character of the field can be generally ascribed to Coriolis force shaping the motions of the highly conducting fluid where thermal convection flow patterns contribute to spatial irregularities and the chaotic secular variation including occasional flips in polarity. The sequence of polarity reversals in fact appears to have a distinctive temporal sequence where each reversal has hardly any memory of a previous occurrence (Cox 1969; McFadden & Merrill 1993). The chronology of reversals in a geomagnetic polarity time-scale (GPTS) must therefore be determined empirically.

The Earth's climate system also has a basic zonal character, in this case related to the way insolation is distributed as the planet rotates, precesses, tilts and orbits around the Sun. There is an overall equator to pole(s) variation in Earth's climate and indeed, the latitudinal distribution of palaeoclimate indicators is a classic means

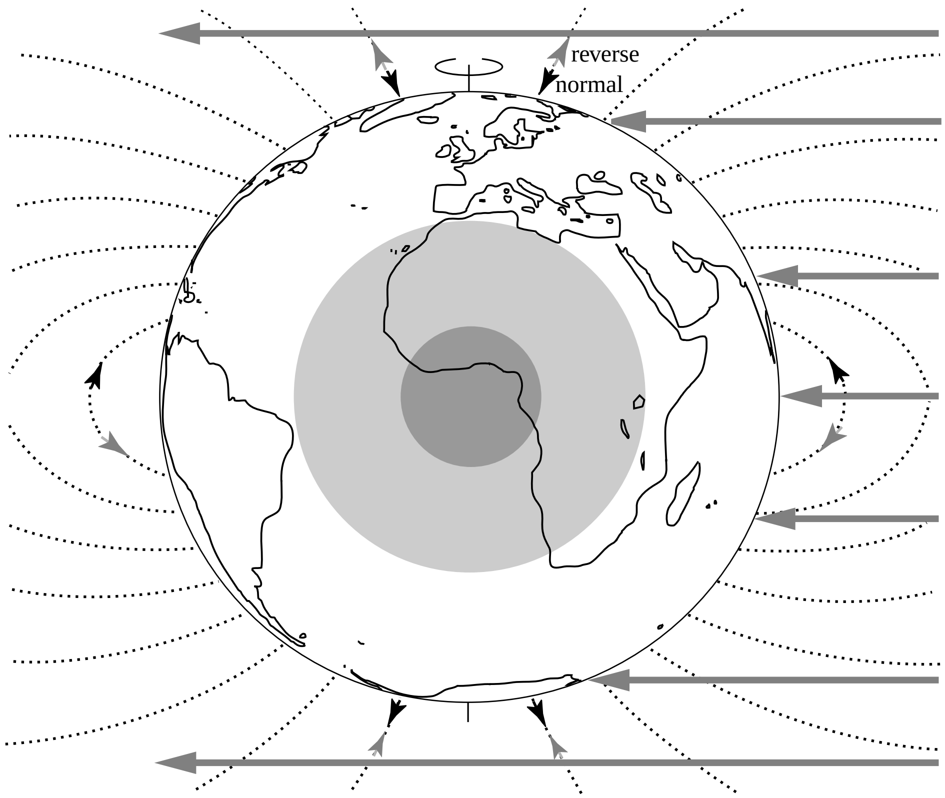

Figure 1. Field of the time-averaged geomagnetic axial dipole (dotted curves) and annualized mean path of insolation (heavy straight lines with arrows), illustrating systematic dependencies with respect to the rotation axis (i.e. a function of latitude). Dynamo action in the fluid outer core (projected as a lightly shaded circle with the inner core as a darker shaded superposed circle) accounts for stochastic properties of geomagnetic polarity reversals, whereas long-term temporal variations in climate tend to be periodic due to regular orbital motions.

by which the geocentric axial dipole hypothesis can be tested for the remote geologic past (Briden 1968). However, a fundamental difference from geomagnetism is that climate has a predictable long-term component of change resulting from Earth's clockwork orbital motions. The juxtaposition of these two global phenomena, geomagnetism, which has a stochastic polarity reversal record with its distinctive, non-repetitive signature that is ideal for correlation, and climate, with a rich periodic spectrum of temporal change that is ideal for parsing time, results in an ideal universal geochronometer whose ultimate expression is the astronomical polarity time-scale (APTS) (figure 2).

This paper comments on (1) the development of the APTS in the late Neogene and its extension to the late Miocene, with comparisons to the current GPTS based mostly on interpolation of marine magnetic anomalies; (2) prospects for the use of astronomical calibration of polarity sequences in the pre-Cenozoic, using the Late Triassic as an example; and (3) speculations of a fundamental relationship between Milankovitch cyclicity and geomagnetic field.

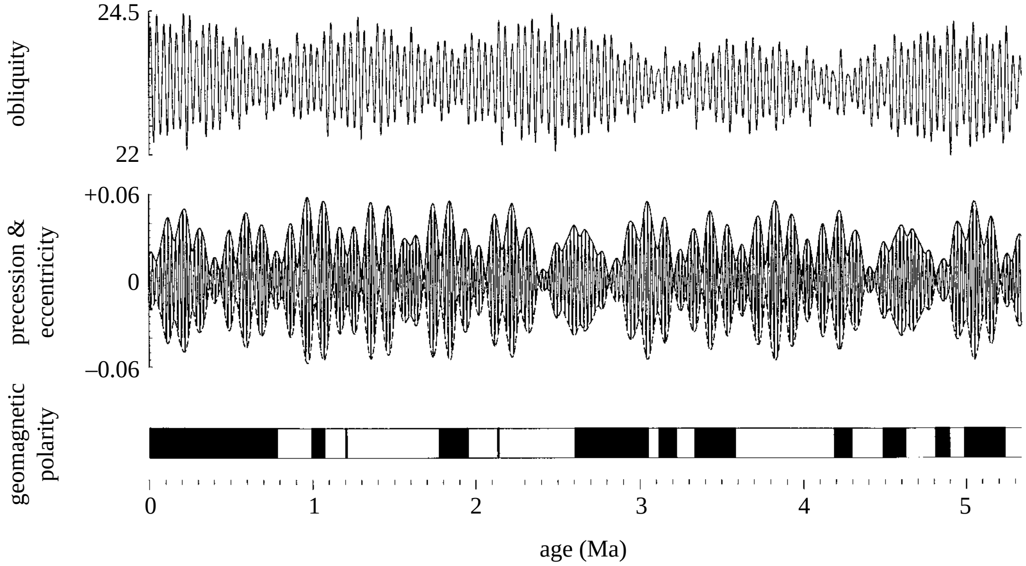

Figure 2. Comparison of the stochastic sequence of geomagnetic polarity reversals (APTS: Shackleton *et al.* 1990; Hilgen 1991*a*, *b*) and periodic variations of orbital eccentricity modulating precession and of obliquity (Laskar (1990) as plotted with AnalySeries; Paillard *et al.* 1996) for 0–5.34 Ma. Distinctive temporal sequence of polarity reversals provides a bar code that is ideal for correlation in marine and non-marine sediments as well as igneous rocks including the sea floor, whereas periodic character of Milankovitch climate change recorded in sediments serves as a precise timing tool.

2. Brief history of APTS (0–5.23 Ma)

Emiliani (1966) published the first continuous climate record for the Pleistocene, defining oxygen-isotope stages 1–17 from an analysis of Caribbean deep sea sediment cores. The chronology of the climate fluctuations was based on extrapolation of sedimentation rates from radiocarbon and protactinium/thorium measurements near the top of the cores. However, Emiliani's chronology was contested by Broecker & Ku (1969), who suggested that it should be stretched by 25% on the basis of a different analysis of the isotopic dates. For example, the stage 13–14 boundary was 313 ka according to Emiliani (1966), but 392 ka according to Broecker & Ku (1969); correspondingly, the age of *ca.* 425 ka for stage 17 at the base of Emiliani's climate record would be increased by 25% to 532 ka (figure 3).

The oxygen-isotope time-scale problem was resolved by the magnetic polarity stratigraphy of a 1600 cm long core (V28-238) recovered from the Ontong–Java Plateau (Shackleton & Opdyke 1973). The pelagic calcareous sediments had a superb oxygen-isotope record that could be correlated to and extended beyond Emiliani's record. The Brunhes–Matuyama (B–M) boundary was located in stage 19 at 1200 cm depth and corresponded to 0.70 Ma, the accepted age of the B–M boundary at the time. The chronology of Late Pleistocene oxygen-isotope stages could thus be determined by interpolation and was found to be considerably older than previous time-scales. For example, the stage 17–18 boundary had an interpolated age of 647 ka, more than 50% greater than the age extrapolated for stage 17 by Emiliani (1966) and even *ca.* 20% greater than in the revised time-scale of Broecker & Ku (1969).

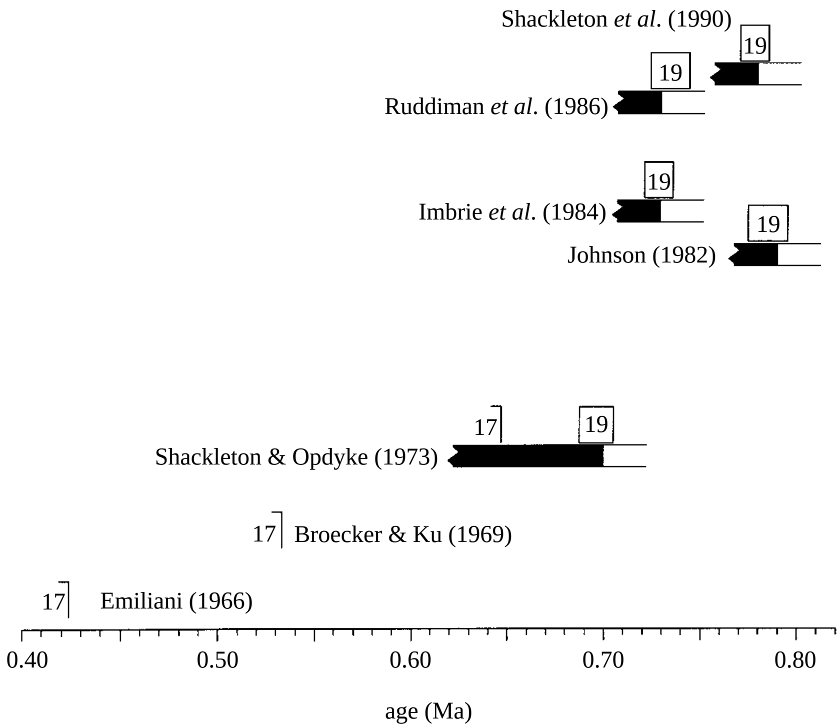

Figure 3. Comparison of oxygen-isotope chronologies for the Late Pleistocene. Age estimates for oxygen-isotope stage 17 by Emiliani (1966) and Broecker & Ku (1969) are based on extrapolation of sedimentation rates. Shackleton & Opdyke (1973) extended the oxygen-isotope record and correlated stage 19 with the B–M boundary; their interpolated age of stage 17 is shown for comparison with the earlier studies. Subsequent studies either regarded the radiometrically dated age of the B–M boundary to be consistent with orbital tuning (e.g. Imbrie *et al.* 1984; Ruddiman *et al.* 1986) or calculated a new age for the B–M boundary based on orbital tuning (Johnson 1982; Shackleton *et al.* 1990), which is now regarded as providing the best age estimates.

The combined oxygen-isotope and magnetostratigraphic record for V28-238 (and V28-239; Shackleton & Opdyke 1976) showed the importance of the orbital eccentricity cycle in the Late Pleistocene and provided the long-term temporal framework for more detailed studies of Milankovitch climate cyclicity.

The landmark pacemaker study (Hays *et al.* 1976) presented compelling evidence from spectral and time-series analysis for orbital forcing of Late Pleistocene glacial–interglacial cycles, even though the composite record from two deep-sea cores that were studied was only *ca.* 400 ka long. This meant that it was not possible to tune climate cycles for the entire Brunhes, and hence to make an independent astronomical age estimate for the B–M boundary, which was regarded from radiometric dating to be *ca.* 0.72 Ma. The last half of the Brunhes became the focus of the SPECMAP project for at least the next decade (e.g. Martinson *et al.* 1987).

Johnson (1982) derived an astronomical age of 0.79 Ma for the B–M boundary from the best available oxygen-isotope records for the Brunhes, which basically were still from cores V28-238 and V28-239 (Shackleton & Opdyke 1973, 1976). This age estimate was 60 ka older than 0.73 Ma for the B–M boundary as was recently determined in a comprehensive assessment of radiometric age data (Mankinen & Dalrymple 1979). The glaring magnitude of the discrepancy was probably why the prescient and basically correct interpretation by Johnson was virtually ignored for the rest of the decade.

The breakthrough in acceptance of an older astronomical age for the B–M boundary came in 1990 with the publication of a high-resolution composite oxygen-isotope record for the Pleistocene from multiple cores at equatorial Pacific Site 677 (Shackleton *et al*. 1990). Ironically, there was no magnetostratigraphic record for this site, but the positions of the major polarity reversals could be approximated by biostratigraphic datums and the distinctive character of the oxygen-isotope record. For example, the B–M boundary was assumed to be coeval with oxygen-isotope stage 19, as originally observed in V28-238 and found in virtually all subsequent oxygen-isotope records with magnetic polarity stratigraphy (see, for example, DeMenocal *et al*. 1990). Shackleton *et al*. (1990) determined an age of 0.78 Ma for the B–M boundary level using eccentricity-modulated precession cycles to calibrate the climate record in Ocean Drilling Program (ODP) Site 677. In some other Brunhes high-resolution records, such as from Deep Sea Drilling Project (DSDP) Leg 94, the climate proxies were tuned to orbital obliquity; however, an obliquity (*ca*. 41 ka) cycle was evidently missed in the Brunhes in these studies and an age of 0.73 Ma for the B–M boundary had been retained (Ruddiman *et al*. 1986).

Hilgen (1991*a*) confirmed the astronomical ages of the Olduvai subchron and the Gauss–Matuyama boundary obtained by Shackleton *et al*. (1990) by associating clusters of sapropels and individual carbonate cycles and sapropels in Mediterranean sections with eccentricity and precession variations. Hilgen (1991*b*) was then able to extend the astronomical dating of magnetic polarity reversals to the Miocene–Pliocene boundary by successively correlating the sedimentary patterns to the 400 ka and 100 ka eccentricity variations and finally to individual precession cycles. The combined chronologies of Shackleton *et al*. (1990) and Hilgen (1991*a*, *b*) can be regarded as the first astronomical polarity time-scale (APTS) which encompassed the Pliocene and Pleistocene, extending to the base of the Thvera polarity subchron with an astronomical age of 5.23 Ma.

For reasons that were not immediately obvious, the ages of polarity reversals in the APTS were systematically 5–7% older than in the widely used radiometrically dated GPTS of Mankinen & Dalrymple (1979). For example, the Gauss–Gilbert boundary was 3.58 Ma in the APTS compared with 3.40 Ma in the GPTS of Mankinen & Dalrymple (1979) as well as time-scales such as Berggren *et al*. (1985) that commonly used this radiometric age estimate as a tiepoint for extrapolation of the sea-floor spreading magnetic anomaly sequence. It soon became evident that uncertainty in the K–Ar database was at fault (Tauxe *et al*. 1992), rather than gross inaccuracies in the fundamental isotopic decay constants or the astronomical solutions. This was highlighted by new, high precision $^{40}Ar/^{39}Ar$ dating of the B–M boundary in lavas which yielded very nearly the same age (0.783 Ma; Baksi *et al*. 1992; see also Singer & Pringle 1996) as was determined astronomically in sedimentary records.

At about the same time, work was in progress on a complete revision of the GPTS for the Late Cretaceous and Cenozoic based on marine magnetic anomalies (Cande & Kent 1992). The sea-floor spreading record is ubiquitous and highly redundant, allowing a very complete record of reversals to be assembled. However, the marine magnetic reversal sequence requires external age calibration because the ocean floor

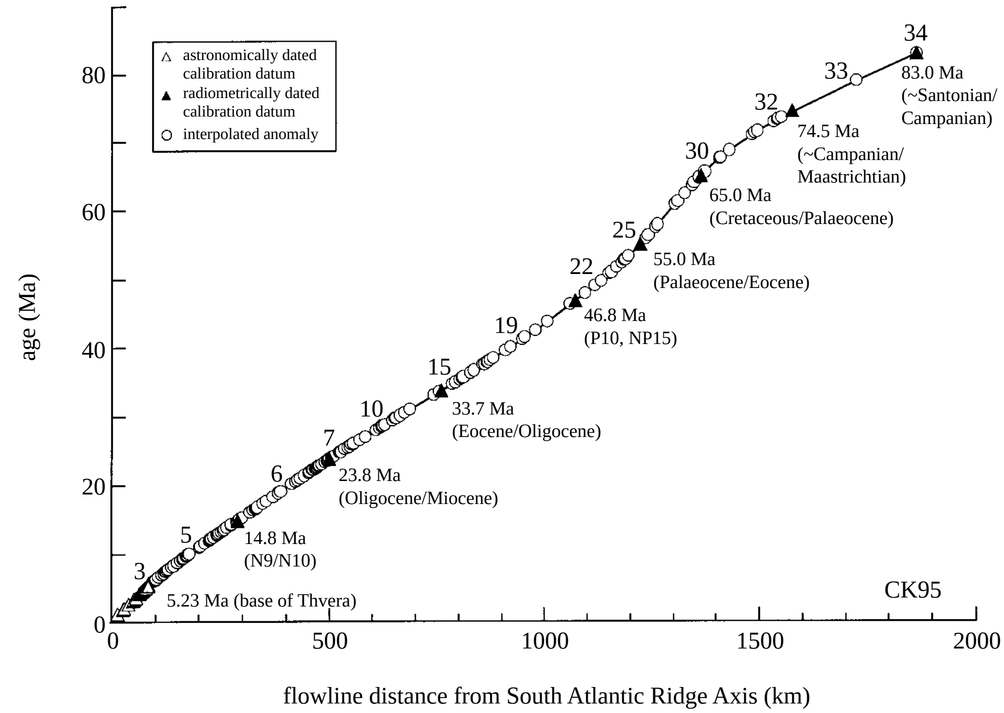

Figure 4. Age calibration of distances of magnetic anomalies projected along a synthetic flowline on the South Atlantic ridge system used in the geomagnetic polarity time-scale of Cande & Kent (1995; CK95). Ages for polarity chrons C1 to C3n (0–5.23 Ma, to base of Thvera subchron or base of chron C3n.4n) are taken directly from APTS of Shackleton *et al*. (1990) and Hilgen (1991*a*, *b*); ages of older polarity chrons determined by cubic spline interpolation through radiometrically dated tiepoints shown by solid triangles, plus the base of the Thvera from the APTS. See Cande & Kent (1992, 1995) and Berggren *et al*. (1995*b*) for more details.

source rocks are extremely difficult to date adequately. Cande & Kent (1992) adopted APTS ages for the B–M (0.78 Ma) and Gauss–Matuyama (2.60 Ma) boundaries as tiepoints in the youngest part of the anomaly–reversal sequence; the next tiepoint for cubic spline interpolation was for anomaly 5B (chron C5Bn) with an estimated radiometric age of 14.8 Ma. There was excellent agreement to the base of the Gauss (base of chron C2An) between the interpolated ages in CK92 and the direct APTS. However, the ages of the normal polarity subchrons in the Gilbert (collectively chron C3n) were systematically younger by *ca.* 150 ka in CK92 compared with the APTS (Berggren *et al*. 1995*a*).

The discrepancy between CK92 and the APTS was attributed to an artefact of the magnetic anomaly profiles used in CK92 to represent this interval (Wilson 1993). Moreover, the excellent agreement between the APTS and high precision $^{40}Ar/^{39}Ar$ dating (Renne *et al*. 1994; Clement *et al*. 1997) emphasized the reliability of the APTS. Accordingly, the APTS was incorporated *in toto* into the revised GPTS of Cande & Kent (1995), where the APTS age of 5.23 Ma (Hilgen 1991*b*) for the base of the Thvera subchron (older end of chron C3n.4n) was used to anchor the cubic spline interpolation for the older parts of the sequence (figure 4).

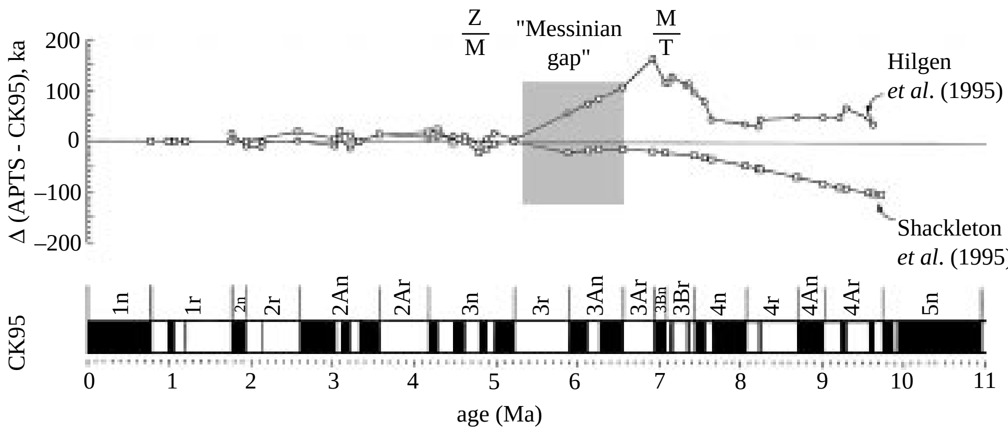

Figure 5. Geomagnetic polarity time-scale of Cande & Kent (1995: CK95) for 0–11 Ma, compared with age estimates from astronomical time-scales of Hilgen *et al.* (1995) and Shackleton *et al.* (1995). Data plotted as given in table 1 of Hilgen *et al.* (1995). CK95 already includes the APTS of Shackleton *et al.* (1990) and Hilgen (1991*a, b*) for 0–5.23 Ma (to base of chron C3n). For older than 5.23 Ma, CK95 is based on interpolation of magnetic anomalies (see figure 4). Stratigraphic positions of Messinian–Tortonian (M–T) and Zanclean–Messinian (Z–M) boundaries, and the 'Messinian gap' are from Hilgen *et al.* (1997).

3. Extension of APTS to the late Miocene (5.23 Ma to *ca.* 10 Ma)

To assess the consistency of pre-Pliocene time-scales, two recently published astronomical polarity time-scales for the late Miocene: one by Hilgen *et al.* (1995) based on sedimentary cycles in Mediterranean sections and the other by Shackleton *et al.* (1995) based on GRAPE records from ODP Leg 138 cores from the equatorial Pacific, can be compared with the nearly contemporaneously published GPTS of Cande & Kent (1995: CK95). Magnetic polarity stratigraphy was available in both astronomical time-scale studies (Krijgsman *et al.* 1995; Schneider 1995), allowing direct correlation to CK95. A comparison of differences in ages of polarity reversals in the two APTS with respect to CK95 is shown in figure 5 which along with the ensuing analysis is based on data tabulated by Hilgen *et al.* (1995).

The astronomical ages of Hilgen *et al.* (1995) are consistently older in the late Miocene than the astronomical ages of Shackleton *et al.* (1995) or the anomaly interpolated ages in CK95. Interestingly, the maximum difference is between these two astronomical time-scales (182 ka for the younger end of chron C3Bn) with the anomaly interpolated ages of CK95 generally falling in between. For 18 reversal boundaries in common between chrons C3An and C4Ar (*ca.* 5.9–9.6 Ma), the APTS of Hilgen *et al.* (1995) is on average 80 ka older, whereas the APTS of Shackleton *et al.* (1995) is on average 50 ka younger than CK95. The mean difference between the two published APTS over this late Miocene time-interval is 118 ka, or roughly one eccentricity cycle. This value can be considered a practical measure of uncertainty in an extended APTS when based on independent studies. Such differences in correlation to a target orbital curve nevertheless amount to only *ca.* 1% of the estimated age in the late Miocene and are largely a function of quality of data rather than simply age of sedimentary sections.

Astronomical age estimates for the Miocene are also now available from ODP Leg

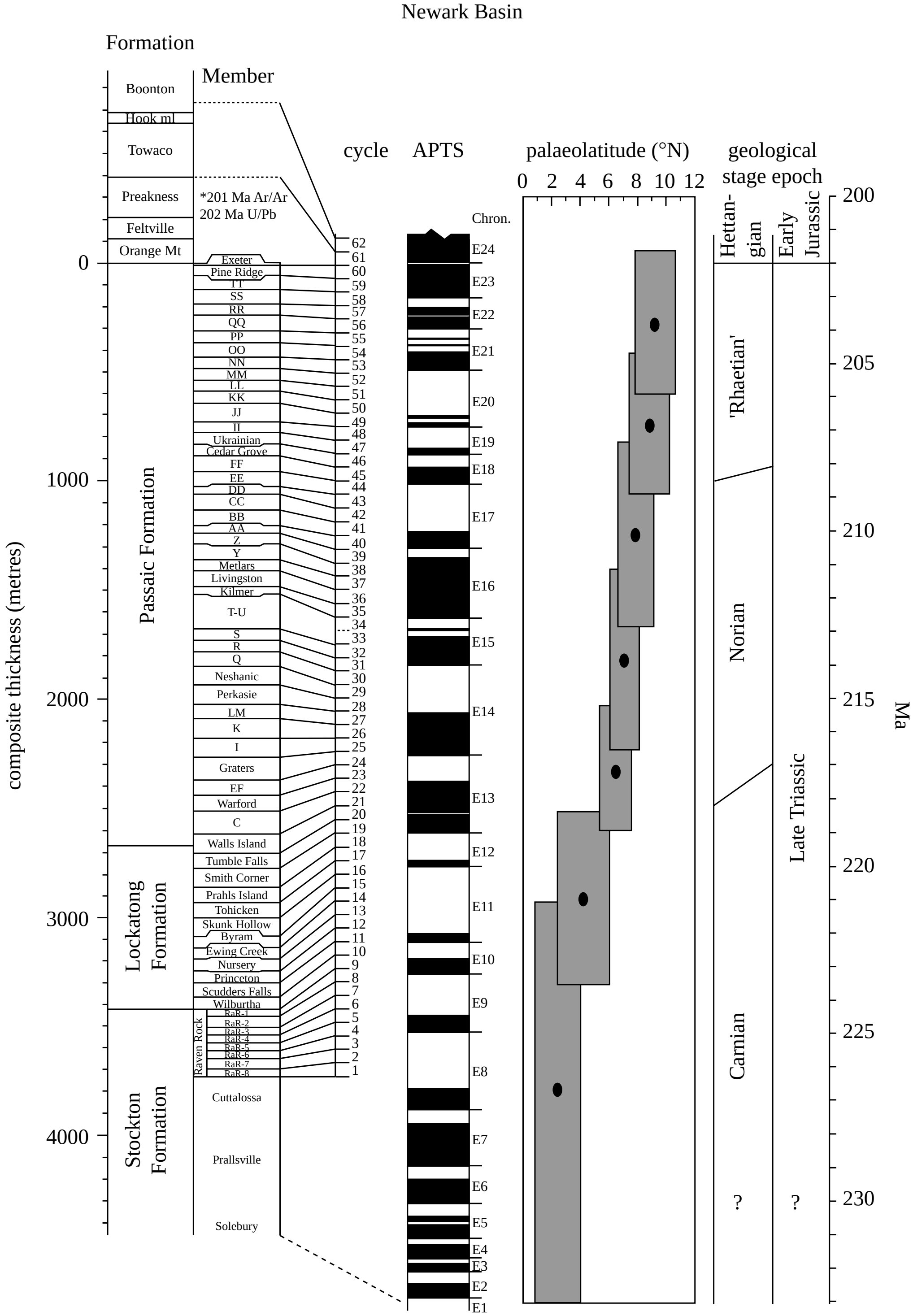

Figure 6. For description see opposite.

154 sediments (Shackleton & Crowhurst 1997), but the lack of magnetic polarity stratigraphy for the equatorial Atlantic ODP Leg 154 cores will complicate detailed comparisons with other time-scales. In the Mediterranean, a difficulty in extending the APTS is the 'Messinian gap' (Hilgen *et al*. 1997), an interval in the latest Miocene with poor data due to a concentration of evaporites and other sedimentary facies that are not conducive for detailed magnetic or cycle stratigraphies. The 'gap' separates cyclic Pliocene sediments for which an APTS has been developed (i.e. Hilgen 1991*b*) from underlying cyclic Miocene sections that are available in the Mediterranean region for further extension of the APTS. Although Messinian sections in Morocco are not punctuated by evaporites, it is difficult to link the available magnetic polarity and stable isotope stratigraphies to the Plio-Pleistocene APTS (Hodell *et al*. 1994). The 'Messinian gap' contributes to the large deviations in the latest Miocene in the APTS of Hilgen *et al*. (1995) compared with the chronologies from either Leg 138 sediments (Shackleton *et al*. 1995) or magnetic anomalies (Cande & Kent 1995). Precise biostratigraphic links to the Leg 154 astronomical record conceivably may provide a bridge over the 'Messinian gap'.

4. Astronomical dating in the pre-late Miocene (greater than 10 Ma)

Orbital solutions have been computed to 200 Ma but their reliability was judged to be reasonable only over 10–20 Ma due to chaotic elements; the reliability of solutions for precession and obliquity of the Earth are further constrained to time spans of only a few million years due to fundamental uncertainties in tidal terms (Laskar 1990; Laskar *et al*. 1993). In practical terms, Hilgen *et al*. (1995) suggested that astronomical dating errors may be *ca.* 5 ka at 5 Ma and may increase to 10–20 ka around 10 Ma. However, uncertainty in correlation of sedimentary cycles to orbital target curves is likely to be a more significant source of error than inaccuracies in the astronomical solutions themselves over this time range as suggested, for example, by the difference of about an eccentricity cycle between the APTS of Hilgen *et al*. (1995) and Shackleton *et al*. (1995) discussed above.

Errors in the astronomical solutions and perhaps correlations will presumably increase with increasing age, necessitating a modified strategy for developing an astrochronology in more remote geologic epochs. The greatest uncertainties in the astronomical solutions are for precession and obliquity which are directly affected by poorly constrained tidal dissipation components. On the other hand, changes in the frequency of the planetary system are thought to be much smaller so that eccentricity periods calculated for the past few million years are likely to have remained stable for hundreds of millions of years (Berger *et al*. 1992). It should thus be possible to use eccentricity modulation of precession cycles to develop precise albeit floating chronologies for earlier geologic epochs.

Figure 6. Late Triassic APTS based on cycle stratigraphy and magnetic polarity stratigraphy of composite section from drill cores in the Newark basin (Kent *et al*. 1995; Olsen & Kent 1996; Kent & Olsen 1999). Stratal thickness was converted to age by assuming each lithologic member is a 404 ka orbital eccentricity cycle and anchoring the interpolated relative chronology to an age of 202 Ma for the Triassic–Jurassic boundary in the early part of the Exeter Member (informal cycle no. 61). For APTS column, filled bars represent normal magnetic polarity and open bars are reverse polarity. The Newark Basin section was deposited at low palaeolatitudes based on palaeomagnetic inclination data from the drill cores which can account for the dominance of eccentricity-modulated precession and the general absence of an obliquity component in the climate record.

An example of a floating APTS comes from cycle stratigraphic and magnetostratigraphic analyses of a 5000+ m composite section of Late Triassic and earliest Jurassic continental sediments and interbedded volcanics obtained by scientific coring in the Newark rift basin of eastern North America (Olsen *et al*. 1996). This long, continuous and uniformly sampled record is characterized by 60 normal and reverse polarity magnetozones and a full spectrum of Milankovitch cyclicity as reflected by the lithologic facies response to climatically induced lake level variations (Kent *et al*. 1995; Olsen & Kent 1996). The prominent 404 ka eccentricity climate cycle was the basis for scaling the stratigraphic section in time (figure 6).

The resulting astronomically calibrated polarity time-scale spans from about 201 to 233 Ma when indexed to available radiometric dating, i.e. 202 Ma for the Triassic–Jurassic boundary. As described in detail elsewhere (Kent & Olsen 1999), the magnetic polarity interval lengths range from about 20 to 2000 ka and have a mean duration of 530 ka. There is no significant polarity bias. Moreover, the Late Triassic polarity reversals apparently occurred relatively rapidly (mean of *ca.* 8 ka for 42 transition records of 35 polarity reversals) and the intervals between reversals are closely approximated by an exponential distribution, consistent with a stochastic process. The overall statistical properties suggest that the behaviour of the geomagnetic field in the Late Triassic was not very different than in the Cenozoic, a conclusion only made possible by astronomical calibration of a virtually complete geomagnetic polarity record.

5. Obliquity-modulated precessional geodynamo?

Malkus (1968) suggested that Earth's precession could supply significant energy for the geodynamo due to fluid motions produced by the difference in dynamic ellipticity of the core and the mantle. Although subsequent work indicates that precession is unlikely to be the main source of energy to drive the geodynamo (e.g. Rochester *et al*. 1975), any contribution from precession might vary with the obliquity. This would make obliquity-modulated precession a testable forcing function for the geodynamo. In this regard, geomagnetic field intensity variation recorded in sediments over the Brunhes was suggested to have a prominent variation with a periodicity of around 41 ka that may be related to obliquity (Kent & Opdyke 1977). A difficulty is that such relative sedimentary palaeointensity records often have climatically induced lithologic variations at Milankovitch wavelengths, including obliquity, which may strongly bias estimates of geomagnetic intensity change (e.g. Kent 1982). Recently, Channell *et al*. (1998) reported spectral analyses of sedimentary records of relative palaeointensity from two ODP sites in the North Atlantic that show significant power at orbital eccentricity as well as obliquity frequencies. The eccentricity power was also evident in bulk magnetic properties and could thus be dismissed as due to climatically induced lithologic changes. The obliquity power, however, was not present in bulk magnetic properties and provides tantalizing evidence for orbital forcing of the geomagnetic field.

6. Conclusions

The demonstrated success of astronomical tuning in the late Neogene gives us reason to believe that this precise geochronological methodology will soon be extended to the formulation of the next generation of geomagnetic polarity time-scales for the entire Cenozoic and beyond. Problematical assumptions about the rate of sea-floor formation, whether it was smoothly varying in the South Atlantic (Cande & Kent 1992) or the Pacific (Wilson 1993) or at some average rate (Heustis & Acton 1997), will no longer be necessary. Instead, the APTS will allow us to answer critical questions about the timing and rate of changes in spreading rate on different ridges and reveal tectonic problems that are at present hidden by the assumptions we use.

The stability of the periods of the eccentricity cycles (Berger *et al*. 1992) suggests that the relative accuracy of astronomical tuning may not be very different over long geologic intervals. This was the premise for the development of an APTS for a 30 Ma interval of the Late Triassic (Kent & Olsen 1999; see also Olsen & Kent, this issue). Indications are that the combination of periodic Milankovitch cyclicity for tuning and stochastic polarity reversals for correlation will make the APTS a powerful geochronometer for the Cenozoic, Mesozoic and even older eras.

Finally, the development of astronomical time-scales for earlier periods should provide settings where magnetic records have a much smaller lithologic response than in the Pleistocene and even where almost all the lithologic signal is in the precession band. This will make it easier to find a clean geomagnetic intensity variation in the obliquity band to gauge the importance of a precessionally driven dynamo.

The author thanks N. J. Shackleton for the invitation to participate in the Discussion Meeting and for providing very useful comments on the manuscript. This paper represents research under various grants over several years from the US National Science Foundation, whose support is gratefully acknowledged. Lamont-Doherty Earth Observatory contribution no. 5907.

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