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A revised magnetic polarity timescale for the Cretaceous and Cainozoic†

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Magnetostratigraphic correlations of biostratigraphic stage boundaries have established calibration points for dating the polarity reversal sequence derived from marine magnetic anomalies. Interpolation between the best-estimate ages for these tie points gives a revised magnetic polarity timescale for the Cainozoic and Cretaceous. Recomputed sea-floor spreading rates for this time prove to be high during the Cretaceous quiet interval at several plate margins, but remained remarkably constant in the central Atlantic. The geomagnetic reversal frequency, when averaged over intervals of several megayears duration, has exhibited a steadily increasing trend since the late Cretaceous.

Introduction

The most continuous and detailed record of geomagnetic field polarity reversals is that interpreted from the magnetization of oceanic crust that formed by sea-floor spreading since the early Jurassic, when the continents forming Pangaea began to disperse. It is characterized by four distinct episodes: a mid-Jurassic quiet interval when no polarity reversals were occurring, a late Jurassic and early Cretaceous interval of reversals represented in the oceanic crust by the M-sequence anomalies, a mid-Cretaceous quiet interval, and the late Cretaceous and Cainozoic anomaly sequence which continues until the present. Much of the oceanic record of reversal history has now been confirmed independently in diverse palaeomagnetic investigations. In particular, magnetostratigraphic work has led to improved calibration of the magnetic polarity timescale for the Cretaceous and Palaeogene.

Magnetostratigraphic investigations of geomagnetic reversal history

Combined magnetic polarity and K–Ar age determinations on lavas resulted in a magnetic polarity timescale for about the last 5 Ma of reversal history (Cox 1969; Mankinen & Dalrymple 1979); magnetostratigraphic studies in dated lava sequences permit extension to about the last 10 Ma (McDougall *et al.* 1976). The youngest marine magnetic anomalies were dated by correlation to the last 3.35 Ma of this directly established polarity timescale and older anomalies of the Cainozoic and late Cretaceous sequence were then dated by linear extrapolation (Heirtzler *et al.* 1968). This polarity timescale, spanning about 80 Ma, has served as the basis for timing plate tectonic events, determining ocean-floor spreading rates and analysing the frequency of geomagnetic reversals in this time.

Magnetostratigraphic studies in non-indurated deep-sea sediment, sampled with a piston-type coring device that causes little disturbance of the sediment or its magnetization, revealed

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a magnetic reversal sequence equivalent to that found in continental lavas (Opdyke 1972). Global comparisons of biostratigraphic dating schemes were made and sedimentation rates in the deep ocean basins were evaluated. The record of independently confirmed polarity history was extended further by seeking out sampling sites where hiatuses at the sediment–water interface were known to exist. The polarity record for most of the Neogene has been pieced together from overlapping cores with the aid of palaeontological tie points (Opdyke 1972; Theyer & Hammond 1974).

Magnetic stratigraphy studies in marine sediment cores taken in the Deep Sea Drilling Project (D.S.D.P.) have often been plagued by disturbance due to the drilling process and by sporadic core recovery which resulted in important hiatuses in the record. Nevertheless, Keating & Helsley (1978) established the main features of Cretaceous polarity history, and Hailwood *et al.* (1979) numerically calibrated the late Neogene biostratigraphic zonation scheme for the North Atlantic and also pieced together a partial magnetic stratigraphy for the Palaeogene by using the magnetic stratigraphy in D.S.D.P. sediment cores. Ryan *et al.* (1974) indirectly correlated the magnetic polarity and biostratigraphy sequences for the entire Neogene from continental sections and D.S.D.P. and oceanic magnetic anomaly records. Recently D.S.D.P. introduced a hydraulic piston corer that gives practically undisturbed samples. The magnetic stratigraphy of Leg 73 holes clearly matches the marine magnetic reversal history of the Palaeogene and late Cretaceous (Tauxe *et al.* 1980).

Independent investigation of Cretaceous and Palaeogene magnetic reversals has been carried out in thick continental exposures of sedimentary rocks from the Mediterranean realm. The pelagic carbonate rocks of the southern Alps and northern Apennines in Italy were deposited on the southern margin of the former Tethyan ocean in a uniform, continuous fashion and have good palaeomagnetic properties. Magnetostratigraphic zonations in these formations have confirmed the magnetic polarity history deduced from oceanic magnetic anomalies. Allowing for fluctuations in carbonate sedimentation and sea-floor spreading rates, the correlation is almost perfect from the earliest Miocene anomaly 6C to middle Barremian anomaly M-4 (Roggenthen & Napoleone 1977; Lowrie & Alvarez 1977*a*, *b*; Channell *et al.* 1979; Lowrie *et al.* 1980*a*, 1982; Channell & Medizza 1981). Associated palaeontological zonations of planktonic foraminiferans and calcareous nannofossils located the biostratigraphic stage boundaries within the marine magnetic reversal sequence (figure 1). Especially important are the locations of the Palaeocene–Eocene boundary in the negative zone between anomalies 24 and 25 and the Tertiary–Cretaceous boundary in the negative zone between anomalies 29 and 30. Recently, Ogg (1981) has located the Tithonian–Berriasian (Jurassic–Cretaceous) boundary near the base of the positive polarity interval immediately younger than anomaly M-18.

In non-marine sedimentary sequences from the western United States the Cretaceous–Tertiary boundary dated by mammal fossils was located within the positive magnetozone corresponding to anomaly 29 (Butler *et al.* 1977). This may imply a discrepancy between the mammal and microfossil dating schemes, or it could be caused by hiatuses in the sections (Alvarez & Vann 1979). The position of the Palaeocene–Eocene boundary between anomalies 24 and 25 in continental deposits from Wyoming (Butler & Coney 1981; Butler *et al.* 1981) agrees with its location in the Umbrian marine limestones (Lowrie *et al.* 1982).

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Magnetic polarity timescales

The location of the Cretaceous–Tertiary boundary between anomalies 29 and 30 (Lowrie & Alvarez 1977*a*, *b*; Roggenthen & Napoleone 1977) was used by LaBrecque *et al.* (1977) in revising the Heirtzler *et al.* (1968) magnetic polarity timescale. The age of the older reversal boundary of anomaly 2A was taken to be 3.32 Ma and that of the Cretaceous–Tertiary boundary as 65 Ma. Intervening Cainozoic reversal boundaries were dated by interpolation, and late Cretaceous anomalies 30–34 were dated by extrapolation. Ness *et al.* (1980) made a thorough review and assessment of previous magnetic timescales. They introduced an additional calibration level at the base of anomaly 5 and used the revised decay constants and abundances for K–Ar dating (Steiger & Jaeger 1977) to modify the magnetic polarity timescale. However, the timescales of LaBrecque *et al.* (1977) and Ness *et al.* (1980) inherently assumed that ocean-floor spreading, particularly in the South Atlantic, remained constant for a very long time and the positions of stage boundaries relative to the polarity sequence were located indirectly at the levels appropriate to the boundary ages.

Lowrie & Alvarez (1981) proposed a revised geomagnetic reversal timescale for the Cainozoic and late Cretaceous in which 11 calibration points, established in magnetostratigraphic studies in Italian limestones, were used in addition to the 0 Ma age datum (figure 1). Absolute ages of the calibration points were provided by the best available dates for Palaeogene and late Cretaceous stage boundaries (Hardenbol & Berggren 1978; Obradovich & Cobban 1975), as recomputed with revised decay constants by Ness *et al.* (1980). The ages of reversal boundaries were calculated by linear interpolation between the calibration points.

An important result of magnetostratigraphic work in Italy and the western United States is the location of the Palaeocene–Eocene boundary between anomalies 24 and 25 (Butler & Coney 1981; Lowrie *et al.* 1982). This boundary had been placed within younger anomalies in earlier timescales (Heirtzler *et al.* 1968; LaBrecque *et al.* 1977; Ness *et al.* 1980). The revised position gives an improved fit between the ages of basal sediments in D.S.D.P. holes and the age of the anomaly on which the hole was drilled (Lowrie & Alvarez 1981). It changes reversal boundary ages in the late Palaeocene – early Eocene by as much as 3 Ma compared with the timescale of LaBrecque *et al.* (1977).

M-sequence magnetic timescales (Larson & Pitman 1972; Larson & Hilde 1975; Vogt & Einwich 1979) have been calibrated by the palaeontological ages of basal sediments retrieved from D.S.D.P. holes drilled to igneous basement on appropriate magnetic anomalies. The problem of dating the M-sequence anomalies is compounded by the paucity of reliable radiometric ages for the early Cretaceous stage boundaries. The ages used here have been recomputed with the new radiometric decay constants from those of Van Hinte (1976), which were largely intuitive. They give estimated ages of about 117 Ma for anomaly M-0, which is of early Aptian age (Channell *et al.* 1979), and 138 Ma for the Jurassic–Cretaceous boundary, which lies in the normal interval between anomalies M-17 and M-18 (Channell *et al.*, this symposium). The ages of other early Cretaceous M-sequence anomalies have been interpolated linearly between these two calibration points (figure 1).

Cox (1982) has prepared an alternative magnetic timescale that uses most of the magneto-stratigraphic correlations of Lowrie & Alvarez (1981) and minimizes changes in sea-floor spreading rate by modifying the ages of several late Cretaceous boundaries while keeping them within their ranges of uncertainty. For the M-sequence of reversals he adopted the early

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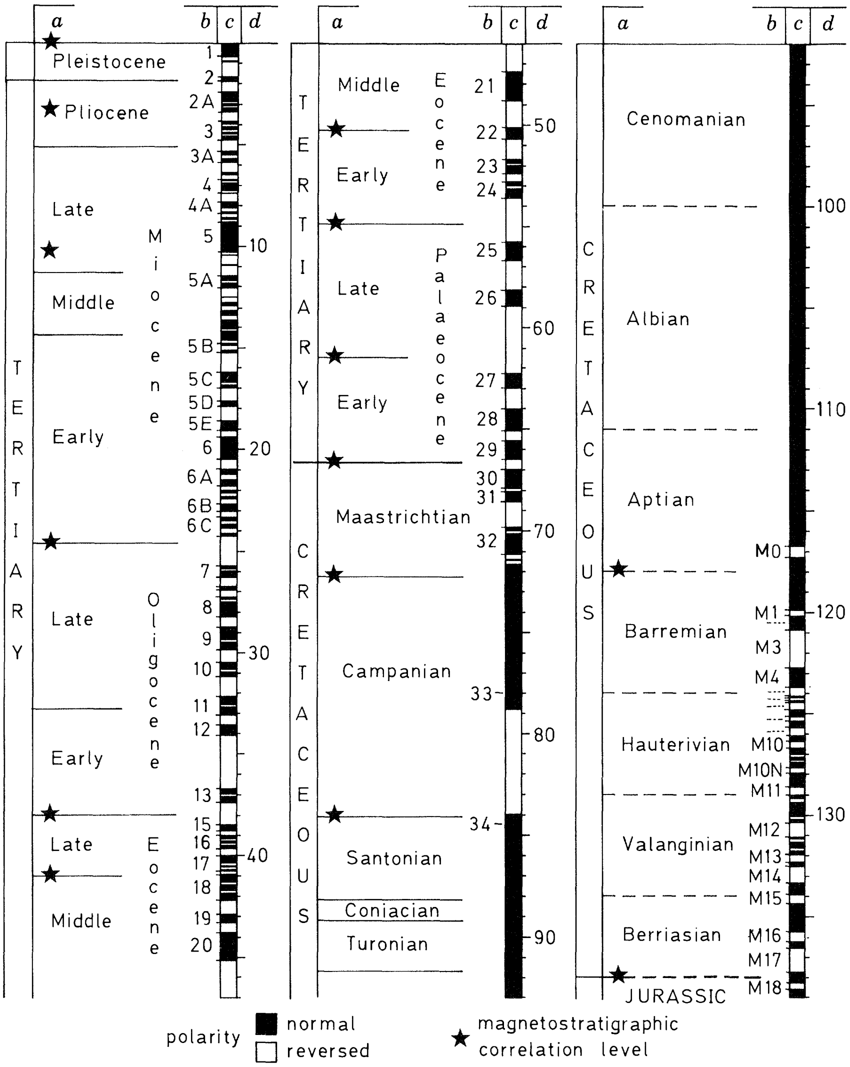

FIGURE 1. A magnetic polarity timescale for the Cainozoic and Cretaceous. Younger than 84 Ma this is the timescale of Lowrie & Alvarez (1981) with stage boundary ages after Ness *et al.* (1980). The older part of the timescale is modified from Larson & Hilde (1975) with stage boundary ages recomputed from Van Hinte (1976) by using revised K–Ar decay constants. Explanation of lettered columns: (*a*) geological epoch or stage, (*b*) oceanic magnetic anomaly number, (*c*) geomagnetic polarity, (*d*) estimated age in megayears.

Aptian age of anomaly M-0 (Channell *et al.* 1979; Lowrie *et al.* 1980*a*, *b*) and the Oxfordian age of D.S.D.P. hole 105 and dated the intervening reversals again by linear interpolation. Each stage from Aptian to Kimmeridgian was allotted the same duration (6 Ma). The age of the Jurassic–Cretaceous boundary becomes 144 Ma, which is within the range of radiometric estimates, but forces it to fall near the base of the normal polarity interval between magnetic anomalies M-15 and M-16. However, magnetostratigraphic correlations locate this boundary near the base of the normal interval between anomalies M-17 and M-18 (Ogg 1981; Channell *et al.*, this symposium).

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Sea-floor spreading rates in the Atlantic and Pacific oceans

Larson & Pitman (1972) computed sea-floor spreading rates for various accreting plate boundaries in the Atlantic and Pacific oceans by using the timescale of Hiertzler *et al.* (1968) for the Cainozoic and late Cretaceous and their own new timescale for the older M-sequence anomalies. They postulated an apparent world-wide increase in spreading rate during the Cretaceous quiet interval. By combining their published data (Larson & Pitman 1972, figure 8)

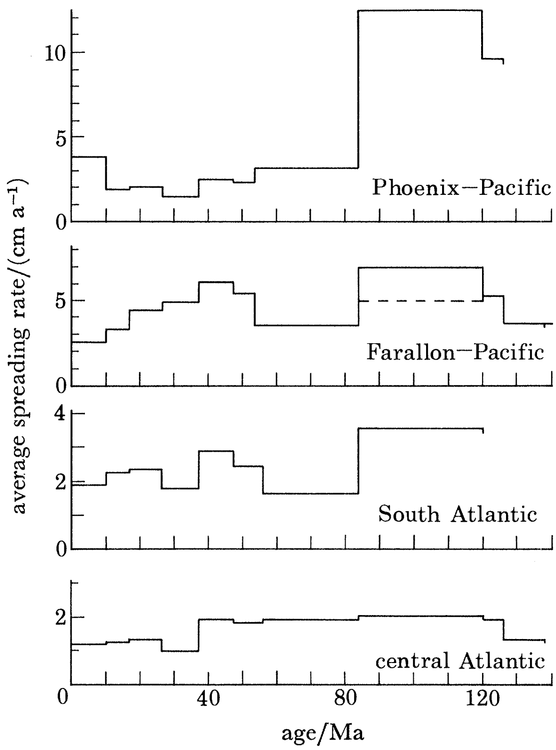

Figure 2. Average sea-floor spreading rates at accreting plate boundaries in the Atlantic and Pacific oceans. The rates have been recalculated from Larson & Pitman (1972), assuming the oceanic magnetic anomaly dates in figure 1.

with the new timescale in this paper (figure 1), the average spreading rates between selected oceanic magnetic anomalies have been recalculated (figure 2). Although lower than Larson and Pitman's estimates, average spreading rates during the Cretaceous quiet interval (age 84–120 Ma) are high at all plate boundaries except in the central Atlantic.

Changes in spreading rates are probably not as abrupt as suggested by figure 2. Cox (1982) notes that the timescale of Lowrie & Alvarez (1981) results in rapid apparent changes in Palaeogene sea-floor spreading rates compared with the timescale of LaBrecque *et al.* (1977). To minimize these changes he dropped the Eocene and Palaeocene intra-epoch calibration points, which are not as well constrained by available radiometric data as the beginning and end of these epochs (Hardenbol & Berggren 1978). However, the LaBrecque *et al.* (1977) timescale is founded on that of Heirtzler *et al.* (1968) and inherently assumes constant sea-floor spreading in the South Atlantic. Heirtzler *et al.* (1968) made this region their choice for

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a standard magnetic profile after comparison with the North Pacific, South Pacific and Indian oceans, in each of which marine magnetic profiles showed some inadequacy.

Authors of subsequent timescales (e.g. LaBrecque *et al.* 1977; Ness *et al.* 1980) have questioned the assumption of constant sea-floor spreading in the South Atlantic but continued it from necessity. Figure 2 shows that this assumption is probably not justified. However, the spreading rates in the central Atlantic imply long periods of constant spreading in this region. They remain within 5% of the average rate of 1.9 cm a^{-1} from middle Hauterivian (anomaly M-9, age about 126 Ma) to the end of the Eocene (anomaly 13, age about 37 Ma) and from then to the present they are reasonably constant at about 1.2 cm a^{-1}.

Frequency of geomagnetic field reversals

Analysis of the frequency of polarity reversals in the Heirtzler *et al.* (1968) magnetic timescale showed that older polarity chrons tended to last longer than younger chrons and that the frequency of reversals before about 45 Ma ago was lower than since then (Heirtzler *et al.* 1968; Cox 1969).

The numbers of reversals per megayear in the Cainozoic and late Cretaceous timescale of Lowrie & Alvarez (1981) have been averaged over age intervals of 2, 5 and 10 Ma duration, respectively (figure 3). The 2 Ma averages show large, apparently irregular, fluctuations in reversal frequency; averaging over 5 Ma and especially 10 Ma intervals indicates that these fluctuations are superposed on an almost linear trend. This suggests that the average reversal frequency of the geomagnetic field has been steadily increasing since the late Cretaceous.

The observed trend could result from a progressive loss of older reversals; however, the polarity sequences in the oceanic crust and in pelagic limestone sections are almost identical for this time interval. Oceanic basalts contain titanomagnetites, or titanomaghaemites, which have a thermoremanent magnetization acquired during initial cooling of the lava; pelagic limestones contain detrital magnetite that acquired a post-depositional remanence by statistical alignment with the ambient field shortly after deposition of the sediment. It is unlikely that the original reversal record would be subsequently eroded in exactly the same way in rocks with such disparate magnetic mineralogies and magnetization processes.

Cox (1969) found that the lengths of polarity intervals in the reversal timescale based on radiometrically dated young lavas, and in the Heirtzler *et al.* (1968) timescale for ages less than 10.6 Ma, conformed well to an approximately Poissonian distribution (Cox 1968), but that older reversals deviated markedly. He attributed the discrepancy to short polarity events (of duration 0.03 Ma or less) that were unresolved in the oceanic magnetic record, and noted that about 30 randomly distributed short events before 45 Ma age would be needed to change the observed distribution to a Poissonian one.

The magnetostratigraphic investigations in pelagic limestone sections have not clarified this point because their resolution is comparable with that of the oceanic magnetic anomaly studies. The pelagic limestones in which magnetostratigraphic correlations established the calibration tie points (figure 1) had sedimentation rates of 3–15 m Ma^{-1}. The palaeomagnetic sampling interval was adjusted to give a resolution of about 0.05 Ma in the magnetostratigraphic polarity zonations. Cox (1969) inferred that the smallest width of a strip of magnetized oceanic crust detectable with a magnetometer at the sea surface is around 1 km. With the average spreading rates of figure 2 this represents a resolution of about 0.03 Ma at the fast spreading

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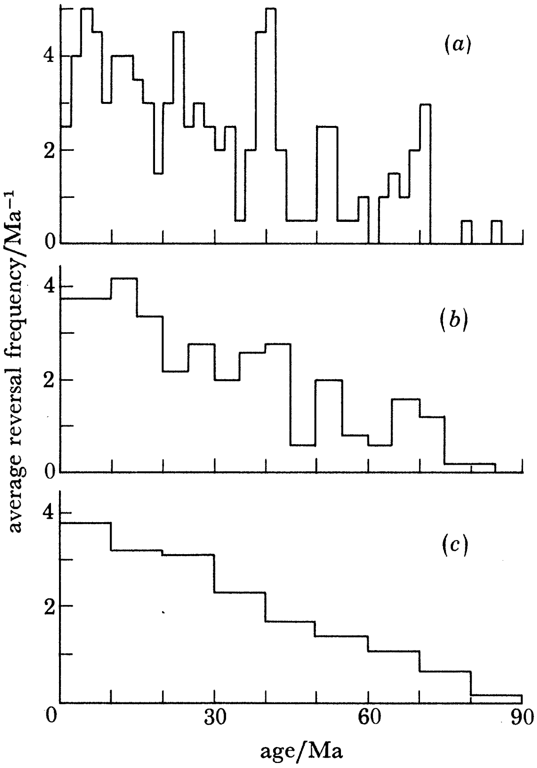

FIGURE 3. The frequency of geomagnetic polarity reversals since the late Cretaceous, averaged over intervals of (*a*) 2, (*b*) 5 and (*c*) 10 Ma, respectively.

Farallon–Pacific plate margin and about 0.06 Ma at the comparatively slowly spreading central Atlantic plate margin. Polarity chrons of duration 0.04 Ma or shorter could be missing from both records. Several additional reversals not found in the oceanic record have been reported in magnetostratigraphic studies, especially in the Cretaceous (Keating & Helsley 1979; Lowrie *et al.* 1980*b*).

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