Magnetostratigraphy and the geomagnetic polarity record

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ABSTRACT

The global nature of geomagnetic polarity reversals makes magnetic polarity stratigraphy a powerful method of correlating one stratigraphic section with another, especially when carried out in conjunction with a biostratigraphic study. Evaluated together, the two data sets provide a more reliable and finer correlation of stratigraphic sections with each other —or with a magnetic polarity time scale— than is possible with one data set alone. An interesting and potentially valuable application of integrated magneto- and biostratigraphy studies is in sedimentary basin analysis, for example in determining sedimentation rates, or as the time base for analysis of cyclical patterns in sedimentation. Where different paleontological dating schemes have been used to date sections, the magnetic stratigraphy allows these to be compared and synchronised with each other. An important result of magnetostratigraphic research has been the correlation and dating of the polarity history obtained from oceanic magnetic anomalies.

The interpretation of lineated marine magnetic anomalies has provided a continuous record of geomagnetic polarity since the onset of the present phase of sea-floor spreading, following the break-up of Pangea in the early Jurassic. The dominant characteristics of this time interval are two sequences of alternating geomagnetic polarity, the younger of which has lasted from the Late Cretaceous to the present. The older M-sequence falls in the Late Jurassic and Early Cretaceous, and is separated from the younger sequence by the Cretaceous quiet interval, lasting about 35 Myr, in which no magnetic reversals took place. Almost all the geomagnetic po-

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larity intervals in these reversal sequences have been corroborated and dated by coordinated magnetostratigraphic and biostratigraphic research in pelagic sedimentary sections. Magnetic polarity times scales for the late Mesozoic and the Cenozoic adopt the oceanic magnetic polarity sequence as basis, as it is the best continuous record available for this time period. Several versions of the magnetic polarity time scale have been proposed. They differ from each other mainly in the number of tie-levels used and the optimum ages accepted for the tie-levels, and reflect different critical evaluations of the acceptable radiometric age data base.

INTRODUCTION

Although the causes are not understood, the fact that the dipole component of the earth's main magnetic field of internal origin has changed polarity in the geological past is well documented in the remanent magnetizations of rocks. Early in the present century, Brunhes (1906) found volcanic rocks that had magnetization directions opposite to that of the present field, and Matuyama (1929) showed that the polarities of oppositely magnetized groups of lavas were dependent on their stratigraphic position relative to one another. However, self-reversal of the thermoremanent magnetization by a magnetomineralogical mechanism was the favored explanation of many rock magnetists for the reversed magnetizations in these igneous rocks. Laboratory experiments demonstrated that, during cooling, certain igneous rocks acquired a magnetization which was antiparallel to the ambient field. With the development of a precise radiometric dating technique and its application to lavas of known magnetization polarity, the history of geomagnetic polarity reversals within the last few million years was convincingly documented. The specter of self-reversal as the cause of all observed reversed polarities was finally laid to rest by the discovery of the same polarity sequence in modern deep sea sediments (Opdyke *et al.*, 1966). A completely different, isothermal process is responsible for the depositional remanent magnetization of sediments and thermally activated self-reversal can be ruled out. This does not discount the possibility that reversed magnetizations in some igneous rocks may in fact be due to a self-reversal mechanism.

MAGNETIC FIELD BEHAVIOR DURING A POLARITY REVERSAL

The behavior of the earth's magnetic field during a polarity change has been studied in igneous and sedimentary rocks. The change of paleofield direction to an antipodal value is observed as simultaneous transitions of the inclination and declination. Paleointensity results from modern deep-sea sediments and slowly cooled igneous intrusives indicate that the intensity of the magnetic field decreases to a low value during the transition (Opdyke *et al.*, 1973; Dodson *et al.*, 1978; Prévot *et al.*, 1985). This is interpreted as evidence for disappearance of the dominant dipole component of the geomagnetic field.

Although the dipole field decays during a reversal, the non-dipole field evidently does not and a standing non-dipole field component remains during the reversal (Hillhouse and Cox, 1976). The morphology of the transitional fields is assumed to be axisymmetric. Current research is aimed at establishing whether the transitional field configuration is controlled by the quadrupole component or the octupole component of the non-dipole field (Fuller *et al.*, 1979).

From transition studies in oceanic sediments with constant sedimentation rates, the duration of a directional transition has been estimated to last about 4,000-5,000 yr; the intensity change takes about twice as long as the directional change (Opdyke *et al.*, 1973). For magnetostratigraphy these are rapid changes. In a pelagic limestone section with sedimentation rate 10 m/Myr the transition is represented by a mere 4-5 cm of section. The average length of a polarity interval in the Cenozoic was about 450,000 yr (Heirtzler *et al.*, 1968). Thus each transitional interval is only about 1 % as long as the average polarity interval. Accordingly, about 1 % of the paleomagnetic samples from a magnetostratigraphic section may be expected to have intermediate directions.

MAGNETOSTRATIGRAPHY

Field Sampling

The usual purpose of a magnetostratigraphic study is to establish the record of magnetic polarity in a geological formation of known age. Although it is possible to piece together the record from a series of short outcrops, this procedure is usually not precise enough. It is desirable that the formation be exposed for sampling in as long and continuous sections as possible. The criterion of continuity of sedimentation is extremely important. To be sure that there are no gaps in sedimentation, as well as to date the reversals found, a magnetostratigraphic study must be accompanied by a thorough paleontological study. The combination of magnetostratigraphy and biostratigraphy in the same section provides mutually supportive data, which result in a more finely resolved stratigraphic evaluation and a larger number of reliable tie-levels for correlations between sections.

The sampling interval of a magnetostratigraphic section is determined by the required resolution. As can be inferred from the description of field behavior during a polarity transition, the shortest possible magnetic polarity interval must last longer than the time for a double transition, and is therefore around 10,000 yr. The shortest known, well documented intervals of constant magnetic polarity are known from the oceanic magnetic record to have lasted about 20,000 yr. A rough calculation shows that this is about the optimum resolution to be expected. The maximum magnetic anomaly V over the middle of a narrow block of oceanic crust with vertical magnetization contrast M, thickness t and width 2d, at a depth h below the magnetometer is approximately given by $V = 4MJ(\phi_1 - \phi_2)$, where $\phi_1 = \arctan\{d/h\}$ and $\phi_2 = \arctan\{d/(h+t)\}$. Although a significant proportion of the oceanic magnetic anomaly signal arises in the gabbroic seismic Layer 2B of the oceanic crust (Kent *et al.*, 1978), the observed signal is generated predominantly in the layer of basaltic lavas that form Layer 2A, which has a thickness of about 0.5 km and an average basalt magnetization of 5 A/m (Lowrie, 1979). Assuming a water depth of 5 km and a shipboard magnetometer resolution of around 50 nT, the resolvable width of a magnetized block is about 1 km, a figure reached by Cox (1969) from other considerations. On an anomaly system with a fast spreading rate (5 cm/yr) this corresponds to a time interval of about 20,000 yr. This is approximately the maximum resolution of the oceanic magnetic reversal record.

The goal of a magnetostratigraphic study is usually to match this resolution. The desire to locate and define accurately the shortest magnetozones in the section must be balanced against the effort of sampling, measurement and data processing. When the sampling density is so high that the number of samples taken is much greater than the number of magnetozones expected in the section, the manner in which the samples is distributed stratigraphically is not important. If the goal of the study is to define the magnetic polarity zonation completely but economically with as few samples as necessary, the geometry of the sampling scheme becomes important and optimum sampling is achieved with uniform stratigraphic spacing (Johnson and McGee, 1983). In many sections of pelagic limestone, the sedimentation rate is typically around 10 m/Myr. In order to achieve a resolution of 20,000 yr, a sampling interval of 20 cm would be needed, which in a 100 m thick section requires taking 500 samples. Every sample must be processed and remeasured many times, and consequently the amount of effort required for this resolution quickly becomes disproportionate to the goal. A suitable compromise consists of taking samples at 1 m intervals, corresponding to 100,000 yr resolution. After comparison of the magnetostratigraphy with the polarity sequence in the oceanic record, the section can be resampled in more detail where shorter intervals are expected. This practical compromise emphasises the confirmation of known reversals at the expense of discovering short events missing in the oceanic record.

Laboratory and Analytical Procedures

The samples from a magnetostratigraphic section must be subjected to the same rigorous progressive demagnetization procedures as samples from a conventional paleomagnetic investigation. Only in this way can the stable characteristic remanent magnetization direction be determined properly, preferably as the straight line leading to the origin of a vector diagram (As and Zijderveld, 1958). Less rigorous methods are inadequate and can lead to false interpretation. For example, in a magnetostratigraphic study of Oligocene sections in the Western U.S.A., Prothero *et al.* (1982) classified many samples with positive inclinations as reversed because the directions showed tendencies towards the negative polarity hemisphere during progressive demagnetization. The fact that stable directions were not established implies that the demagnetization technique was inadequate or the samples were unstable. As a result, the positions of reversal boundaries in the interpreted polarity sequence were wrongly identified. Absolute dates for ash flows in the magnetostratigraphic section were wrongly correlated with the polarity sequence, and incorporated subsequently as incorrect datum points in the magnetic time scale of Berggren *et al.* (1985).

The history of geomagnetic polarity is usually portrayed as a binary sequence of normal and reversed polarities, intermediate directions being regarded as not representative of a stable state. The polarity data are customarily shown as a column of alternately black (conventionally used for normal polarity) and white (for negative polarity) magnetozones. The interpretation of the limits of magnetozones can be made from plots of inclination and declination against stratigraphic position (Fig. 1). The individual points in each plot should be connected by straight lines to aid the viewer, as it is otherwise difficult to obtain a correct visual impression of the quality of the data. When the data points are unconnected, it is easier to overlook incompatible directions corresponding to disturbed parts of the section, and short intervals of opposite polarity to the dominating polarity of a magnetozone are more easily «lost». A convenient way of combining inclination and declination in a single polarity parameter is to calculate the latitude of the geomagnetic pole when the formation magnetization was acquired. For this calculation, declinations must be corrected for any post-depositional rotation and the paleolatitude of the site, calculated from the mean inclination, must be used instead of the present latitude. The boundaries for the magnetozones are located at the zero-crossings of the plot of polar latitude (sometimes misleadingly labelled VGP latitude), or of the inclination, or at the corresponding intersections of declinations with a value 90° away from the vector mean declination (Fig. 1).

Ideally each magnetozone should be represented by several samples. However, in many sections individual samples are found with polarity an-

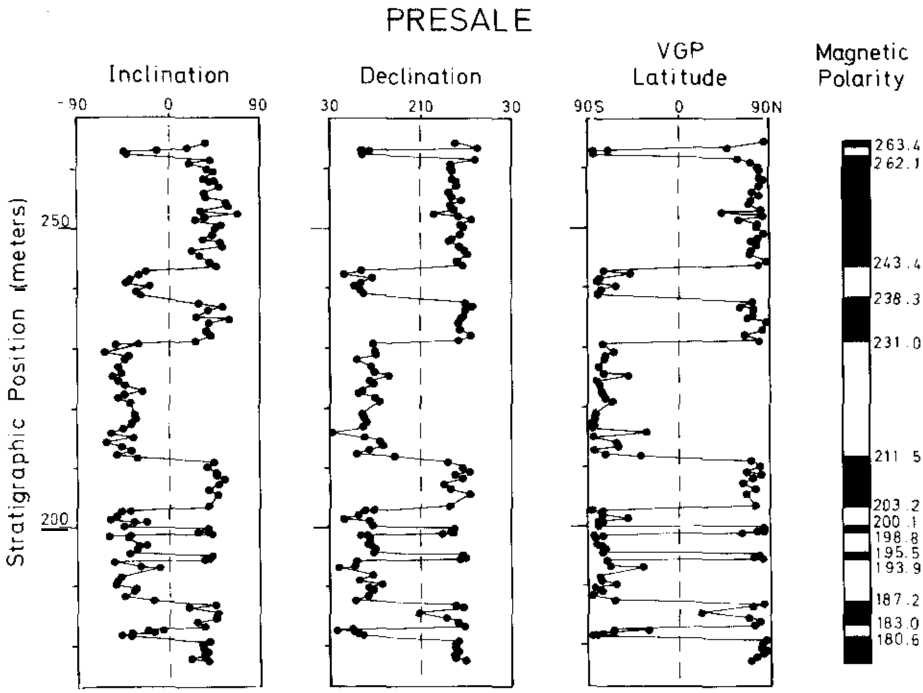


Fig. 1. Stratigraphic variations of inclination, declination and the latitude of the virtual geomagnetic pole in the Presale section of the Lower Cretaceous Maiolica limestone in Umbria, Italy (Lowrie and Alvarez, 1984).

tipodal to that of their neighbours. The anomalous directions may be due to unstable magnetizations, or to disturbance of the section. They may have a geomagnetic origin and represent only temporary excursions of the magnetic pole away from a stable position, or they may in fact represent very short polarity chrons which are too short to be resolved in the conventional oceanic magnetic record. A simple but satisfactory method of designating single-sample magnetozones in a polarity column is to draw them only half the width of the column (see, for example, Ogg *et al.*, 1984). Extra, short magnetozones found in a stratigraphic section which have no counterpart in the oceanic record may be real, and their existence is extremely important, but unless they are verified in other sections, their reality is suspect.

Magnetostratigraphy and Sedimentary History

Identification and Correlation of Magnetic Polarity Zones

To identify individual magnetic polarity chrons in a magnetic time scale, it is necessary to use a suitable nomenclature. Cox (1982) proposed a simple and logical convention which is employed with minor modification here. The chrons are numbered sequentially backwards in time from the present, in correspondence with the number of the oceanic magnetic anomaly from which each polarity chron is interpreted. Each chron consists of a normal polarity subchron and a negative polarity subchron, which are distinguished by attaching the letter n or r to the chron number. Thus the present interval of normal geomagnetic polarity corresponds to chron 1n, while the Cretaceous- Tertiary boundary is correlated with the upper part of chron 29r (Alvarez *et at.*, 1977). This nomenclature will be used in the rest of this paper.

The most common method of identifying a sequence of magnetozones is to find the equivalent pattern in the oceanic record. If a distinctive «fingerprint» is not present, an unambiguous correlation or identification may not be possible. The pattern recognition is usually done visually, but an alternative, quantitative method is by the use of correlograms (Langereis *et al.*, 1984). This involves computing the cross-correlation function of polarity interval lengths at successive matching positions as one polarity sequence is moved progressively past the other (only alternate matches in which intervals of like polarity are matched need be considered). Lowrie and Alvarez (1984) used correlograms (Fig. 2) to correlate the sequences of magnetozones in Italian magnetostratigraphic sections of Late Jurassic and Early Cretaceous age to the M-sequence polarity record (Fig. 3). The correlations of the magnetozone record in the Bosso section with chrons M13-M19 is the only one that is statistically significant. Three significant correlations are possible in the Gorgo a Cerbara section, and four are possible in the Presale section. The oldest correlations can be ruled out as they would imply the age of the section to be Jurassic, whereas it is known to be Early Cretaceous. Lithological arguments favor the youngest significant correlation in each section.

The correlation of sections by means of their magnetostratigraphy necessitates that they have the same history of sedimentation rate. For correlation with the magnetic reversal record derived from interpretation of oceanic magnetic anomalies, the sedimentation rate and sea-floor spreading rate must both be constant over the time interval investigated. Variable sedimentation rate distorts the thicknesses of magnetozones and makes recognition of a distinctive pattern difficult or even impossible. Channell and Grandesso (1987) found a poor visual match between the

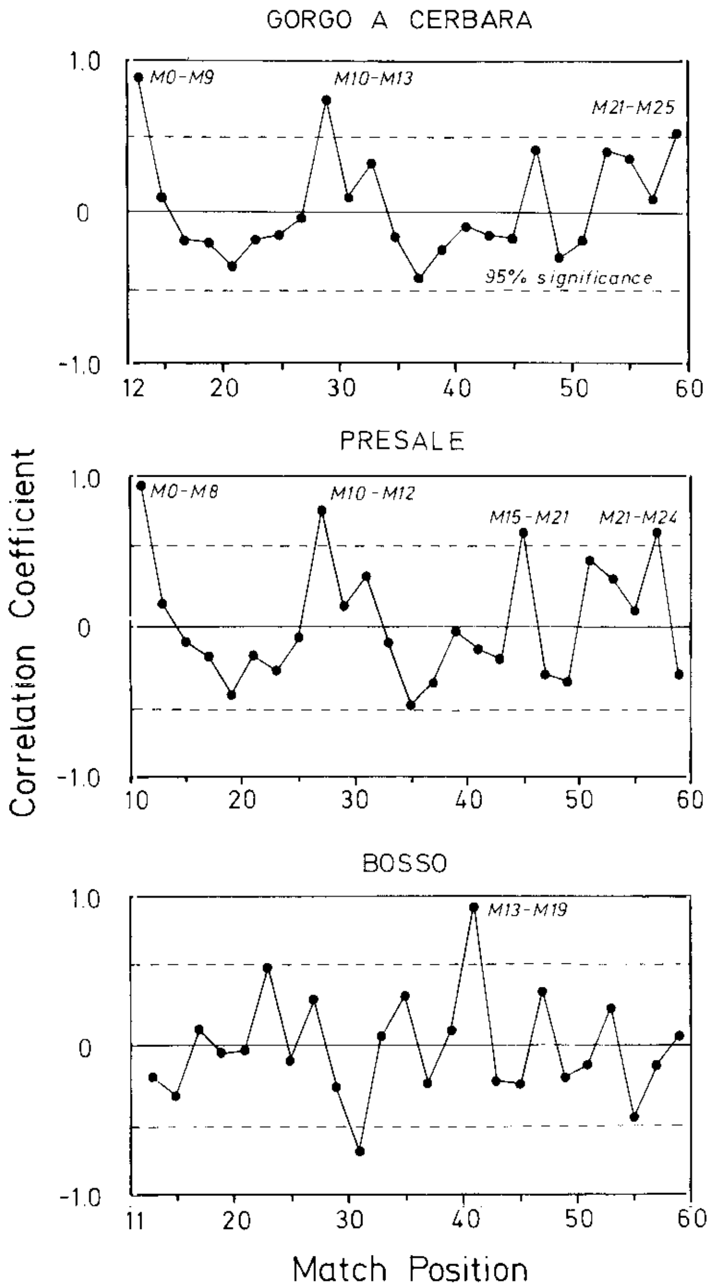


Fig. 2. Correlograms for the possible correlations of the Bosso, Presale and upper Gorgo a Cerbara sections with the M-sequence magnetic reversal record (Lowrie and Alvarez, 1984).

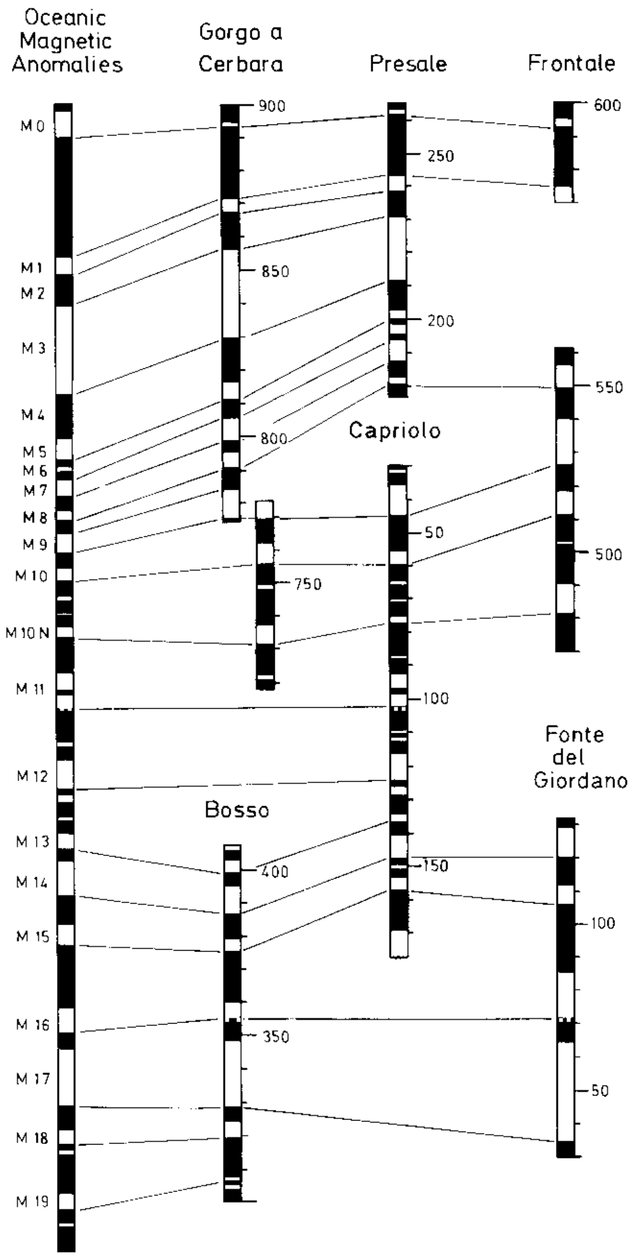


Fig. 3. Correlations of Lower Cretaceous M-sequence magnetostratigraphic sections in Umbria and the Southern Alps, Italy. Descriptions of sections: *Umbria:* Bosso (Lowrie and Channell, 1984a), Gorgo a Cerbara, Presale, Frontale (Lowrie and Alvarez, 1984), Fonte del Giordano (Cirilli *et al.*, 1984); *Southern Alps:* Capriolo (Channel *et al.*, 1987).

marine polarity record and the magnetic stratigraphies in the Xausa and Frisoni sections in the Italian Southern Alps, which they postulated was due to changing sedimentation rate in these sections, or to variation in sea-floor spreading rate during formation of the corresponding M-sequence anomalies. The close agreement of the coeval magnetic stratigraphy in the Umbrian Bosso section with the marine polarity record favored the interpretation of changing sedimentation rates in the Southern Alps sections and allowed their magnetostratigraphies to be corrected correspondingly.

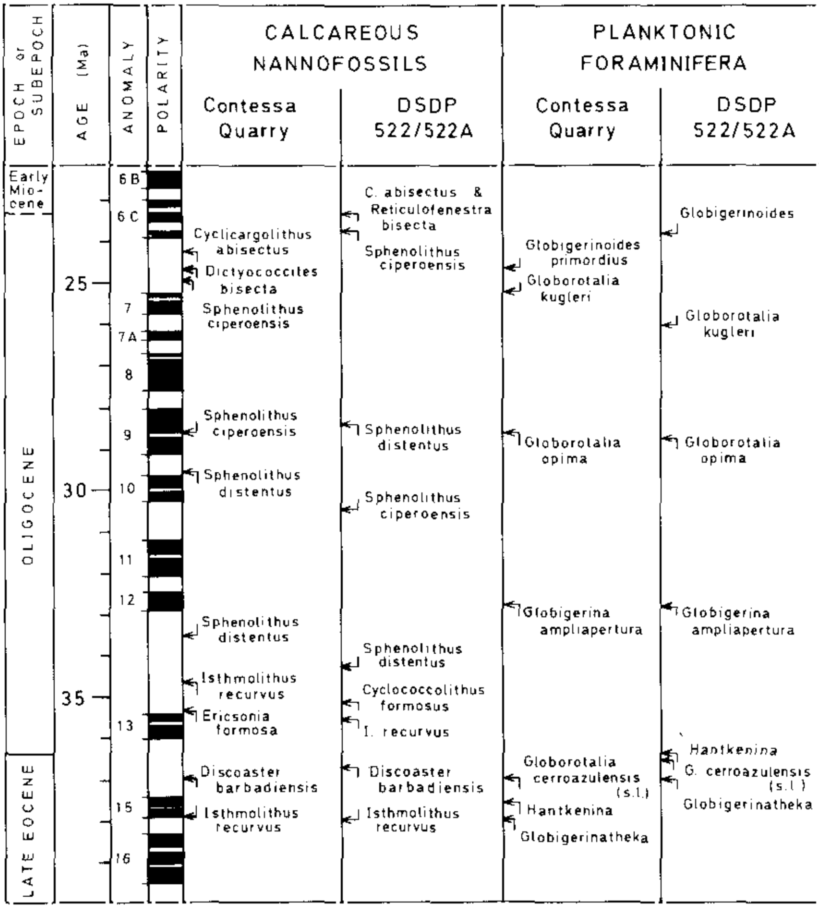
Magnetostratigraphy and Basin Analysis

The ability to provide information about sedimentary history makes magnetostratigraphy a powerful tool for basin analysis. Once a section has been correlated with the magnetic time scale, the sedimentation rate can be estimated. The tie-lines of the correlations also dramatize minor fluctuations of sedimentation rate within individual sections. Magnetostratigraphic correlations in the Maiolica limestone in the Italian Appenines (Fig. 3) give valuable information about the Early Cretaceous development of the Umbria-Marches sedimentary basin. Contemporaneous differences in sedimentation rate within the basin are seen by comparing the Presale section, which has a sedimentation rate of 7.1 m/Myr, with the upper part of the Gorgo a Cerbara section in which the sedimentation rate (9.5 m/Myr) is more than 30 % greater. The difference reflects the positions of the sections on opposite sides of a listric normal fault that was active during sedimentation. The Presale section represents deposition on a seamount on the upthrown side, while the Gorgo a Cerbara section was deposited in the deeper basin on the downthrown side (Lowrie and Alvarez, 1984). Further scrutiny shows a marked variation in sedimentation rate during Maiolica deposition. The sedimentation rate in the Bosso section, which spans the Jurassic/Cretaceous boundary, is 9.6 m/Myr. Higher sedimentation rates are found in the coeval lower parts of the Gorgo a Cerbara and Frontale sections (15 m/Myr and 19 m/Myr, respectively), although these sections are about 50 km apart. The increase in sedimentation rate is accompanied by a marked decrease of the intensity of magnetization due to dilution of the principal magnetic mineral (magnetite) in the sections. This dilution results in deterioration of the paleomagnetic signal, which becomes too weak to measure reliably. Polarity chrons M11r to M13n have not been found in Umbrian magnetostratigraphic research, but they have been securely located in the Capriolo section in the Southern Alps (Channell *et al.*, 1987).

COORDINATION OF MAGNETOSTRATIGRAPHY WITH BIOSTRATIGRAPHY

Mercanton (1926) first recognized that a paleomagnetic reversal is a globally synchronous event; the record of polarity is independent of geographic location. One of the major interests of magnetostratigraphic research has been in the correlation and dating of the polarity history obtained from oceanic magnetic anomalies, and its most valuable application has invariably been in conjunction with a detailed biostratigraphic study. The two disciplines reinforce each other and increase the number of usable tie-levels for correlation. The location of paleontological stage boundaries of known radiometric age relative to the magnetic polarity record provides invaluable tie-points for the development of magnetic polarity time scales. In turn, this enables the estimate of numeric ages for paleontological zones, first appearance datum (FAD) and last appearance datum (LAD) levels.

An interesting example of the combined use of magnetostratigraphy and biostratigraphy is seen in the results from pelagic carbonate sections in the Contessa valley in Umbria, Italy (Lowrie *et al.*, 1982) and site 522/522A of the Deep Sea Drilling Project (DSDP) in the South Atlantic (Poore *et al.*, 1982). The polarity sequences in both sections are excellent, unambiguous records. They agree with the polarity sequence derived from marine magnetic anomalies and expressed in the form of a magnetic polarity time scale (LaBrecque *et al.*, 1977). Both sections were dated using foraminifera and calcareous nannofossils, and the stratigraphic positions of key LAD and FAD levels were determined and located relative to the magnetic polarity time scale (Fig. 4). The numeric ages of appearances and extinctions of the same species in the Tethys (Contessa section) and in the South Atlantic (DSDP 522) were calculated. The average age of FAD's from the Tethys was 0.45 (± 0.39) Myr younger than in the South Atlantic, and the average LAD age in the Tethys was 0.57 (± 0.28) Myr older than in the South Atlantic (standard errors in parenthesis). The samples are small and the interpretation is open to speculation. The differences may be a real expression of evolutionary rates in the Tethys and South Atlantic, or they may express systematic discrepancies in the dating schemes used by different paleontologists (LaBrecque *et al.*, 1983). The example demonstrates two valuable applications for magnetostratigraphy, in permitting direct comparison of paleontologically dated sections with a common independent framework, and in determining the absolute ages to be associated with important paleontological events.



Type of event	Number of events	Age Difference (DSDP - Tethys)
FAD	6	$+0.45 \pm 0.39$ Myr
LAD	8	-0.57 ± 0.28 Myr

Fig. 4. Magnetostratigraphic positions of key LAD and FAD levels in the Contessa quarry section and DSDP site 522 (Lowrie, 1986; after Porre *et al.*, 1982), using the magnetic polarity time scale of LaBrecque *et al.* (1977).

GEOMAGNETIC POLARITY HISTORY

The history of geomagnetic polarity is now well established for Cenozoic and late Mesozoic time and quite accurate paleomagnetic polarity time scales have been developed on the basis of the dated reversal sequences. The detailed record of geomagnetic polarity from the late Middle Jurassic to the present is derived from the interpretation of lineated oceanic magnetic anomalies generated by the sea-floor spreading process (Heirtzler *et al.*, 1968). This reversal record is the longest continuous history of geomagnetic polarity available, and is the standard of comparison for all magnetostratigraphic studies in this range of geological ages. The reversal record can be divided broadly into three parts: a mixed polarity interval from the present back to the Late Cretaceous, a quiet interval of apparently constant normal polarity in the Middle Cretaceous, and a mixed polarity interval in the Early Cretaceous and Late Jurassic.

Neogene Magnetic Polarity History

Direct determination of radiometric ages gives a magnetic polarity time scale for chrons younger than 5 Myr (Mankinen and Dalrymple, 1979). The same polarity sequence has been confirmed also in modern deep sea sediments (Opdyke *et al.*, 1966; Foster and Opdyke, 1970; Opdyke, 1972). The precision of the K/Ar method is not adequate to resolve older short chrons by direct dating. However, it has been possible to extend absolute dating back to about 10 Myr in thick lava sequences by incorporating the stratigraphic relationships of the paleomagnetic sites (McDougall *et al.*, 1976). The polarity chrons of the uppermost Miocene were confirmed in a detailed magnetostratigraphic study of paleontologically dated marine clay sections on Crete (Langereis *et al.*, 1984). Although several studies of Pliocene and Late Miocene polarity history have been carried out in very long, overlapping piston cores of unconsolidated deep-sea sediments (e.g. Foster and Opdyke, 1970; Theyer and Hammond, 1974), the correlations with the magnetic time scale are sometimes ambiguous and some of the correlations have recently been questioned (Berggren *et al.*, 1985). Due to poor recovery in DSDP and ODP cores, and the scarcity of suitable exposures on land, most of the Miocene geomagnetic polarity history has not been verified and dated by independent magnetostratigraphic investigations.

Paleogene and Late Cretaceous Magnetostratigraphy

Virtually all the magnetic polarity chrons of the Oligocene, Eocene, Paleocene, and Late Cretaceous have been confirmed independently in magnetostratigraphic studies of pelagic sediment sections in the Tethyan realm (Fig. 5), and in DSDP cores. Detailed paleontological dating of the magnetostratigraphic sections has tied the positions of major paleontological stage boundaries to the geomagnetic polarity sequence. The association of absolute ages with these tie-levels has enabled the construction of magnetic polarity time scales, which in turn allow biozonations to be related to a framework of absolute ages, and provide dates for important paleontological events. Among the important stage boundary correlations in this time are the Oligocene/Miocene boundary near the chron 6Cn/6Cr boundary, the association of the Eocene/Oligocene boundary with chron 13r (Lowrie *et al.*, 1982), the position of the Cretaceous/Tertiary boundary near to the young margin of chron 29r, and the end of the Cretaceous quiet interval very close to the old edge of chron 33r (Alvarez *et al.*, 1977; Lowrie and Alvarez, 1977).

Early Cretaceous and Late Jurassic Magnetostratigraphy

The period from early Aptian to the end of the Campanian is characterised by constant normal polarity. Ocassional reports of negative polarity zones within the Cretaceous quiet interval have not been verified by repetition in other magnetostratigraphic sections. The start of this quiet interval, after the youngest polarity chron of the M-sequence (M0), has been established as very early Aptian, just younger than the Aptian-Barremian boundary, in pelagic carbonate sections in the Southern Alps (Channell *et al.*, 1979) and in Umbria (Lowrie *et al.*, 1980). Most of the Lower Cretaceous polarity chrons were confirmed in magnetic stratigraphy investigations in Umbria (Lowrie and Alvarez, 1983; Cirilli *et al.*, 1984), with the exception of chrons M11r to M13n where the limestone magnetization was too weak to be measureable (Fig. 3).
Recent studies in the Italian Southern Alps have independently confirmed and dated the polarity chrons from M8 to M23, including the interval M11r-M13n (Channell *et al.*, 1987). The magnetic stratigraphy of the Berriasian type section (Galbrun and Rasplus, 1984) was difficult to correlate unambiguously with the polarity record. The base of the section did not reach the Jurassic/Cretaceous boundary.

The location and correlation of the Jurassic/Cretaceous boundary to the magnetic polarity time scale illustrates a further application of magnetostratigraphy to the synchronisation of different paleontological dating schemes. There is no unanimous agreement among paleontologists on the definition of this boundary, due in part to the difficulty of correlating one paleontological dating scheme with another. Results from Umbria, the Southern Alps and DSDP sites (Fig. 6) give different relative positions of the boundary, depending on which paleontological scheme is

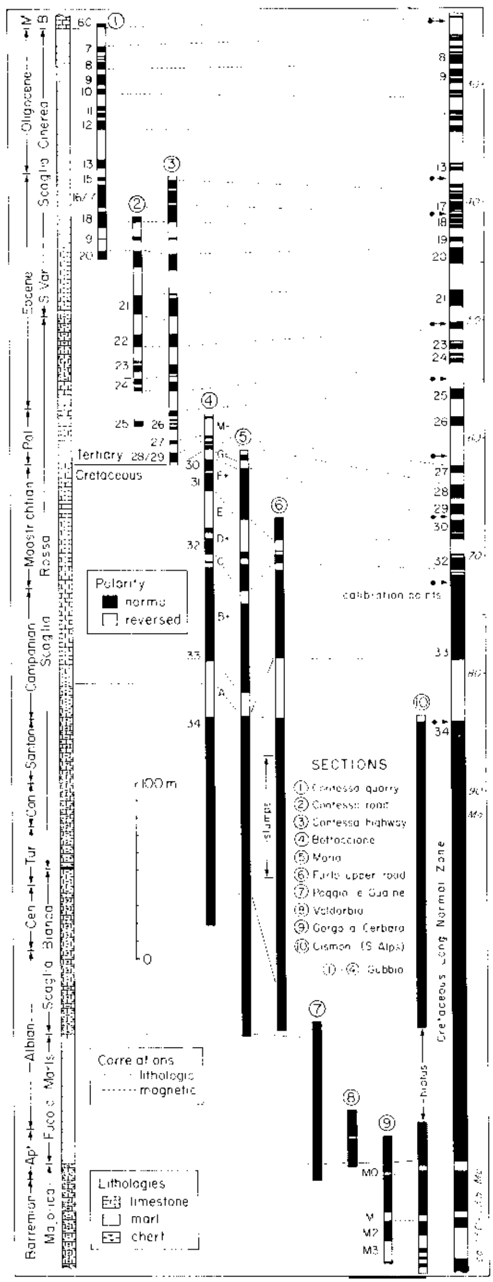


Fig. 5. Summary of magnetostratigraphic results from sections of Cenozoic, Late and Middle Cretaceous age in Umbria and the Southern Alps, and correlation with the oceanic magnetic polarity record (Lowrie and Alvarez, 1981).

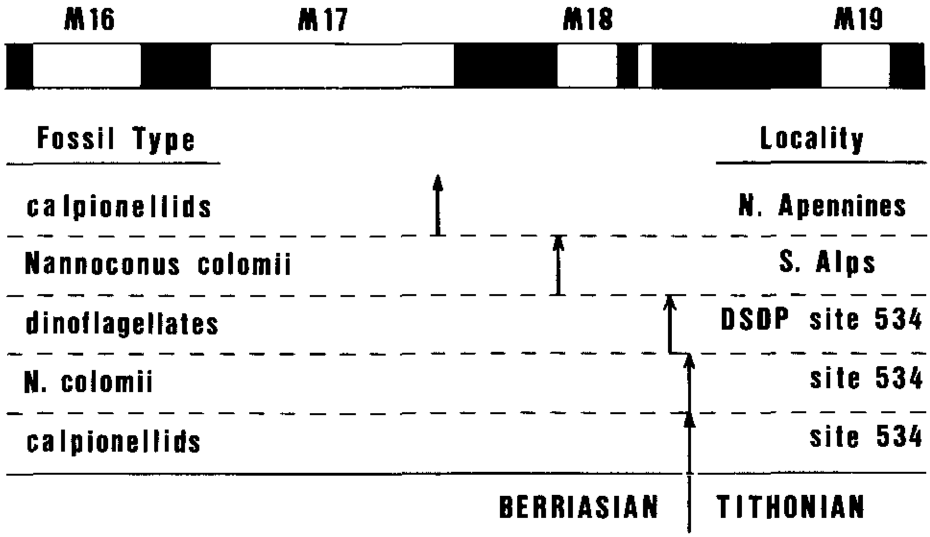


Fig. 6. Comparison of different locations of the Jusassic/Cretaceous boundary based on different fossil dating schemes (Lowrie and Channell, 1984b).

used (Ogg, 1984). Most investigations place the boundary near reversed polarity chron M18r, but it has been correlated as young as the old edge of chron M17r, on the basis of calpionellids in the Umbrian Maiolica limestone (Lowrie and Channell, 1984a), and as old as the middle of chron M19n, based on nannofossils, dinoglagellates and calpionellids in DSDP site 534 (Roth, 1983). Based on calpionellid biozonations, the Jurassic/Cretaceous boundary was placed in the base of chron M17r in the Umbrian Bosso section (Lowrie and Channell, 1984a). Channell and Grandesso (1987) reinterpreted the Umbrian calpionellid zonation and suggested a correlation within M18n. Depending on how the Jurassic/Cretaceous boundary is defined in terms of ammonite zonation and tied to the calpionellid zonation, they found alternative correlations in Southern Alps sections with the base of M17r or the top of chron M19n. Ogg and Lowrie (1986) have suggested resolving the uncertainty of paleontological definitions of the Jurassic/Cretaceous boundary by defining it at the base of chron M18r, a readily indentifiable event, and using this definition to calibrate the different paleontological dating schemes against each other. A position of the Jurassic/Cretaceous boundary consistent with this definition has been found in magnetostratigraphic sections in the Southern Alps (Channell *et al.*, 1987).

The M-sequence magnetic polarity history for the Late Jurassic (Fig. 7) has been confirmed in the Foza and San Giorgio magnetostratigraphic sections in northern Italy (Ogg, 1981) and in the Carcabuey and Sierra Gorda sections in southern Spain (Ogg *et al.*, 1984). Although several mag-

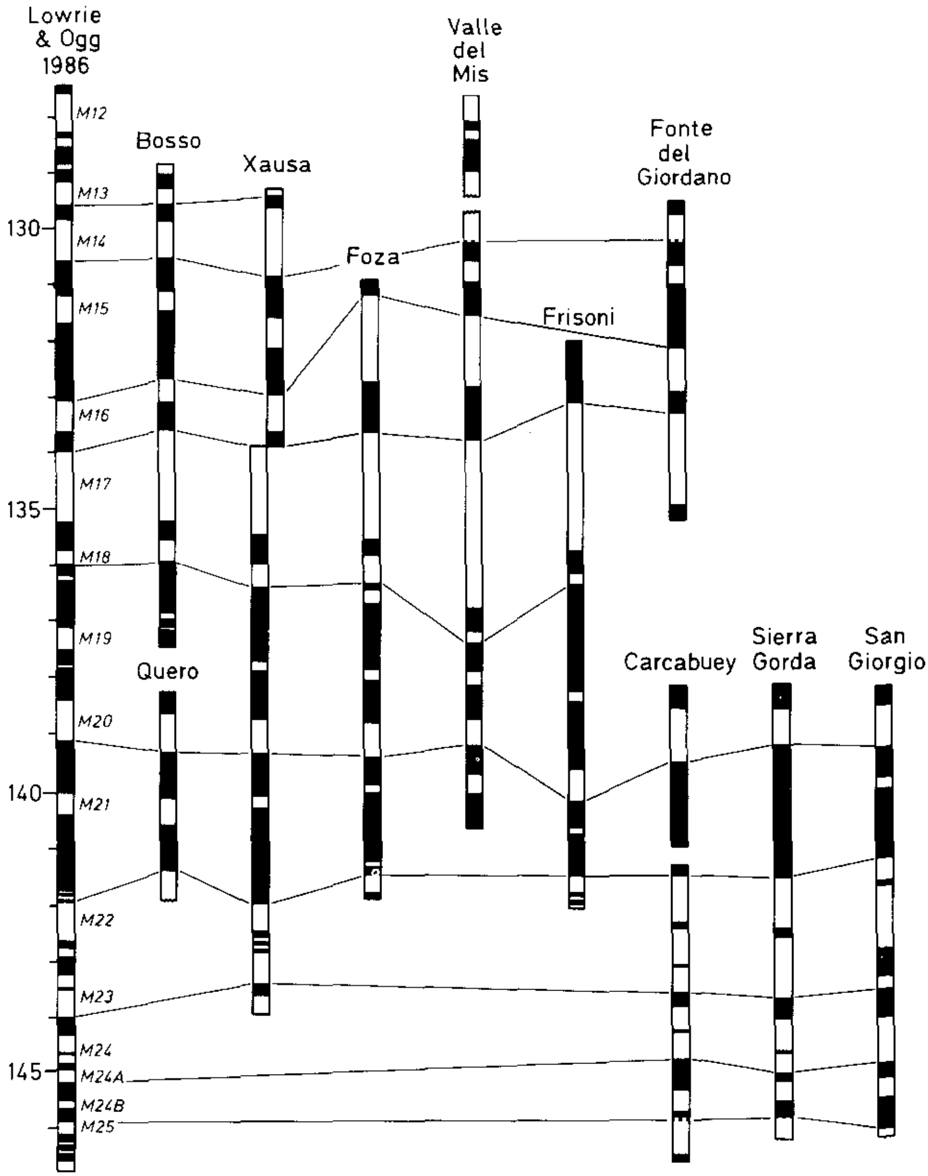


Fig. 7. Summary of magnetostratigraphic results from sections of Early Cretaceous and Late Jurassic age in Umbria, the Southern Alps and Spain, and correlation with the M-sequence magnetic polarity time scale (Lowrie and Ogg, 1986). Descriptions of sections: *Umbria:* Bosso (Lowrie and Channell, 1984a); *Southern Alps:* Xausa, Valle del Mis, Quero, Frisoni (Channell and Grandesso, 1987; Channell *et al.*, 1987), Foza and San Giorgio (Ogg *et al.*, 1984); *Spain:* Carcabuey and Sierra Gorda (Ogg *et al.*, 1984).

netozones are only represented by single samples, the magnetic stratigraphies in these sections can be correlated satisfactorily. An important result from these studies is the correlation of the Oxfordian/Kimmeridgian boundary with polarity chron M25n.

In their description of the M-sequence oceanic magnetic anomalies, Larson and Hilde (1975) identified the oldest negative anomaly as M25, corresponding to polarity chron M25r. Over older oceanic crust they found no correlatable lineated anomalies, and this was ascribed to a Jurassic quiet interval. The Jurassic quiet interval is not so well defined as the Cretaceous one, and there are different theories as to its origin. Subdued, low-amplitude linear anomalies have been correlated within this quiet zone, extending the anomaly sequence to M29 (Cande *et al.*, 1978). However, the anomaly pattern from the Pacific ocean does not agree with that from the Atlantic ocean, possibly due to poor resolution in the slowly spreading Atlantic ocean. Magnetic stratigraphy research in four overlapping sections of Middle-Late Oxfordian pelagic limestone near Aguilon in northern Spain (Steiner *et al.*, 1985) gave alternating polarity zones, but their sequence did not correlate well with the Pacific oceanic pattern.

Middle and Early Jurassic Magnetostratigraphy

For the early Jurassic polarity history before the Jurassic quiet interval there is no oceanic record. Without this guide, the correlation between magnetostratigraphic sections becomes much more difficult. Magnetic polarity sequences, dated with ammonites in some cases, have been reported for early Jurassic sections in Italy, Hungary and Spain but coeval polarity records do not agree well with each other (Channell *et al.*, 1982). Contemporaneous sections from different sedimentary environments might not be expected to show the same polarity sequence if the sections have different histories of sedimentation rate. However, even overlapping sections from nearby localities show poor internal consistency in Umbria (Channell *et al.*, 1982) and in northern Spain (Steiner *et al.*, 1985). The magnetostratigraphic studies were carried out with techniques comparable to those used in successful correlations of younger sections to the known marine record. Although the magnetic stratigraphies contain securely defined intervals of normal and reversed polarity, many of the observed magnetozones are defined only by single samples. This may imply inadequate sampling intervals, or that the frequency of reversals in the early Jurassic was high. Alternatively, the breakdown of consensus between early Jurassic sections may most simply be ascribed to variation of sedimentation rate within and between the investigated magnetostratigraphic sections. Possibly the depositional basins in the Tethys following the break-up of Pangea were small and characterised by inconstant sedimentation. This presents a serious obstacle to the possibility of establishing a standard polarity time scale for earlier times.

MAGNETIC POLARITY TIME SCALES

Magnetic polarity time scales for the last 5 Myr can be formed by direct radiometric dating of lavas, for which the K/Ar method is accurate enough to resolve individual polarity chrons lasting about 100,000 yr or longer. The older polarity chrons in time scales cannot be dated directly. Their ages must be estimated by linear interpolation between known tie-points such as, for example, the paleontological stage boundaries which have been correlated by magnetostratigraphy to the polarity sequence. Within each time scale ages are calculated to the nearest 0.01 Myr (10,000 yr), which is close to the achievable resolution. This relative accuracy is about 100 times better than the absolute accuracy with which tie-point ages are known.

The first magnetic polarity time scale for the Late Cretaceous and Cenozoic was constructed by Heirtzler *et al.*, (1968), primarily on the basis of a very long magnetic anomaly profile in the South Atlantic. The ocean ridge anomalies younger than 3.35 Myr were correlated directly to the radiometrically dated polarity time scale, and older anomalies were dated by extrapolation from this very short base. The time scale of LaBrecque *et al.* (1977) incorporated the following improvements: (1) the sequence of short reversals originally named anomaly 14 was dropped, as it was absent in most marine magnetic anomalies; (2) a revision of the relative lengths of the older Late Cretaceous chrons 29 to 34 was included; and (3) the magnetostratigraphic correlation of the Cretaceous/Tertiary boundary with reversed polarity chron 29r (Lowrie and Alvarez, 1977) was adopted as a tie-point, and associated with an age of 65 Myr. The ages of intervening polarity chron boundaries were calculated by interpolation. The sequence of polarity in the LaBrecque *et al.*, (1977) time scale has been used in all subsequent versions, stretched and contracted in linear segments between tie-points.

The major stage and sub-stage boundaries of the Upper Cretaceous and Paleogene were correlated to the geomagnetic polarity sequence as a result of extensive magnetostratographic research in pelagic marine carbonate sections on land (Alvarez *et al.*, 1977; Lowrie *et al.*, 1982) or sampled in the Deep Sea Drilling Project (LaBrecque *et al.*, 1983). Lowrie and Alvarez (1981) developed a magnetic polarity time scale for the Upper Cretaceous and Cenozoic based upon interpolations between 11 correlated tie-points, for which revised absolute ages had been calculated. Some of these ages were poorly chosen, and resulted in apparent sudden changes in sea-floor spreading rate. To minimize these effects while retaining as many of the magnetostratigraphic correlations as possible, Cox (1982) dropped the sub-stage boundaries as tie-points, and adjusted the absolute ages of tie-points within their error limits so as to obtain a time-scale which was consistent with a model of nearly constant sea-floor spreading.

A more radical polarity time scale for the Cenozoic was prepared by Berggren *et al.* (1985), who disregarded the magnetostratigraphic correlations of stage boundaries. They substituted a handful of selected radiometric age dates which met their own criteria. At least two of their tie-points are wrong, because they derive from the misinterpreted Oligocene magnetostratigraphy of Prothero *et al.* (1982).

On balance, the most appropriate magnetic polarity time-scale for the Late Cretaceous and Cenozoic is probably that of Cox (1982). It makes optimum use of available magnetostratigraphic correlations, while minimizing drastic changes in sea-floor spreading rate.

The construction of a magnetic polarity time scale for the M-sequence polarity chrons is handicapped by the absence of well dated tie-levels. Although an adequate number of the Lower Cretaceous and Upper Jurassic stage boundaries have been correlated satisfactorily to the M-sequence, the absolute ages for these boundaries are not well known. The polarity sequence is basically that obtained by Larson and Hilde (1975) from the interpretation of M-sequence magnetic anomalies in the North Pacific and Atlantic oceans. The original calibration of these anomalies was based upon the paleontological ages of the sediment in contact with igneous basement in DSDP holes drilled near the young end and the old end of the M-sequence. Assuming approximate absolute ages for the ends, the intervening polarity chrons were again dated by linear interpolation. The same method was used by Vogt and Einwich (1979), who introduced an additional early Valanginian calibration point for DSDP site 387 within the M-sequence. It is often uncertain whether a DSDP hole has indeed reached igneous basement, or whether it has merely encountered an intrusion. Also, the length of time between formation of the lava and deposition of the first datable sediments is not known. Finally, the paleontological dates often have large errors, amounting to an entire stage or more, and ages determined with different fossil types can differ appreciably.

As a result of the magnetostratigraphic investigations in pelagic limestone sequences in Italy chron M0 was correlated with the very earliest Aptian, close to the Barremian boundary (Channell *et al.*, 1979; Lowrie *et al.*, 1980). This was incorporated as one of two tie-points in a magnetic polarity time-scale for the M-sequence by Cox (1982); however, the older tie-point was a paleontological age from a DSDP hole. Ogg *et al.* (1984) correlated the Oxfordian/Kimmeridgian boundary with the M-sequence just younger than M25 in magnetostratigraphic sections in Spain (Fig. 7). Kent and Gradstein (1985) derived a magnetic polarity time scale for the

M-sequence, using these magnetostratigraphic ties for M0 and M25 and the age estimates by Harland *et al.* (1982) for the Oxfordian/Kimmeridgian and Barremian/Aptian stage boundaries. They then compared their new time scale with the magnetostratigraphic correlations of the other Lower Cretaceous and Upper Jurassic stage boundaries (Lowrie and Channell, 1984a; Lowrie and Alvarez, 1984; Ogg *et al.*, 1984) and estimated the absolute ages of these boundaries.

The Jurassic/Cretaceous boundary is not clearly defined paleontologically (Fig. 6), and Ogg and Lowrie (1986) proposed that an optimum correlation is with the base of polarity chron M18r. Lowrie and Ogg (1986) added this tie-point to the M0 and M25 correlations. They disregarded the age estimates of Harland *et al.* (1982) for these three tie points as invalid, and adopted the values given by Hallam *et al.* (1986). A magnetic polarity time scale for the M-sequence polarity chrons was then constructed by linear interpolation on the two segments M0-M18 and M19-M25; ages for M25-M29 were estimated by extrapolation of the M19-M25 segment. Ages were also estimated for the other Lower Cretaceous and Upper Jurassic stage boundaries from their correlations with the polarity time scale. Because of the additional Jurassic/Cretaceous tie point and the different ages assumed for the three calibration points, the ages of all polarity chrons in the time scale of Lowrie and Ogg (1986) are 5-10 Myr younger than in the time scale of Kent and Gradstein (1985), although the agree quite closely with earlier estimates of these stage boundaries by Van Hinte (1976a, 1976b). The uncertainties in absolute ages of calibration points result in large differences between M-sequence time scales (Fig. 8).

DISCUSSION

The record of geomagnetic polarity derived from analysis of oceanic magnetic anomalies is dominated by two sequences of mixed geomagnetic polarity, the younger extending from the present until the Late Cretaceous, and the older spanning the Early Cretaceous and Late Jurassic. They are separated by an interval of constant normal polarity —the Cretaceous Quiet Interval. As a result of oceanic magnetic anomaly surveys the polarity sequences have become securely established. They have been largely confirmed independently in magnetostratigraphic studies. Coordinated magnetostratigraphic and biostratigraphic investigations in pelagic sedimentary rocks have dated this history of geomagnetic polarity. The Cretaceous Quiet Interval lasted about 35 Myr in the middle Cretaceous, from the earliest Aptian until the Santonian/Campanian boundary. Apart from this interval —in which magnetostratigraphic correlations are not

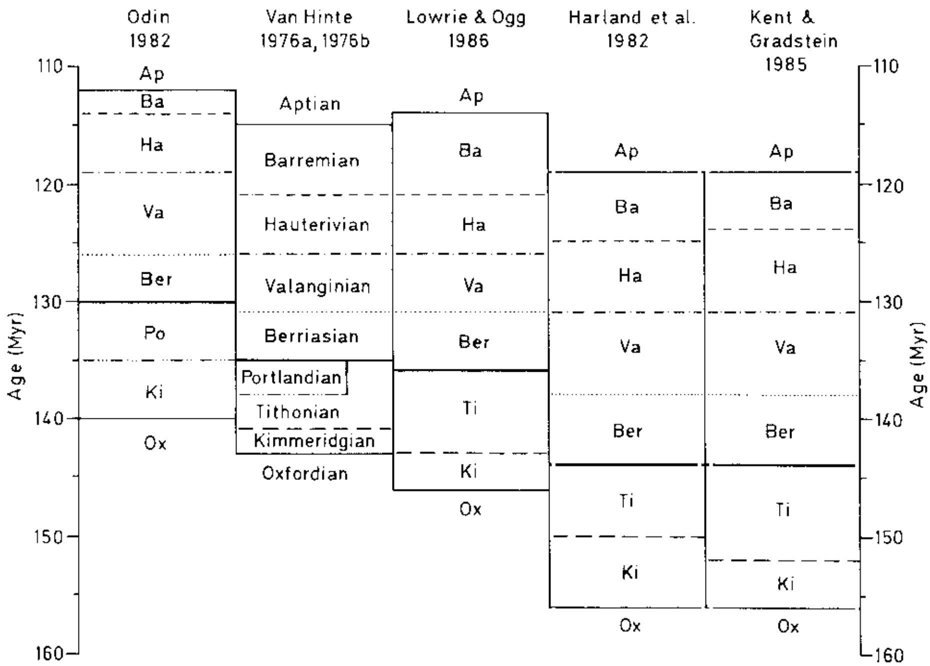


Fig. 8. Comparison of different time scales for the Early Cretaceous and Late Jurassic.

possible— most of the major paleontological stage boundaries from the Oxfordian/Kimmeridgian to the Oligocene/Miocene have now been correlated reasonably satisfactorily to the oceanic polarity sequence. These correlations have fostered the construction of magnetic polarity time scales covering the Late Jurassic, Cretaceous and entire Cenozoic. In turn, undated paleontological stage boundaries, as well as paleontological zones and events, have been dated by their correlations to these time scales. Different versions of the same polarity time scale reflect successive improvements in the definition of the polarity sequence, the number of correlated tie-levels used, and the optimum ages accepted for the tie-levels. They also reflect different critical evaluations of the radiometric age data base. Discrepancies between different versions of the Late Cretaceous and Cenozoic polarity time scale are small (Table 1); the differences for the M-sequence polarity chrons are larger (Fig. 8). In order to resolve them and produce definitive versions of either time scale, improvements in the number and quality of absolute age estimates for the tie-levels are needed.

TABLE 1

STAGE	LaBrecque *et al.*, 1977 Chron	Age	Lowrie and Alvarez 1981 Chron	Age	Cox 1982 Age	Berggren *et al.*, 1985 Age
Miocene						
	6Br	22.5	6Cr	24.6	24.5	23.7
Oligocene						
	15r	38	13r	38	38	36.6
Eocene						
	23	55	24r	54.9	55	57.8
Paleocene						
	29r	65	29r	66.7	65	66.4
Maastrichtian						
	top 33	72.5	top 33	72.3	73	75
Campanian						
	33r-34	80	33r-34	84.1	83	84
Santonian						

Table 1. Ages (in Myr) of Late Cretaceous and Paleogene stage boundaries in different polarity time scales. In the time scale of LaBrecque *et al.* (1977) the correlations of stage boundaries are inferred from their assumed ages, except for the Cretaceous/Tertiary boundary and in the Late Cretaceous. The magnetostratigraphic correlations of stage boundaries with the polarity chrons in the time scale of Lowrie and Alvarez (1981) are also used in the time scales of Cox (1982) and Berggren *et al.* (1985).

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