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

Magnetic Polarity Dating of Tectonic Events at Passive Continental Margins [and Discussion]

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Magnetic polarity dating of tectonic events at passive continental margins

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Palaeomagnetic studies of continental margin sediments drilled during Ipod Legs 47a (NW African margin), 47b (NW Portugal margin) and 48 (NE Bay of Biscay and SW Rockall margins) have provided magnetic polarity records for the late Neogene, early Palaeogene and Cretaceous periods. The general pattern of geomagnetic field polarity reversals during late Neogene times is now well established, and the absolute age of major polarity transitions during the past 5 Ma or so has been determined from palaeomagnetic studies of radiometrically dated igneous rocks (see, for example, McDougall *et al.* 1977). Consequently, a comparison of the late Neogene polarity sequences identified in the sediments of NE Biscay and the NW African margins, with this 'standard' polarity reversal time scale, allows the precise determination of sediment accumulation rates, and the times at which significant changes in these rates occurred. These comparisons have led to the recognition of a short but significant hiatus in early Pleistocene times in both areas. Such information has importance in evaluating the recent geological evolution of these margins.

For earlier geological periods the correlation between the magnetic polarity time scale and the geological time scale is less well established but studies of the type described here can still provide important information on the timing of certain geological events recorded in the magnetic anomaly patterns of the oceanic lithosphere. For example, in the early Palaeogene sediments cored at Sites 403–405, off the SW Rockall continental margin, a sequence of magnetic reversals has been identified, which shows a good correlation with marine magnetic anomalies 22–24. Since anomaly 24 is the oldest recognizable anomaly in the Atlantic, and lies immediately adjacent to the continental margins, the biostratigraphic age of this anomaly, determined from the nannofossil and dinocyst zonal determinations at Site 404, provides important information on the date of initial rifting of Rockall (together with the rest of NW Europe) from Greenland. The Site 404 results indicate that this important tectonic event occurred in early Eocene times, at about 52 Ma b.p., rather than at 60 Ma b.p. as was originally proposed by Heirtzler *et al.* (1968).

A further example of the potential value of this type of study is provided by the Cretaceous sediments cored at Sites 397, 398, 400A and 402A. A long section of predominantly normal polarity sediments at the latter three sites appears to correlate with the long Cretaceous interval of dominantly normal polarity identified in marine magnetic anomaly patterns. The combination of palaeomagnetic and biostratigraphic studies allows useful constraints to be placed on the maximum duration of this interval, and on the age of short reversals within and below it. This information has direct relevance to the interpretation of Mesozoic marine magnetic anomaly patterns in terms of the history of seafloor spreading and evolution of continental margins during the early stages of opening of the South Atlantic in Cretaceous times.

1. Introduction

The use of deep-drilling techniques to investigate the geological evolution of the ocean basins and their margins depends critically on an accurate knowledge of the geological ages of the sediments encountered. Without such knowledge, comparisons of contemporaneous palaeoenvironments in different regions cannot be made and the recognition of important geological events, represented by such features as sedimentary hiatuses and changes in sedimentation rates, becomes impossible. The most widely used technique for determining the relative age of deep sea sediments is micropalaeontology, and the contribution of this subject to our current understanding of the geological evolution of ocean basins cannot be over-emphasised. However, this technique does suffer from certain fundamental limitations. In particular, it requires the presence of suitable conditions for the existence and preservation of ancient marine organisms,

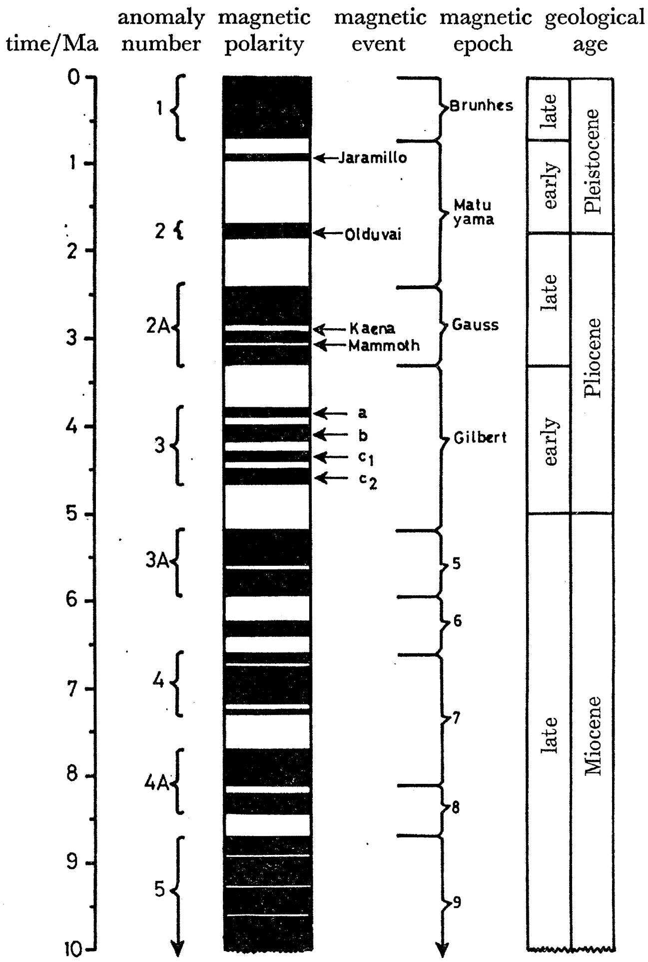

FIGURE 1. Magnetic polarity timescale for the past 10 Ma, compiled from the data of Cox (1969), Talwani *et al.* (1971) and Blakely (1974). Black, normal polarity; white, reversed polarity.

and the time resolution depends upon the identification of particular assemblages of rapidly evolving species. Unfortunately, the first appearance and extinction of species and changes in their relative abundances are often complicated by local ecological factors, or masked in the sedimentary record by dissolution effects. This can lead to local difficulties in identifying boundaries between micropalaeontological zones, and in ascribing a precise synchroneity to the same boundary identified in different regions.

In recent years an alternative dating technique has seen significant development. It is based on the fact that ancient geomagnetic field polarity reversals are commonly recorded in sedimentary sequences at their time of deposition. Since geomagnetic field reversals are essentially synchronous and worldwide in extent, this technique offers certain important advantages over micropalaeontological dating methods, but is itself subject to other limitations. For example, it is first necessary to demonstrate that the remanent magnetism of the sediment is stable, and was acquired at, or close to, the time of deposition. Furthermore, the pattern of magnetic reversals recorded in a particular sedimentary sequence will depend on both the reversal frequency at the time of deposition, and on the sedimentation rate. In the case of slow sedimentation rates at times of high reversal frequency, certain magnetic polarity events may be too

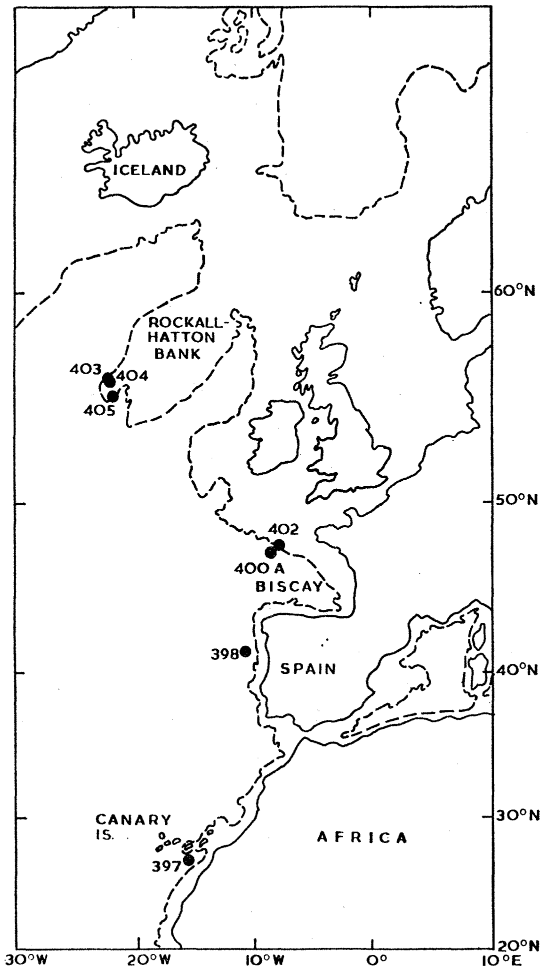

FIGURE 2. Locations of IPOD Leg 47 and 48 sites discussed in this paper.

short in duration to be recorded at all, and the occurrence of sedimentary hiatuses will considerably complicate the identification of particular magnetic polarity intervals. However, if the approximate age of the sediment is known from micropalaeontological information, it should be possible to 'focus in' on a particular part of the established polarity timescale, and thus to identify specific reversals in the sedimentary sequence. The combination of magnetic and biostratigraphic studies may then allow much more precise dating than is possible from palaeontological information alone. Furthermore, under favourable circumstances, magnetostratigraphy may be used to date sediments which contain impoverished fauna, or are entirely barren of characteristic zonal fossils.

The success of this approach depends ultimately on the reliability of the polarity reversal sequence, and on the establishment of a precise correlation between this sequence and biostratigraphic zonal schemes. The pattern of geomagnetic field reversals is now well established for the Cainozoic and late Mesozoic, from combined studies of marine magnetic anomalies, and the palaeomagnetism of sediments and radiometrically dated igneous rocks. The sequence of reversals during the late Neogene is shown in figure 1, compiled from the data of Cox (1969), Talwani *et al.* (1971) and Blakely (1974). However, the precise correlation between this polarity reversal sequence and biostratigraphic zonations is at present less well established. The only completely satisfactory means of achieving such a correlation is to undertake combined palaeomagnetic and biostratigraphic studies on the same sedimentary sequences, and the purpose of this paper is to present the results of such work, carried out on sediments of Quaternary and late Neogene, early Palaeogene and Cretaceous age, from the continental margins of Africa and northwest Europe, drilled on IPOD Legs 47 and 48 (figure 2). The implications of these results for the oceanographic and tectonic evolution of Atlantic passive margins will then be briefly reviewed.

2. Methodology

A high proportion of the palaeomagnetic results presented in this paper were obtained from shipboard measurements, carried out in the palaeomagnetic laboratory installed on D/V *Glomar Challenger* at the start of Leg 47. Remanence measurements were made on a Digico computerized magnetometer, with a noise level of about 0.03 μG,† and the stability of the magnetism was tested by means of an alternating field (a.f.) demagnetizer with a maximum field capability of 1000 Oe ($\approx$ 80 kA m^{-1}).

In the case of soft unconsolidated sediments, palaeomagnetic samples were taken by pushing 2.5 cm diameter non-magnetic plastic tubes, of 2.2 cm length, into the split core sections, and sealing the ends with adhesive tape. In more highly lithified sediments, 2.5 cm cylindrical samples were drilled from the core sections, or cube-shaped samples were cut by diamond saw. In all cases the uphole direction was carefully recorded on the sample by means of an orientation arrow before removal from the core section, and only sediments showing no visible signs of deformation were sampled.

Because of the rotary technique used for drilling IPOD cores, relative rotation frequently occurs between different segments of sediment within the core, and this may cause apparent changes in the declination of stable remanent magnetization. Consequently in this study the magnetic polarity has been assigned on the basis of the inclination of the stable remanent magnetization alone. Since all sites are situated at moderate to high latitudes in the northern hemisphere, positive (downward directed) inclinations are taken to signify a normal polarity, and negative (upward directed) inclinations reversed polarity.

In order to assess the reliability of the polarity determinations and to remove unwanted low coercivity components of remanence, the majority of samples were each separately subjected to progressive a.f. demagnetization until a stable directional endpoint was achieved. Results from samples exhibiting no such endpoint were discarded on the grounds that they were either unstable or carried a complex remanence that could not be satisfactorily decomposed by available laboratory procedures. In occasional groups of samples taken from particular units having a uniform lithology throughout, the behaviour of representative 'pilot samples' during detailed

† 1 gauss (G) = 10^{-4} T.

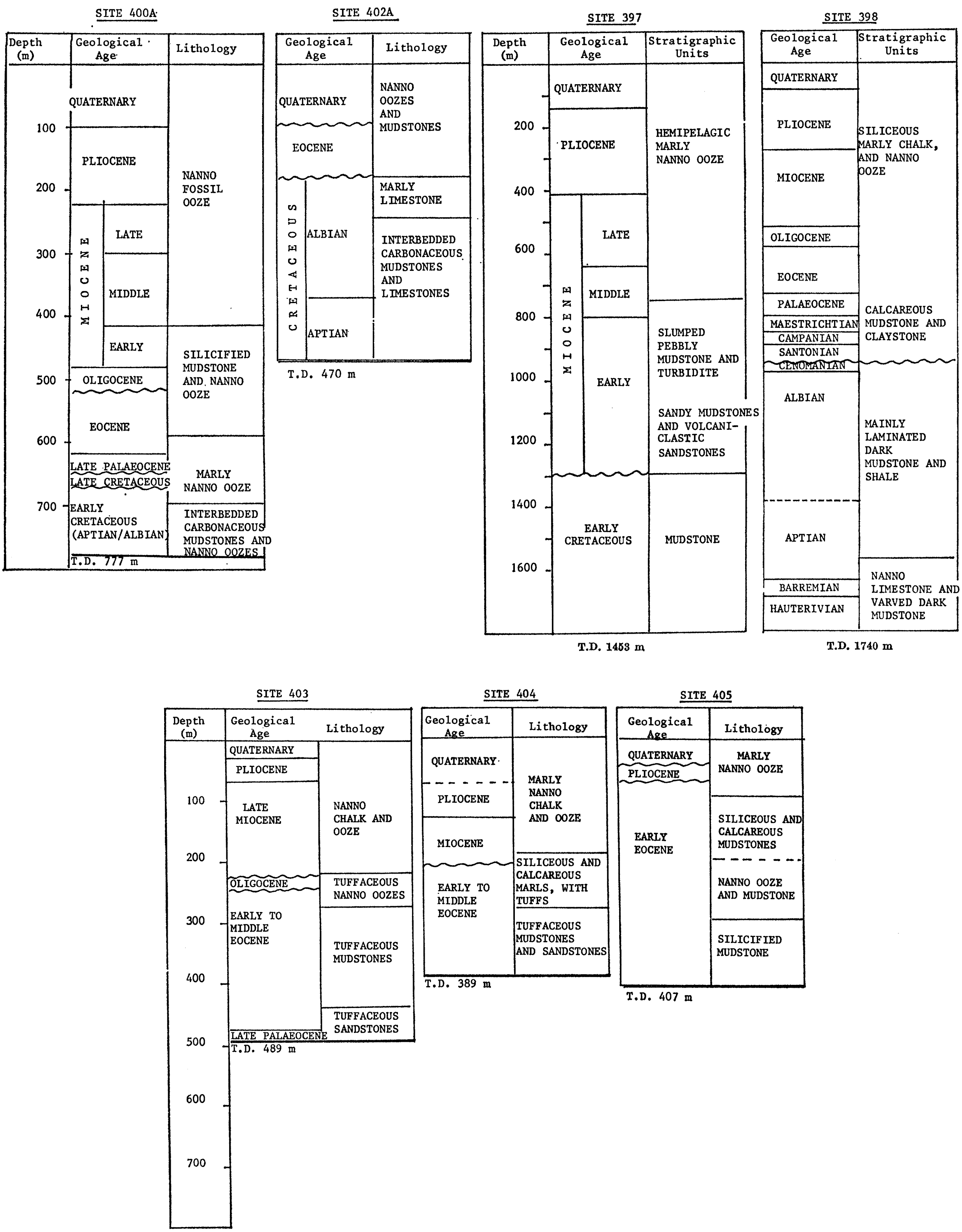

FIGURE 3. Lithological summaries. (T.D. = Total depth.)

demagnetization was used to determine the 'optimum' demagnetizing field, and the remaining samples from that particular unit were then treated at that single alternating field value alone. The results of these stability tests are discussed in detail elsewhere (Hailwood 1979; Hamilton 1979; Morgan 1979). This paper is concerned principally with the interpretation of the magnetic polarity results and their usefulness in dating certain important geological events recorded in the sedimentary sequences.

3. Quaternary and late Neogene magnetostratigraphy

Continuous and undisturbed Quaternary and late Neogene sections of sufficient length to allow the identification of geologically useful magnetic polarity sequences were obtained from Site 397 on Leg 47a, and Site 400A on Leg 48.

(a) *Site* 400*A*

(i) *General polarity sequences*

At this site, located at the northeastern margin of the Bay of Biscay, a fairly complete sequence of undeformed marls and nanno-chalk oozes of Pleistocene, Pliocene and late Miocene age was recovered from the depth range 70–310 m below sea floor (figure 3). One palaeomagnetic sample was taken from each 1.5 m core section recovered, yielding a total of 80 samples. The intensity of magnetization of these samples was very weak, typically in the range of 0.04–0.20 μG with 50% of the values weaker than 0.10 μG. Since the instrument noise level was typically 0.03 μG, the directions of magnetization are frequently subject to an uncertainty of 10° or more. None the less, in most cases partial a.f. demagnetization at several different applied field values in the range 25–300 Oe allowed an unambiguous determination of polarity to be made.

The downhole variation of stable magnetic inclination for these sediments is shown in figure 4, and a number of clearly defined magnetic reversals may be discerned on the basis of changes in sign of inclination. With a few exceptions, each of the major polarity zones are defined by several samples of the same polarity. A comparison between the Site 400A late Neogene polarity sequence and the 'standard' polarity timescale shown in figure 1 reveals a convincing degree of correlation. On the basis of micropalaeontological studies, the uppermost 110 m of sediment at Site 400A are known to be of Pleistocene age (figure 3), and the dominantly reversed polarity correlates well with the upper part of the Matuyama reversed polarity epoch (figure 1). This dominantly reversed polarity sequence at Site 400A extends to a depth of 125 m, and is underlain by a dominantly normal polarity sequence, extending down to 163 m. By comparison with figure 1, it is inferred that the base of the Matuyama reversed polarity epoch, corresponding with a radiometric age of 2.4 Ma, occurs at a depth of 125 m, and the base of the Gauss normal polarity epoch, representing an age of 3.3 Ma, lies at 163 m.

Coring gaps in the underlying 60 m of sediment render correlation with the standard polarity time scale more subjective, but four thin normal polarity zones in the depth range 195–215 m sub-bottom appear to correspond with the four short normal polarity events within the Gilbert reversed polarity epoch.

With these calibration points, the Site 400a magnetic polarity sequence is plotted against the 'standard' polarity timescale in figure 5 (*b*), and a tentative attempt is made to extend the correlation to include magnetic polarity epochs 5–9.

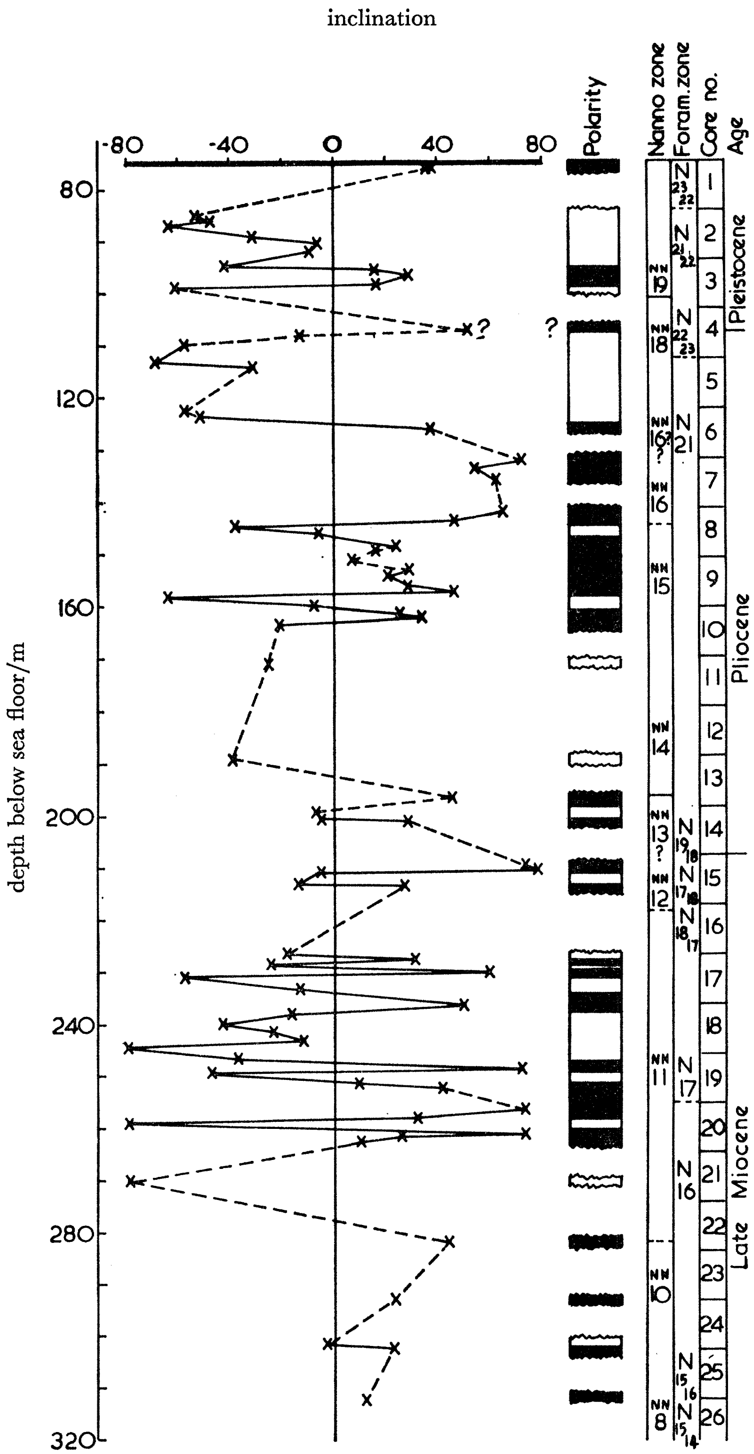

FIGURE 4. Downhole variation of stable magnetic inclination, inferred polarity and biostratigraphic zones for late Neogene sediments at Site 400A. Normal polarity: black; reversed polarity: white.

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(ii) *The Olduvai event and the Plio–Pleistocene boundary at Site* 400*A*

Two apparently separate thin normal polarity zones are present in the depth range 95–110 m at Site 400a, within the sequence assigned to the Matuyama reverse epoch. The upper of these is well defined on the basis of three separate samples, and appears to correspond well with the position of the Olduvai event (figure 5*b*). The underlying thin normal polarity zone at 107 m sub-bottom is based on a single sample with fairly low magnetic stability, but it is possible that this represents one of the short 'Reunion' events which may precede the Olduvai event

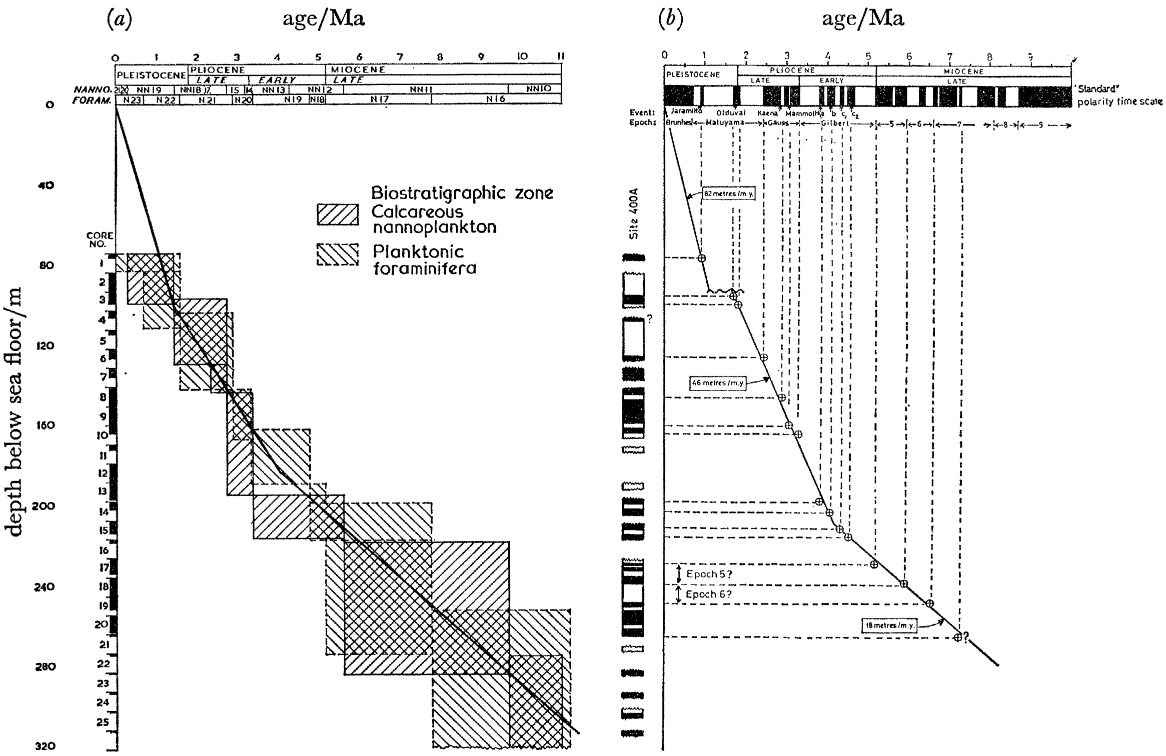

FIGURE 5. (*a*) Late Miocene to Pleistocene biostratigraphic summary for Site 400A. Horizontal sides of boxes represent age range of successive biostratigraphic zones, and vertical sides represent depth extent of these zones at Site 400A. (*b*) Proposed correlation between magnetic polarity record at Site 400A and the 'standard' polarity timescale of figure 1. Note how this correlation allows detection of the early Pleistocene hiatus, which could not be recognized from the biostratigraphic information in (*a*) alone.

(Grommé & Hay 1971). On the basis of combined palaeomagnetic and biostratigraphic studies of stratotype sections in Europe, Ryan *et al.* (1974) propose that the beginning of the Olduvai event correlates with the Plio–Pleistocene boundary, at about 1.8 Ma B.P. The corresponding palaeomagnetic identification of this boundary at Site 400A agrees well with its palaeontological identification at about 107 m sub-bottom (figure 4).

A thin normal polarity zone at 75 m sub-bottom appears to correlate with the Jaramillo event (about 1 Ma B.P.), and, assuming a constant sedimentation rate through to the present time, the palaeomagnetic data imply the existence of a short, but significant hiatus close to the Plio–Pleistocene boundary (figure 5*b*).

The micropalaeontological age determinations for the late Neogene sediments at Site 400A are plotted in figure 5(*a*). The sides of the 'boxes' in this figure represent the age range of each successive biostratigraphic zone, and the corresponding depth range of the microfossil assemblages at Site 400A. It is clear from figure 5 that the palaeontological and palaeomagnetic age determinations at this site are broadly compatible with each other, but that the resolution of the palaeomagnetic data is significantly greater. This allows a more precise determination of sediment accumulation rates, and a better definition of the times at which significant changes in sedimentation rate occurred. Furthermore the palaeomagnetic data clearly define the existence of the early Pleistocene hiatus, which could not have been identified from the available palaeontological information alone.

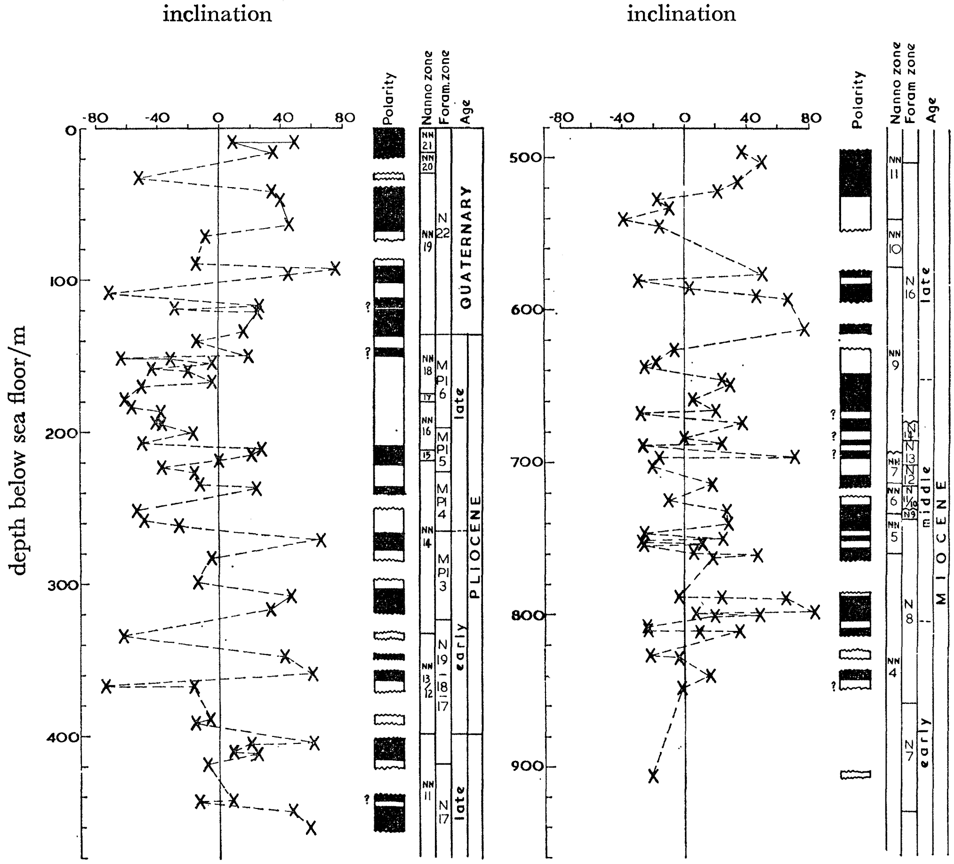

FIGURE 6. Palaeomagnetic and biostratigraphic summary for late Neogene sediments at Site 397. Symbols as in figure 4.

(*b*) *Site* 397

This site is located off the coast of northwest Africa (figure 2) in a region which experienced persistently high sediment accumulation rates throughout the entire Neogene. Consequently the total thickness of Neogene sediments is nearly three times greater than that at Site 400A. At this site a very complete Quaternary to early Miocene section was recovered and, as at Site 400A, the most convincing correlation with the 'standard' polarity reversal timescale exists for the late Pliocene and Pleistocene part of the polarity record (figure 7).

An interval of dominantly reversed polarity occurs between a depth of 70 m, and the inferred top of the Gauss normal epoch, at 208 m sub-bottom, and is assigned to the Matuyama reversed epoch. The palaeontologically determined Plio–Pleistocene boundary, at 140 m sub-bottom, corresponds closely with the start of a well defined short normal polarity interval within the Matuyama epoch sediments, and as at Site 400A (§3*a*), this interval is taken to represent the Olduvai event. Also in agreement with Site 400A, a separate and distinct short normal polarity interval (located at 140 m sub-bottom) appears to precede the Olduvai event at this site.

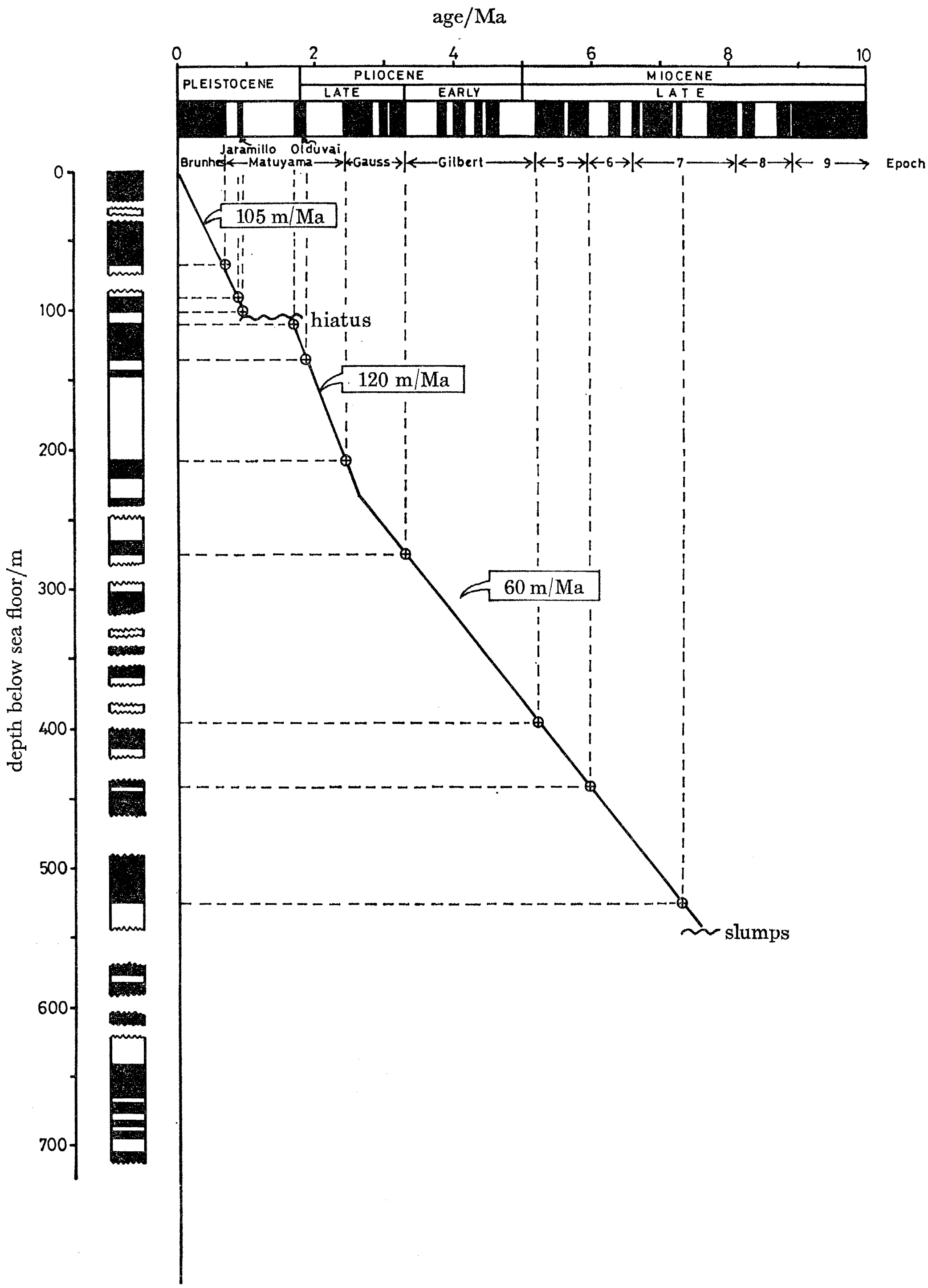

FIGURE 7. Proposed correlation of Neogene magnetic polarity record at Site 397 with polarity time scale of figure 1. Note inferred presence of early Pleistocene hiatus, as at Site 400A.

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Early Pleistocene hiatus

At Site 397, the Olduvai is followed by a further normal polarity event within the Matuyama epoch sediments at a depth of about 100 m, which correlates well with the Jaramillo event. According to this interpretation a significant sedimentary hiatus occurs close to the Plio–Pleistocene boundary, in an identical position to that proposed on the basis of palaeomagnetic evidence at Site 400A.

It is important to note the absence of direct palaeontological evidence for such an hiatus, since all standard late Pliocene and Pleistocene calcareous nannoplankton and planktonic foraminifera zones appear to be present at both sites (figures 4 and 6). Nevertheless, at Site 397 there are other independent lines of evidence supporting the possible existence of such an hiatus (Leg 47a Initial Report). Since the proposed hiatus lies close to the NN18–19 nannofossil zone boundary, and the N21–22 foraminiferal zone boundary, it could be represented simply by a shortening of these zones. It is possible that a re-examination of the nannofossils by the refined Quaternary zonation scheme of Gartner (1977) could allow its palaeontological recognition. The identification of an early Pleistocene hiatus at both Site 397 (northwestern African continental margin) and Site 400A (Bay of Biscay) suggests the occurrence of a contemporaneous oceanographic event whose influence may have been sufficiently widespread to affect much of the eastern Atlantic, and was probably associated with the Pleistocene glaciation.

4. Early Palaeogene magnetostratigraphy

Very complete sedimentary sections of late Palaeocene to early Eocene age were drilled at IPOD Sites 403, 404 and 405, near the southwestern extremity of the Rockall plateau. The sediments at Sites 403 and 404 are dominated by volcanogenic siltstones and glauconitic mudstones, and were deposited in a deltaic environment during the initial stages of rifting and separation of Rockall from East Greenland. The patterns of oceanic magnetic anomalies formed during the early stages of sea floor spreading in this region are well preserved, and individual anomalies are readily identifiable. Consequently, it should be possible to recognize similar patterns of polarity reversals, recorded vertically in the sedimentary sequences at IPOD Sites 403–405, and simultaneously in the horizontal magnetic anomalies of the underlying oceanic lithosphere. A correlation between these two polarity reversal sequences would provide an important opportunity for using the palaeontological dating of the reversals in the sedimentary section to assign an age to the oldest marine magnetic anomaly (number 24), and so determine the date of initiation of sea floor spreading between Rockall and Greenland.

The magnetic polarity records derived from the sediments at Sites 403–405 are summarized in figure 8. At Sites 403 and 404 a thick reversed polarity sequence was encountered in the lower parts of the holes, and this has been assigned to the late Palaeocene and early Eocene nannofossil zones NP10 and NP11 (D.S.D.P. Leg 48 Initial Reports). The total interval of time represented by these reversely magnetized sediments is some 2.3 Ma and the only reversed polarity period of this length close to the Palaeocene–Eocene boundary is the long reverse interval that immediately preceded anomaly 24. Consequently, the first normal polarity sediments to occur above this interval may be assigned with some confidence to anomaly 24. These are identified at the base of nannofossil zone NP12 at IPOD Site 404 (figure 8). According to the geological time scale of Hardenbol and Berggren (in press), compiled from a correlation of microfossil zones with radiometrically dated horizons, this nannozone boundary corresponds with a numerical age of 52 Ma, and consequently this may be taken to represent the date of initiation of sea-floor spreading in the North East Atlantic between Greenland and Rockall.

In general, the preservation of planktonic foraminiferal fossils in the Palaeogene sediments of Sites 403–405 is very poor, so that foram zone determinations are only of limited value. However, the nannofossil and dinoflagellate cyst zones are generally well defined (Muller 1979;

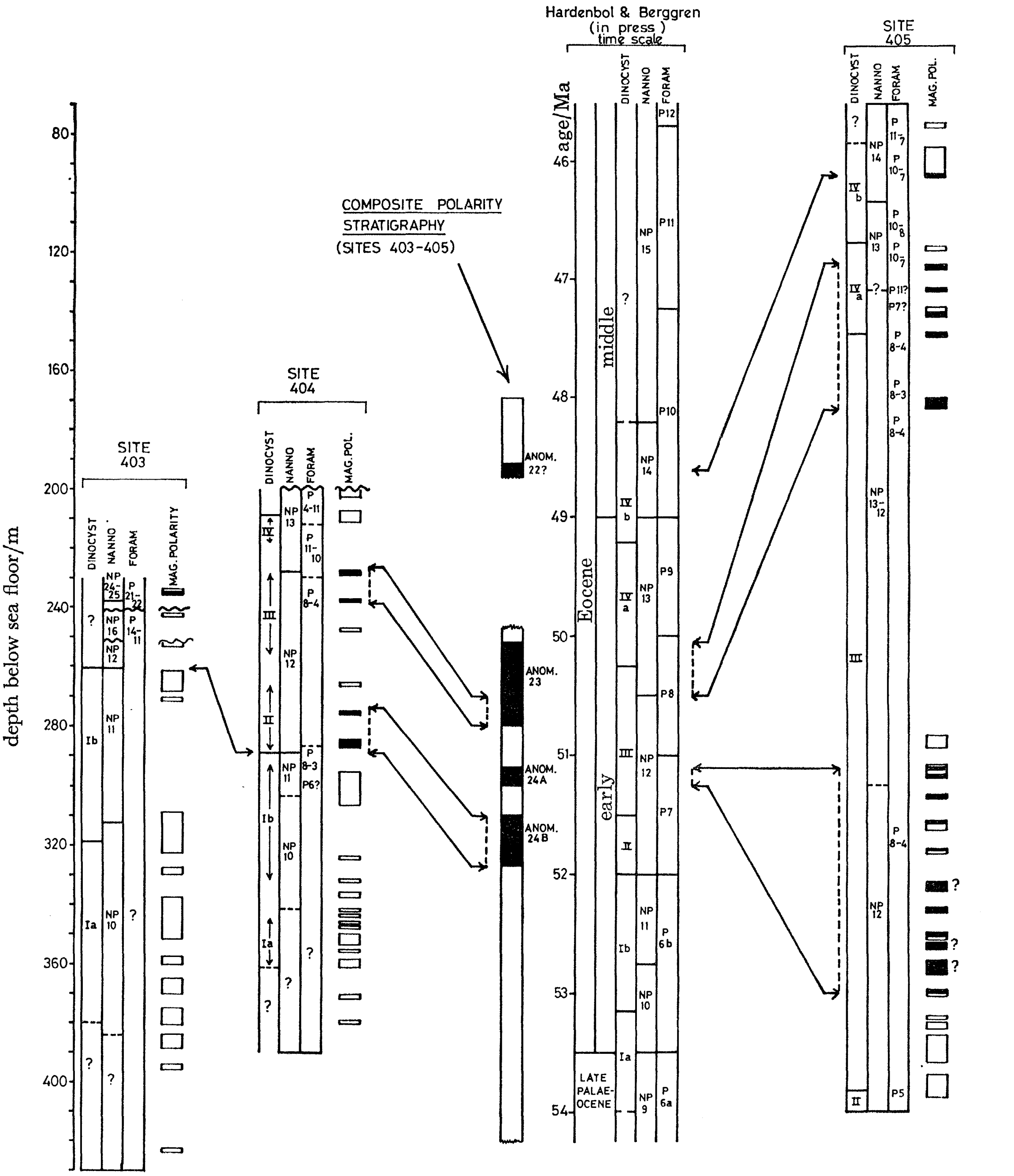

Figure 8. Early Palaeogene magnetic polarity and biostratigraphic summary for Sites 403–405, Rockall Plateau. Composite polarity scale in central column is calibrated via biostratigraphic zone determinations with the Palaeogene timescale of Hardenbol & Berggren (1979).

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Costa & Downie 1979). On the basis of a simultaneous comparison between the magnetic polarity determinations and biostratigraphic zones at the three sites, the composite magnetic polarity stratigraphy shown in the central column of figure 8 may be derived, and an excellent correlation is obtained with the pattern of geomagnetic field reversals recorded as marine magnetic anomalies 24B, 24A, 23, and part of 22. This correlation is supported by the fact that the normal polarity sediments attributed to anomaly 24 at Site 404 may be traced westwards by means of seismic reflexion records, and are observed to overlie immediately the oceanic basement of anomaly 24 age, as deduced from marine magnetic anomalies (D. G. Roberts, personal communication).

The proposed age of 52 Ma for the start of anomaly 24, and hence for the start of opening of the NE Atlantic, is significantly different from the values of 60 Ma proposed by Heirtzler *et al.* (1968), 56 Ma proposed by Sclater *et al.* (1974), and 49 Ma by Tarling & Mitchell (1976). However, it is believed to be more reliable than these previous estimates, since it is based on direct palaeomagnetic and biostratigraphic studies of continuous sedimentary sequences spanning the time interval represented by anomalies 22–24. This proposed age for the start of anomaly 24 is also in agreement with the radiometric and biostratigraphic age of the East Greenland basalts, which were erupted during the long reverse polarity interval prior to anomaly 24 (see discussion in Hailwood *et al.* 1979).

5. Cretaceous magnetostratigraphy

Thick sequences of Cretaceous carbonaceous mudstones, spanning the Aptian and Albian stages, were recovered from Sites 400A and 402A on IPOD Leg 48, and Site 398 on Leg 47b (figure 3). At the latter site, these extend downwards into the late Hauterivian and are unconformably overlain by brown and red chalks and calcareous mudstones of Santonian to Maestrichtian age. At Site 397 on Leg 47a, 150 m of late Hauterivian to early Barremian mudstones were penetrated, beneath the early Miocene unconformity.

The downhole variations in inclination of stable remanence, together with the inferred magnetic polarity, and available micropalaeontological age determinations, are summarized in figures 9 and 10. One of the most conspicuous features of the polarity records for all four sites is the predominance of normal magnetic polarities throughout the Cretaceous sections. This is consistent with their deposition during the long Cretaceous interval of dominantly normal polarity (Helsley & Steiner 1969). The duration of this long normal polarity episode, together with the age of certain short period events within it, is of considerable importance to the interpretation of marine magnetic anomalies within the Cretaceous 'magnetic quiet zones' identified in the oceanic lithosphere, and also to the investigation of Mesozoic seafloor spreading history and geomagnetic field behaviour.

Short period reversals within the Cretaceous normal polarity interval

A number of short period reversed events have been identified within the Cretaceous interval of dominantly normal polarity. Perhaps the best documented of these is the reversed polarity event, of possible duration *ca.* 1 Ma, situated close to the Aptian–Albian boundary (Helsley & Steiner 1969; McElhinny & Burek 1971; Pechersky & Khramov 1973). This reversal has been correlated with marine anomaly 'M–0', and from a study of marine magnetic profiles in the Hawaiian area Larson & Hilde (1975) derive an age of mid- to late Aptian for this anomaly.

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Further evidence in support of this age is provided by the results of D.S.D.P. drilling at Site 417D, situated on top of this anomaly in the North Atlantic. Sediments immediately overlying basement at this site have been dated by nannofossils at late Aptian. In view of this evidence, the age of this anomaly appears to be fairly well established as late Aptian.

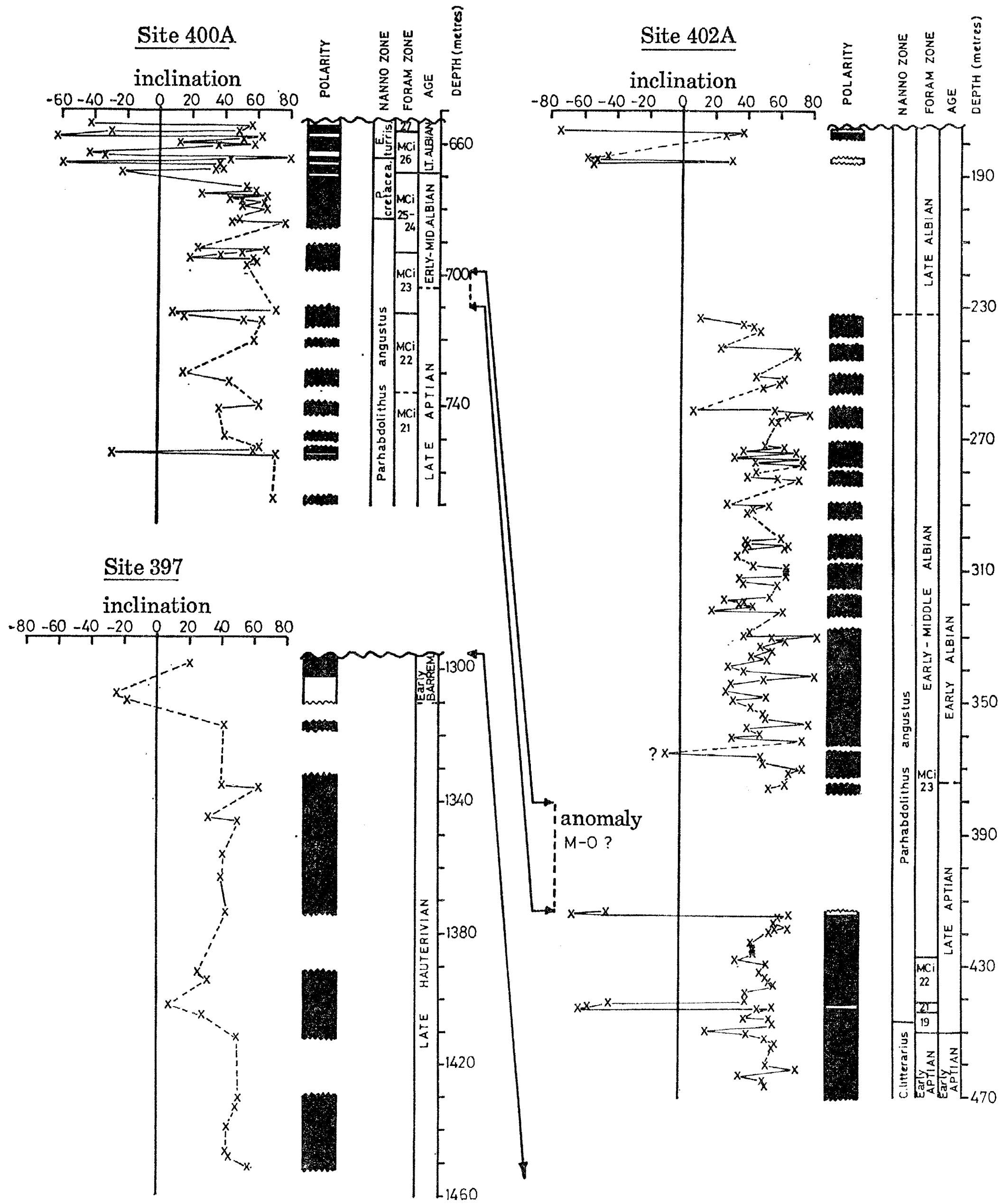

FIGURE 9. Paleomagnetic and biostratigraphic summary for Cretaceous sediments at Sites 397, 400A and 402A.

However, at none of the three sites penetrating Aptian–Albian sediments on Legs 47 and 48 (Sites 398, 400A and 402A) is there any indication of a prolonged reversed polarity event in sediments of this age (figure 9). Careful a.f. demagnetization studies were carried out on all samples from these three sites, and no evidence was found for possible reversed polarity components being masked by a later normal polarity overprint. Accepting the proposed late Aptian age for this anomaly, its absence in the sedimentary record at these three sites may be attributed to a combination of coring gaps and sedimentary hiatuses. At Site 398, a hiatus of uncertain duration has been proposed close to the Aptian–Albian boundary (IPOD Leg 47b Initial Reports 1979). At Site 400a there are several coring gaps representing time intervals of approximately 1 Ma, but at Site 402A, where the sedimentation rate was nearly three times greater, the only coring gap of sufficient length occurs immediately below the Aptian–Albian boundary, within the lower part of the MC 23 (Tricinella bejaouaensis) foram zone of Sigal (1977). Consequently the age of anomaly M–0 may be deduced by default to correspond with this biostratigraphic zone.

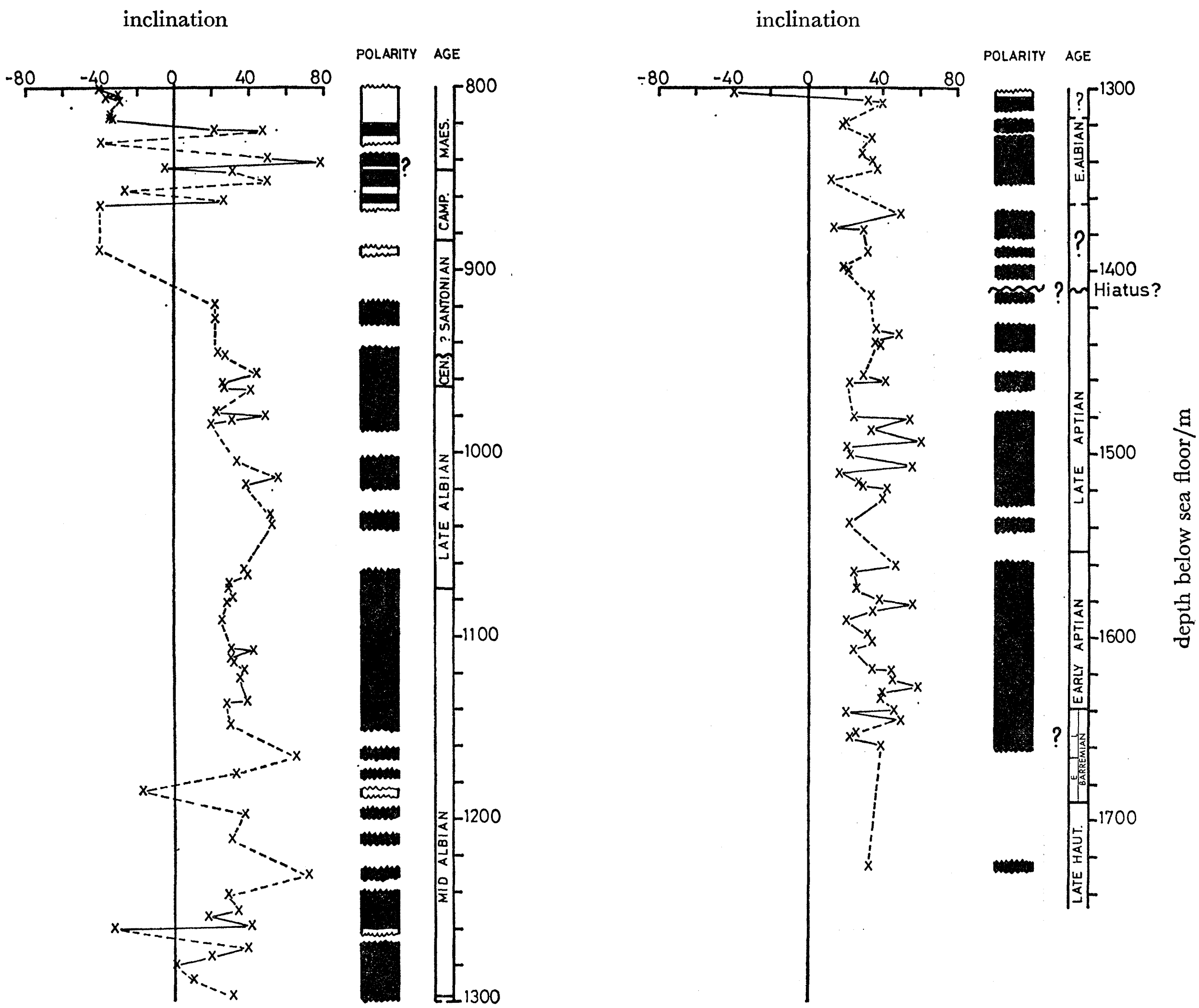

FIGURE 10. Palaeomagnetic and biostratigraphic summary for Cretaceous sediments at Site 398.

Although no prolonged reversed event was observed in the late Aptian sediments, a conspicuous sequence of short-period reversals was identified in the well dated late Albian sediments at Sites 400A and 402A (figure 9). These appear to correlate well with a similar 'mixed polarity zone' identified at D.S.D.P. Site 263, off Western Australia (Green & Brecher 1974; Jarrard

1974). This late Albian mixed polarity zone was not identified at Site 398 (figure 10), but may possibly correspond with the coring gaps between 980 and 1060 m sub-bottom.

Length of Cretaceous normal polarity interval

The Cretaceous polarity determinations from Legs 47 and 48 place some additional constraints on the maximum length of the long Cretaceous normal polarity interval. Apart from the short isolated reversals during the Aptian and Albian discussed above (figure 9), the oldest reversed sample found above the long Cretaceous normal zone occurs in late Santonian sediments at Site 398 (figure 10). This is consistent with the results of Alvarez *et al.* (1977) who place the end of the long normal interval at the Santonian–Campanian boundary (figure 12).

Several alternative ages have been proposed for the beginning of the Cretaceous normal interval, which corresponds with the end of magnetic anomaly M–1. For example, Larson & Pitman (1972) place the base of this normal interval just above the Aptian–Barremian boundary, whereas Larson & Hilde (1975) place it close to this boundary, but within the late Barremian

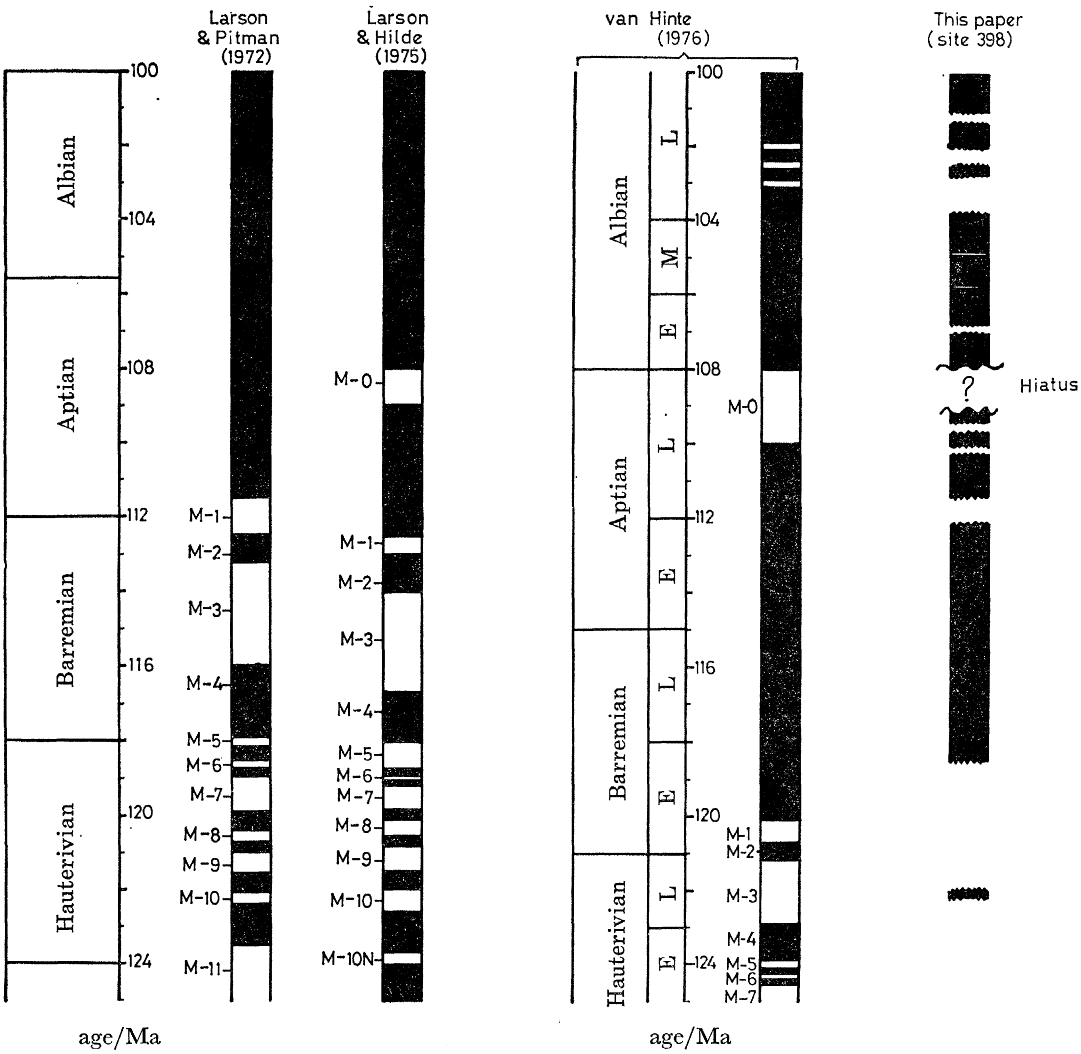

FIGURE 11. Comparison of Site 398 early Cretaceous magnetic polarity record with different versions of the early Cretaceous polarity timescale.

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(figure 11). Van Hinte (1976) suggests a downward revision of 10 Ma so that the start of the Cretaceous normal interval lies close to the base of the Barremian.

The magnetic polarity results from Site 398 allow a test of these alternative interpretations to be made. At this site a consistently normal stable magnetization was observed down to a depth of 1660 m (uppermost lower Barremian, figure 10). As indicated in figure 11, these results are more consistent with the proposed location of anomaly M–1 in the early Barremian

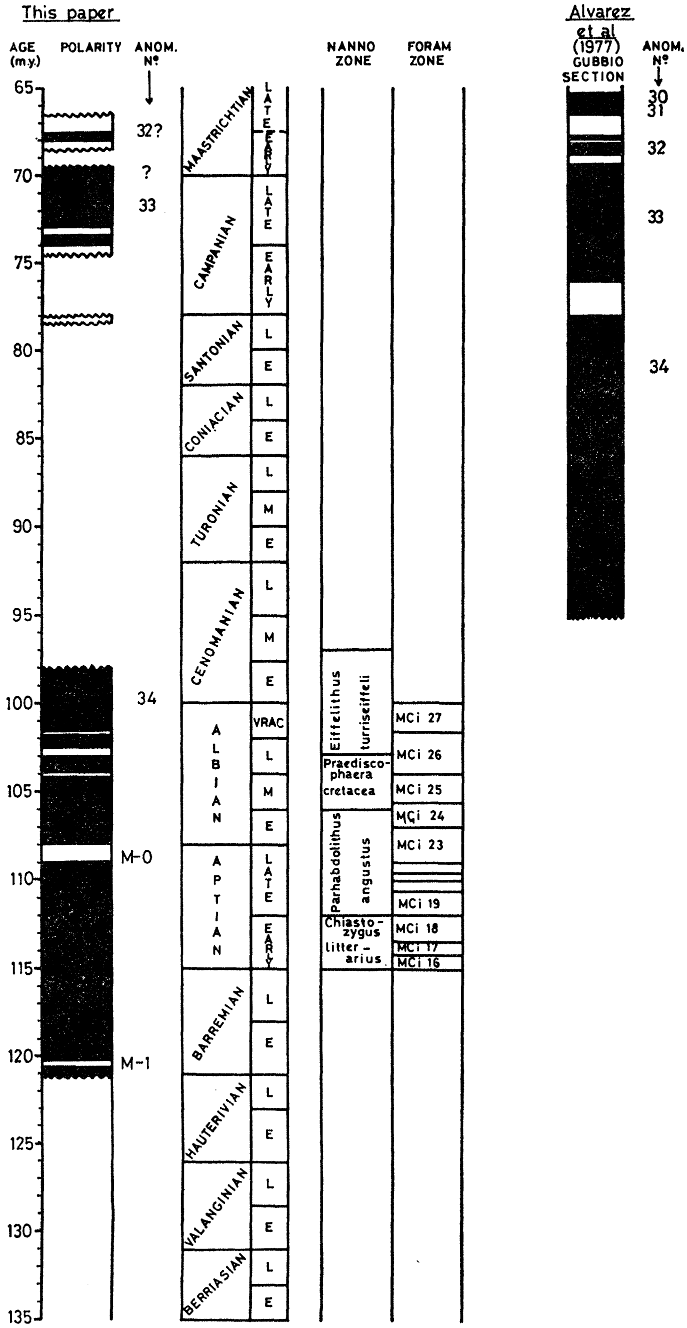

FIGURE 12. Summary of Cretaceous polarity determinations at Sites 397, 398, 400A and 402A discussed in this paper, and comparison of late Cretaceous sequence with polarity timescale of Alvarez *et al.* (1977).

(van Hinte 1976) than in the late Barremian position proposed by Larson & Pitman (1972) and Larson & Hilde (1975).

At Site 397, a thick sequence of early Cretaceous normal polarity sediments, tentatively assigned to the late Hauterivian, is overlain by a relatively thin layer of reverse polarity sediments assigned to the early Barremian (figure 9). This situation cannot be matched on any of the early Cretaceous polarity timescales so far proposed (figure 11), since in all three cases the late Hauterivian is dominantly reversed. Accepting the evidence presented above in support of the van Hinte (1976) early Cretaceous polarity timescale, it is quite possible that the early Barremian thin reverse polarity zone at Site 397 represents anomaly M–1, but it is extremely unlikely that the underlying thick normal polarity sequence represents the dominantly normal period between anomalies M–1 and M–3, since this would require an unrealistically high sediment accumulation rate of some 100 m/Ma. An alternative, and perhaps more likely, interpretation is the existence of a short hiatus at this site representing all, or most of, anomaly M–3. The thick Cretaceous normal polarity sequence at Site 397 would then represent the dominantly normal period between anomalies M–7 and M–3, possibly also including much of the normal period between M–3 and M–1. This interpretation requires a more realistic sedimentation rate of some 45–50 m/Ma.

A summary of the Cretaceous magnetic polarity stratigraphy for the four D.S.D.P. sites discussed in this paper is presented in figure 12.

6. Conclusions

The application of magnetic polarity stratigraphy to the dating of deep sea sediments is still in its infancy, but the examples from Ipod Legs 47 and 48 presented in this paper serve to demonstrate a range of the potential applications of this subject. It has been demonstrated that the sedimentary sections of late Neogene age from Site 397, off the coast of northwest Africa, and Site 400A in the northeast Bay of Biscay, have recorded a pattern of magnetic polarity reversals that correlates well with the known magnetic polarity time scale for the past 5–10 Ma, and allows the determination of precise sediment accumulation rates, and the recognition of a short but significant hiatus in the early Pleistocene, at both sites. This hiatus must represent an oceanographic event which affected a significant part of the eastern Atlantic continental margins, but because of the limited resolution of available micropalaeontological zonations, it could not have been detected by biostratigraphic observations alone.

The sequence of rapidly deposited deltaic sediments of early Palaeogene age, cored at Sites 403–405 near the southwestern extremity of the Rockall Plateau, have recorded a pattern of magnetic reversals which show a good correlation with marine anomalies 22–24. The high-quality nannofossil and dinocyst age determinations from these cores allow the confident assignment of the start of anomaly 24 to the base of the early Eocene mannofossil zone NP12. Since this anomaly is the oldest identifiable marine anomaly in the North East Atlantic between Greenland and Rockall, and occurs close to the continental margins, the corresponding oceanic lithosphere must have been formed during the very early stages of separation of Rockall from Greenland. Consequently, the combination of palaeomagnetic and micropalaeontological studies of sediments from Sites 403–405 allows the accurate determination of the time of initial rifting of Greenland from the Rockall Plateau and Scandinavia.

Finally, a sequence of magnetic reversals has been identified in sediments of Cretaceous age

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at Sites 397, 398, 400A and 402A, and the biostratigraphic age determinations on these sediments provide useful constraints on the duration of the long Cretaceous interval of normal polarity, and on the age of anomalies M–0 and M–1. This information has direct relevance to the interpretation of Mesozoic marine magnetic anomaly patterns in terms of the history of seafloor spreading during the early stages of opening of the South Atlantic ocean basin.

It is clear from the above examples that the combination of palaeomagnetic and biostratigraphic studies on the same deep sea sediment cores can provide important information that has considerable relevance to understanding the evolution of continental margins. The installation of a palaeomagnetic laboratory on board D/V *Glomar Challenger* at the start of Leg 47 greatly facilitated these studies, and in future years should contribute significantly to ascertaining the geological age of other important tectonic events, and relating these to the overall history of evolution of the ocean basins and their margins.

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Discussion

E. J. W. Jones (*Department of Geology, University College London, Gower Street, London, WC1E 6BT*). Is there any evidence to suggest that parts of the sedimentary sequences acquired their present magnetization during diagenesis rather than at the time of deposition?

E. A. Hailwood. The precise time at which sediments acquire their magnetization is always difficult to evaluate, and at the present time there is no universally applicable test that allows a distinction to be made between depositional remanent magnetizations (d.r.ms) acquired at the time of deposition and chemical remanent magnetizations (c.r.ms) acquired as a result of the growth of new magnetic minerals during diagenesis. Unfortunately, both types of magnetization have similar stabilities, so that the standard laboratory demagnetization procedures applied to all samples discussed in this paper, in order to remove possible low stability components acquired during drilling and storage, cannot normally distinguish whether the underlying stable component is a d.r.m. or a c.r.m. However there is some indirect evidence that for most of the samples studied the stable component of magnetization is a d.r.m. since the shape of thermomagnetic curves for samples from Leg 48, and the typical intensities of magnetization of sediments from all three legs suggest that the carriers of magnetic remanence are usually titanomagnetite grains, rather than haematite grains. It is widely believed that titanomagnetities in marine sediments are normally of detrital, rather than chemical, origin since chemically formed iron oxides normally result from the initial precipitation of iron hydroxides, which are later dehydrated to produce haematite rather than magnetite. Thus, as a general rule, c.r.ms in marine sediments tend to reside in haematite grains, which, so far have not been identified in these sediments. None the less, there are mechanisms by which magnetite can sometimes be precipitated in marine sediments, so that the possibilities of the magnetite grains carrying a c.r.m. cannot be completely ruled out. However, we regard this as unlikely in these particular sediments.

Further evidence that the magnetization of these sediments was acquired close to their time of deposition is the fact that for the Quaternary, late Neogene and early Palaeogene sediments, patterns of magnetic reversals have been identified, which can be matched with the established polarity time scale. If the magnetization had been introduced as a result of the variable processes of diagenesis, at significantly different times from the initial deposition, such a close match would be unlikely. Similarly, with the Cretaceous sediments, the recording of certain reversals, such as those in the late Albian 'mixed polarity zone' at Sites 400A and 402A (figure 9) suggests that these particular sediments acquired their magnetization close to the time of deposition. Since the underlying Aptian sediments have a similar lithology, there seems no reason to suppose that diagenesis produced a preferential remagnetization of the Aptian sediments without affecting those of Albian age. We therefore favour the interpretation that the magnetization of these sediments is a primary d.r.m. acquired during deposition, rather than a secondary c.r.m. acquired during diagenesis. This interpretation is supported by the fact that the mean palaeolatitudes deduced for the Site 400A and 402A early Cretaceous sediments, from their mean palaeomagnetic inclinations, are identical with the early Cretaceous palaeolatitude inferred for these sites from continental palaeomagnetic data (Hailwood, in D.S.D.P. Leg 48 Initial Reports). At Site 398 there is some indication of a systematic lowering of the inclination values as a result of compaction effects, (Morgan, in D.S.D.P. Leg 47b Initial Reports) so that some small post-depositional modification of the remanence directions may have occurred at this site, but there is no evidence for the substantial modifications that would be required, in order to influence stratigraphy presented in this paper.