Jurassic to Paleogene: Part 2 Paleogene geochronology and chronostratigraphy

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Jurassic to Paleogene: Part 2
Paleogene geochronology and chronostratigraphy

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SUMMARY: We present a revised Paleogene geochronology based upon a best fit to selected high temperature radiometric dates on a number of identified magnetic polarity chrons (within the late Cretaceous, Paleogene and Neogene) which minimizes apparent accelerations in sea-floor spreading. An assessment of first order correlations of calcareous plankton biostratigraphic datum events to magnetic polarity stratigraphy yields the following estimated magnetobiochronology of major chronostratigraphic boundaries: Cretaceous–Tertiary boundary (Chron C29R), 66.4 Ma; Paleocene–Eocene (Chron C24R), 57.8 Ma; Eocene–Oligocene (Chron C13R), 36.6 Ma; Oligocene–Miocene (Chron C6CN), 23.7 Ma.

The Eocene is seen to have expanded chronologically (~ 21 m.y.) at the expense of the Paleocene (~ 9 m.y.) and is indeed the longest of the Cenozoic epochs. In addition, magnetobiostratigraphic correlations require adjustments in apparent correlations with standard marine stage boundaries in some cases (particularly in the Oligocene). Finally, we present a correlation between standard Paleogene marine and terrestrial stratigraphies.

It is nearly 20 years since Brian Funnell prepared the first relatively precise Cenozoic time-scale based on an assessment of palaeontologically controlled radiometric data in connection with the symposium on the Phanerozoic time-scale sponsored by the London Geological Society, and 10 years since one of us (WAB) presented the first in a series of attempts to further refine Cenozoic geochronology. During the past decade several revisions to the Cenozoic time-scale have appeared and here, at this, the second symposium on the Phanerozoic time-scale sponsored by the Geological Society of London, it is appropriate to present an updated and, hopefully, improved version of the Cenozoic time-scale.

It is opportune that over the past decade direct correlation has been achieved between plankton biostratigraphy in some of the standard European continental marine sections and North American terrestrial vertebrate biochronology and magnetic polarity stratigraphy over much of the Cenozoic Era. The recent improvement in deep sea coring techniques has further extended these correlations on a global scale. It is now possible to make age estimates of epoch boundaries and the extent of time-stratigraphic (standard ages) units in terms of plankton biostratigraphy and magnetic polarity chrons and/or anomalies.

Finally a critical evaluation must be made within a geohistorical context of biostratigraphically controlled radiometric dates and radiometrically dated polarity stratigraphy in order to provide constraints on an internally consistent geologic time-scale.

The revised Cenozoic geochronology has been prepared in two parts: (a) Paleogene; (b) Neogene. In this paper dealing with the Paleogene we first discuss the development of geomagnetic polarity history of the late Cretaceous and Cenozoic. A revised geochronology is then presented which is based upon a best fit to selected high temperature radiometric dates on a number of identified magnetic polarity chrons (in the late Neogene, early Oligocene, middle Eocene, and late Cretaceous) which minimizes apparent acceleration in sea-floor spreading. This is followed by a discussion of the biostratigraphy of the major Paleogene epochs and their boundaries beginning with the Cretaceous–Tertiary boundary. Our revised Paleogene geochronology is presented in a series of figures and reflects our assessment of presently available data from the fields of magneto- and biostratigraphy and radiochronology. The magnetobiochronology of the calcareous plankton (and by extension, the age estimate of the standard epoch and age boundaries) is based on a compilation of first order correlations between biostratigraphic datum levels and magnetic stratigraphy in continental, marine, and deep sea core material. These data are present in tabular form in the appendix.

Paleogene geomagnetic polarity time-scale

The basis for a geomagnetic polarity reversal chronology for the late Jurassic to Recent is the polarity sequence inferred from analysis of marine magnetic anomalies. Although the Paleogene portion of geomagnetic reversal history is of interest here, it is best considered in the context of the magnetic anomaly sequence extending from the present sea-floor spreading axis to the younger limit of the Cretaceous Long Normal or Quiet Zone. Because of the lack of correlatable features in the Cretaceous Quiet Zone, the older (late Jurassic and early Cretaceous) set of anomalies, referred to as the M-sequence (Larson & Hilde 1975), can be treated separately.

The first extended geomagnetic reversal time-scale was presented by Heirtzler et al. (1968) who chose a magnetic profile from the South Atlantic Ocean as representative of geomagnetic reversal history for about the past 80 Ma. Their chronology, hereafter referred to as HDHPL68, was derived by a correlation of the axial anomalies to the 0 to 4 Ma, radiometrically-dated magnetic reversal time-scale (Cox et al. 1965) and by extrapolation to the oldest then recognized polarity interval (anomaly 32). This twenty-plus fold extrapolation assumed that the rate of sea-floor spreading in this area of the South Atlantic was constant over about 1400 km or 80 Ma, at the value calculated from 0 to 3.35 Ma (anomaly 2A). Despite the severe extrapolation required, HDHPL68 has proved its utility in description of sea-floor spreading histories in the world ocean and continues in large part to be the basis for all subsequent revised geomagnetic reversal time-scales (see review by Ness et al. 1980). It is now apparent that HDHPL68 generally comes within 10% of currently accepted ages for this reversal sequence, a remarkable achievement and an indication that the assumption of

sea-floor spreading at a constant rate over prolonged time intervals is a valid approximation.

Although recent magnetostratigraphic investigations have identified large portions of essentially the same magnetic reversal pattern in marine sedimentary sections (e.g. Lowrie *et al.* 1982; Poore *et al.* 1982) and in volcanic sequences with radiometric-date control (McDougall *et al.* 1976), the marine magnetic anomaly record continues to be the standard for determining the relative position of polarity intervals and hence for correlation. This is largely due to the great wealth of marine magnetic anomaly data which can be used to demonstrate that the interpreted record of geomagnetic reversals is relatively smooth and continuous, that is, the same sequence of anomalies (polarity reversals) can be found everywhere, differing over appreciable intervals only by some proportionality factor that reflects formation at different spreading rates. The large number of profiles available also makes possible averaging or stacking of profiles to reduce noise, resulting in a better representation of the true geomagnetic reversal sequence. In contrast, there are few long magnetostratigraphic sections to adequately allow separation of changes in accumulation rates from differences in duration of polarity intervals. There is also the greater probability that sea-floor spreading, on the scale that effects the magnetic anomaly signature, proceeds more regularly over longer time intervals compared to the often cyclic or episodic nature of sediment or lava accumulation. Finally, magnetic anomalies represent an average of the magnetization over substantial portions of oceanic crust and consequently are less likely to reflect small-scale, local variabilities in the recording mechanism than in the discrete sampling in a magneto-stratigraphic study.

For these reasons, the revised magnetic polarity time-scale presented here relies for its continuity and basic structure on the inferred nature of sea-floor spreading history in the world ocean. In particular, we attempt to avoid modifications to the time-scale that would introduce changes in sea-floor spreading rates which are not supported by tectonic or other geological or geophysical evidence.

Use of the marine magnetic anomaly record for the construction of a magnetic reversal time-scale does, however, present the problem of absolute date control since few reliable radiometric-date determinations are available from the sea floor that can be used for direct calibration. Instead, it is necessary to calibrate the magnetic anomaly indirectly, by correlation, often tiered, to relevant material dated elsewhere. Initially, correlation to the 0–4 Ma radiometrically-dated magnetic reversal time-scale was used (e.g. HDHPL68). Unfortunately, it has not proved possible to extend the radiometric-reversal time-scale much beyond present limits of 4–5 Ma, since the usual errors of a few percent in an age determination soon become comparable to the separation of one polarity interval from the next closest one of the same polarity. Since like polarity intervals are distinguishable only by their relative duration within a characteristic pattern of reversals with time, further extension of the radiometrically dated reversal time-scale using an accumulation of radiometric date-magnetization polarity determinations on unrelated lavas is not likely with present radiometric dating methods (Cox & Dalrymple 1967).

Magnetostratigraphic studies provide an additional source of age information that depends on correlation of the measured magnetic polarity zones in a section to the geomagnetic reversal sequence derived from magnetic anomalies; any age-diagnostic property in the section can then be potentially used for calibration. Radiometric dates are sometimes available from the same section investigated for magnetostratigraphy. A notable example is the work on Icelandic lavas (McDougall *et al.* 1976) where it has been possible to directly estimate ages of polarity reversal levels from the stratigraphic distribution of numerous radiometric dates. Much more commonly, however, sedimentary sections which have not been dated directly are studied and numerical age control is derived by biostratigraphic correlation to a geologic time-scale. The accuracy of such ages depends on both the precision of the correlation and the quality of the age estimates for the standard geological stage boundaries. An appraisal of such correlations and age estimates for the Paleogene is presented elsewhere in this paper.

Given a set of ages tied by various correlations to the standard magnetic reversal sequence, several approaches can be used to calibrate it. One method is to fix one or more points in the polarity reversal sequence to the corresponding age estimates obtained by correlation and calculate the ages of other reversals by interpolation or extrapolation. Besides the origin, only a single calibration point was used in HDHPL68, whereas in the time-scale of LaBrecque *et al.* (1977) (hereafter referred to as LKC77), an additional calibration point was added just below (older than) anomaly 29, a position correlated with the Cretaceous–Tertiary boundary (about 65 Ma) by Lowrie & Alvarez (1977).

As more extensive magnetobiostratigraphic correlations become available, further calibration tie points can be fixed. For example, Lowrie & Alvarez (1981) fixed the ages of nine points in the late Cretaceous to Oligocene-Miocene portion of the geomagnetic reversal sequence on the basis of magnetobiostratigraphic correlations in Italian limestones. Such stringent use of calibration tie-points, however, increases the possibility of introducing as artifacts apparent accelerations in sea-floor spreading as the number of calibration tie-points increases within a finite time interval. This is apt to occur because the inherent errors in the age estimates of the calibration points become more important in calculating interval spreading rates as the calibration tie-points used in this way become more closely spaced in time.

An alternative method which we employ here is to assume a minimum number of changes in sea-floor spreading rates that will still satisfy the constraints of the calibration tie-points. Linear segments, each encompassing significant portions of the magnetic reversal sequence are thus identified and a chronology is determined by linear regression analysis. The same age calibration data used by Lowrie & Alvarez (1982) can be analysed in this fashion although many of their Paleogene stage boundary age estimates require revision as discussed elsewhere in this paper. Moreover, we have tried to refrain wherever possible from directly incorporating the age estimates for geological stage boundaries in calibrating the geomagnetic reversal sequence in an effort to produce an independently derived chronology for comparison. As will be shown, most stage subdivisions of the Paleogene and the late Cretaceous are well correlated with the magnetic reversal sequence and it would therefore be of interest to see how well age estimates based, at least in part, on different techniques and assumptions compared. It was in fact because of such a comparison between LKC77 and the Paleogene geological time-scale that we were led to reconsider ages for both, and to make several important modifications as outlined in this paper.

It should, however, be kept in mind that age estimates of geological epoch boundaries are by now difficult to derive

completely independently. This is again due to the fact that rocks elsewhere than the ocean floor provide dates for calibration and both correlation to the geomagnetic reversal sequence and assessment of the dates themselves are often developed within a biostratigraphic framework. The lack of independence is particularly apparent in the Neogene where age estimates of important boundaries are very often already obtained in close conjunction with correlations to the geomagnetic time-scale (e.g. Ryan *et al.* 1974). Unless long lava sequences, devoid of fossils but possible to date radiometrically, are found, or a reliable method is developed to date oceanic crustal rocks, a certain degree of circular reasoning (or more optimistically, positive feedback) is almost inevitable. Nevertheless, there is an impelling motivation and a justification for considering both sets of data simultaneously because the highly developed correlations between biostratigraphy and magnetostratigraphy demand a set of ages consistent within both frameworks. Thus any change in the estimated ages within one framework automatically implies a corresponding change in the other, unless the correlations can be shown to be incorrect.

Nomenclature of magnetic polarity intervals

Several systems of nomenclature have been used in referring to magnetic polarity intervals. According to recommendations of the Subcommission on Stratigraphic Classification (Anonymous 1979), the chron is now the basic unit of geomagnetic polarity. Thus, the intervals of predominantly normal or reversed polarity in the 0–5 Ma radiometric magnetic reversal time-scale are now referred to as chrons instead of epochs, for example, the Brunhes Chron. Shorter intervals of opposite polarity within the chrons can be referred to as subchrons, for example, the Jaramillo Subchron within the Matuyama Chron, and so forth.

Although the four most recent chrons are named after eminent geomagnetic researchers (Brunhes, Matuyama, Gauss and Gilbert), this system was not continued for earlier chron subdivisions. In magnetostratigraphic studies, Hays & Opdyke (1967) introduced an identification scheme in which chrons below the Gilbert were numbered sequentially from 5 (the first four chrons retaining their familiar names). Subchrons were identified by letter suffixes added to the chron numbers. This scheme was extended by Theyer & Hammond (1974a, b) and Opdyke *et al.* (1974) to chron 23 (correlating to near the Miocene–Oligocene boundary).

In subsequent magnetostratigraphic investigations of pre-Neogene sections, even this numbering scheme was discontinued and the magnetic chrons have been named after the correlative magnetic anomaly nomenclature. In the system of LeBrecque *et al.* (1983), a chron is defined as extending from the youngest reversal boundary of one numbered anomaly to the youngest reversal boundary of the next older numbered anomaly; a letter 'C' (for chron) is prefixed to avoid confusion with the pre-existing Neogene chron numbering nomenclature. Other similar schemes have also been proposed (Cox 1982).

These latter nomenclatures recognize the prime importance of the marine magnetic anomaly record in providing a history of geomagnetic reversals. In this paper, we often refer to magnetic anomalies as synonymous to their chron units because we feel such references are less ambiguous until general acceptance of a particular nomenclature emerges. The chron nomenclature of LeBrecque *et al.* (1983) is included in Figs 3, 5 and 6 for comparison of this scheme with the magnetic anomaly sequence.

Where we use this system in the text, the suffix N (e.g., C6CN) refers to the normal polarity interval(s) associated with the magnetic anomaly (e.g., anomaly 6C); the suffix R (e.g., C6CR) refers to the dominantly reversed polarity interval separating the numbered anomaly (e.g., Anomaly 6C) and the next older anomaly (e.g., Anomaly 7).

Revised geomagnetic reversal time-scale

As a representative sequence of geomagnetic polarities for the late Cretaceous to Recent, we use a slightly modified version of LKC77. As discussed in their paper, LKC77 incorporates several refinements to the original HDHPL68 rendition, in particular, revisions in the polarity reversal pattern between anomalies 5 and 6 (Blakely 1974), between anomalies 29 and 34 (Cande & Kristofferson 1977), and up to anomaly 3A (Klitgord *et al.* 1975). The only modification we make to LKC77 is to recalculate the polarity intervals described by Blakely (1974) according to the original age estimate in HDHPL68 for the younger end of anomaly 5, rather than use the slightly different value from Talwani *et al.* (1971) that was used by Blakely. The resulting overall sequence is thus constructed from essentially the same data as in a recent revision suggested by Ness *et al.* (1980), yet is still very similar to LKC77 for ease in comparison.

For the purpose of this discussion, we consider the ages for polarity reversals in LKC77 to be simply a quasi-linear measure of the relative position of the polarity intervals, in effect, a measure of distance or thickness in some idealized section formed at a nearly uniform rate. Unlike HDHPL68 which was largely based on the relative spacing of magnetic anomalies in a single profile, the present standard sequence reflects an aggregate of several segments, each averaged over several profiles and from different spreading systems, and is therefore highly unlikely to be observed anywhere in its entirety, with exactly the same relative spacing. Thus while it would be preferable conceptually to use a true length unit in describing a standard reversal sequence and to refer to actual rates in discussing the implications of its age calibration, the use of time units as common denominator is required to express the best estimate of a geomagnetic reversal sequence synthesized from varied sources. Although this sequence cannot be verified exactly in any single magnetic anomaly profile, it is generally acknowledged that such a composite sequence of many profiles averaged together yields a more complete and reliable record of the geomagnetic reversal pattern. Because LKC77 or any time-scale is an interim scale, we will refer to the units they are given in as apparent time units to facilitate discussion of their recalibration in time.

The age calibration tie-points we use are listed below and plotted with respect to their position in the modified LKC77 reversal sequence in Fig. 1. All ages have been converted where necessary to the new K-Ar radiometric dating system constants using tables in Dalrymple (1979).

(a) 3.40 Ma — Anomaly 2A or the Gauss–Gilbert boundary (Mankinen & Dalrymple 1979). Based on an analysis of radiometric date-magnetization polarity determinations on unrelated lavas. This is presently the oldest well-dated reversal in the classical 0–5 Ma radiometrically dated reversal time-scale and a traditional tie-point in virtually all late Cretaceous to Recent geomagnetic time-scales.

(b) 8.87 Ma — Anomaly 5y (Younger end of anomaly 5). Based on stratigraphic distribution of radiometric date-
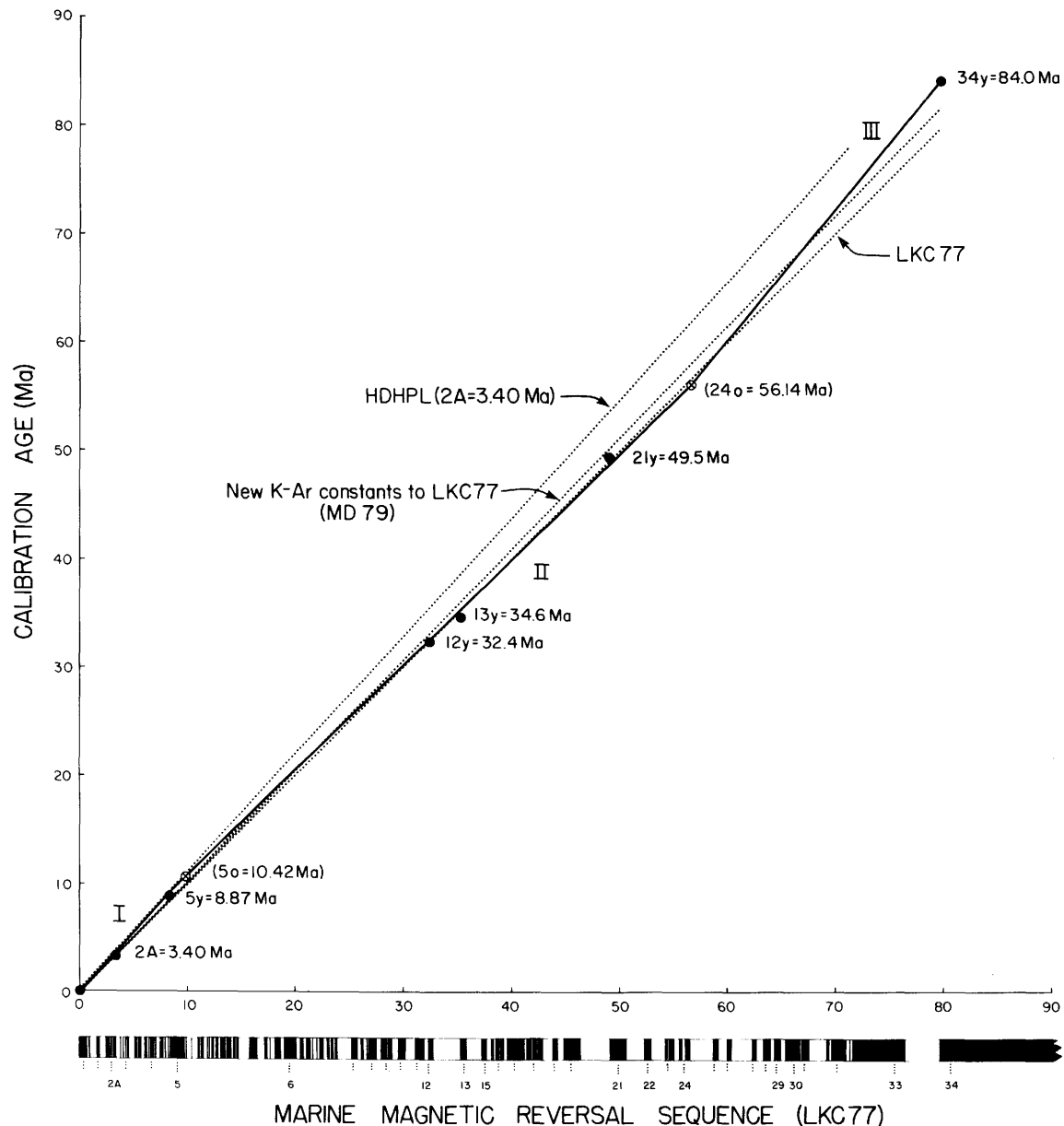
FIG. 1. Revised age calibration of marine magnetic reversal sequence from LaBrecque *et al.* 1977 (LKC77). Solid lines are three linear apparent age-calibration age segments (I, II, and III) which satisfy calibration tie-points indicated by solid circles (see text for details). The two open circles with X's at anomalies 5o and 24o are the inferred inflection points whose ages are derived by extrapolation from linear segments I and II, respectively. Shown for comparison by dotted lines are the geomagnetic polarity time-scales of Heirtzler *et al.* (1968) (HDHPL68 with anomaly 2A set to 3.40 Ma to conform with current estimate) and LaBrecque *et al.* 1977 (LKC77 in original form and modified (MD79) to account for new K-Ar constants as calculated by Mankinen & Dalrymple 1979). Anomaly numbers are indicated below bar graph of geomagnetic reversal sequence (filled for normal, open for reversed polarity).

magnetic polarity determinations on lavas from New Zealand and Iceland. Age represents the mean of 8.90 Ma from New Zealand and 8.83 Ma from Iceland (Evans 1970; Harrison *et al.* 1979).

(c) 32.4 Ma — Anomaly 12y (Chron C12N). Based on magnetostratigraphic studies in Oligocene vertebrate-bearing continental beds in the western United States. Radiometric (K-Ar) date on biotite in volcanic ash stratigraphically overlying normal magnetozone correlated to anomaly 12 (Evernden *et al.* 1964; Prothero *et al.* 1982, 1983).

(d) 34.6 Ma — Anomaly 13y (Chron C13N). Same source as item C; radiometric date (K-Ar) on biotite in volcanic ash stratigraphically overlying normal magnetozone correlated to anomaly 13 (Evernden *et al.* 1964; Prothero *et al.* 1982, 1983).

(e) 49.5 Ma — Anomaly 21y (Chron C21N). Based on magnetostratigraphic studies on Eocene continental and

marine beds in the western United States. Age interpolated from radiometric (K-Ar) dates on lavas and tuffs stratigraphically bracketing the top of a normal magnetozone correlated to anomaly 21 (Flynn 1983a, b). Further details in discussion of Eocene in this paper.

(f) 84.0 Ma — Anomaly 34y (Chron C34N). Age estimates for Campanian–Santonian boundary by Obradovich & Cobban (1975) on basis of K-Ar dates on bentonites from western interior of North America; the Campanian–Santonian boundary lies very near to the upper part of a normal magnetozone, correlated to anomaly 34, in Italian limestones (Lowrie & Alvarez 1977).

A characteristic feature of the above calibration data is that they are all based on the same dating system, K-Ar radiometric dates on high temperature minerals. Except for item f, minimal correlation is necessary to associate the radiometric

date with a magnetozone and both the date and magnetization were usually measured on material from the same section. Correlation of the magnetozones (and associated radiometric age estimates) is also not strongly dependent on biostratigraphy for these items as a group, independently for items (a) and (b), and only partly dependent for items (c), (d), and (e). The use of item (f) for calibration does, however, depend on biostratigraphic correlations since the magnetostratigraphy and radiochronology were determined in different places; the magnetochronological and biochronological age estimates for the Campanian–Santonian boundary are therefore set to be equivalent.

The radiometric age estimate for the younger end of anomaly 5 (8.87 Ma, item (b), above) is very near to the age extrapolated for this anomaly in HDHPL68 (8.92 Ma, using the revised 3.40 Ma date instead of 3.35 Ma for anomaly 2A). This is a strong indication that the original HDHPL68 time-scale provides a good chronologic framework for polarity reversals at least out to this anomaly. Beyond anomaly 5, calibration tie-points (c), (d), and (e) fall off from what would be the extension of the HDHPL68 trend (Fig. 1) and seem to define a different linear relationship between calibration age and apparent age; the change apparently occurs somewhere between the top of anomaly 5 (item (b)) and the top of anomaly 12 (item (c)). This new trend, however, cannot also accommodate the calibration tie-point at anomaly 34 (item (f)) and a change to another relationship must therefore occur somewhere between anomaly 21 (item (e)) and anomaly 34 (item (f)).

A minimum of two changes in the relationship between calibration age and apparent age in modified LKC77 are therefore required to satisfy this set of data. Such changes will have a direct effect on global sea-floor spreading rates and will either introduce or modify accelerations at the point in the anomaly sequence where they are introduced. Accordingly, we seek other evidence of change in the plate tectonic regime to guide the most appropriate placement for these modifications so as to reduce the possibility of producing spurious accelerations that are simply an artifact of an improperly constructed time-scale. We believe the most likely, and at the same time the least disruptive, positions for these calibration age-apparent age inflections occur at around anomaly 5 and at around anomaly 24, for the following reasons.

1. Large changes in sea-floor spreading rates, beyond the likely errors in previous time-scales, have already been noted at around anomaly 5 in the Indian Ocean (Weissel & Hayes 1972) and in the South Pacific (Heirtzler *et al.* 1968). Introduction of a time-scale change at around anomaly 5 would therefore mostly only alter the magnitude of the reported changes in sea-floor spreading and be less likely to introduce new, perhaps spurious, ones.

2. The period at around anomaly 24 is associated with the opening of the Norwegian Sea (Talwani & Eldholm 1977), perhaps the beginning (Weissel & Hayes 1972; but see Cande & Mutter 1982) of separation between Australia and Antarctica, and other evidence for major plate reorganization. A change in sea-floor spreading rates on a global scale at about this time might therefore not be unexpected and was originally observed in the South Pacific (Heirtzler *et al.* 1968). Additional discussion of the anomaly 24 problem is given in Ness *et al.* (1980).

We assume that the inflection points at anomaly 5 and anomaly 24 divide the geomagnetic reversal sequence into three linear calibration age-apparent age segments. Segment I extends from the origin to anomaly 5 and its slope in Fig. 1 is defined on the basis of items (a) and (b), including the origin. We extrapolate this trend to derive an estimated age of 10.42 Ma for the older end of anomaly 5, which compares favourably with a radiometric age estimate of 10.30 Ma (Harrison *et al.* 1979) from the Icelandic lavas. We chose not to use this radiometric age estimate as a calibration tie-point because additional work in progress in Iceland appears to suggest a radiometric age estimate somewhat older than originally reported for the base of anomaly 5 (I. McDougall, pers. comm. 1982).

Segment II is based on a linear best-fit through the data of items (c), (d), and (e) while constrained to join segment I at the 10.42 Ma age derived for the base of anomaly 5. The inflection between segments I and II is therefore fixed at the base of anomaly 5. Note that the trend of segment II lies very near to LKC77 (Fig. 1) and therefore provides a very similar chronology over this interval. Extrapolation of segment II yields an estimated age of 56.14 Ma for the base of anomaly 24 at which point we assume the second inflection occurs to accommodate the 84 Ma date for anomaly 34.

Segment III is simply an interpolation between the age derived for the base of anomaly 24 and the inferred age of anomaly 34 (item (f)). Note that the difference in trend between segments II and III is appreciably larger than between segments I and II, a possible reflection of a larger alteration in plate tectonic regime (sea-floor spreading rates) at around anomaly 24 than anomaly 5. It is also paradoxical that the change in trend at around anomaly 24 brings segment III toward the original HDHPL68 time-scale. Thus the original HDHPL68 gives 76.33 Ma for the base of anomaly 32 compared to our estimate of 73.55 Ma, a difference of less than 4%.

Ages for magnetic polarity intervals or chrons are calculated according to the linear regression equations of these three segments. A tabulation of these ages is presented in Table 1. The calculated ages are given to the nearest 0.01 Ma to reflect the precision in determination of the relative duration of the polarity chrons. The accuracy of a chron age ultimately depends on the uncertainty in the calibration age estimates which are typically quoted as a few percent of the calculated date. However, the method of calibration which we employ reduces sensitivity to the error in any calibration age determination to the extent that the assumption of linear calibration age-apparent age segments is valid. As a result of this procedure, the calculated age also will not necessarily correspond exactly to the calibration age of a tie-point; for example, the age derived for anomaly 13y is 35.39 Ma compared to an age of 34.6 Ma (item (d)) used in calibration. The differences give some indication of the magnitude of likely error in the absolute age of any given anomaly (in the context of the present data set) and encouragingly these differences seem to lie within the range of error associated with the actual calibration date determinations.

Comparison with biochronology

The magnetobiostratigraphic correlations reviewed and discussed elsewhere in this paper allow a detailed comparison of Paleogene epoch boundary age estimates. In Fig. 2 we plot the portion of the revised geomagnetic reversal time-scale between anomaly 6 time and anomaly 31 time against bio-

TABLE 1. Revised geomagnetic polarity time-scale for Cenozoic and late Cretaceous time.

Normal Polarity Interval (Ma)	Anomaly	Normal Polarity Interval (Ma)	Anomaly
0.00– 0.73	1	24.04–24.21	6C
0.91– 0.98		25.50–25.60	7
1.66– 1.88	2	25.67–25.97	7
2.47– 2.92	2A	26.38–26.56	7A
2.99– 3.08	2A	26.86–26.93	8
3.18– 3.40	2A	27.01–27.74	8
3.88– 3.97	3	28.15–28.74	9
4.10– 4.24	3	28.80–29.21	9
4.40– 4.47	3	29.73–30.03	10
4.57– 4.77	3	30.09–30.33	10
5.35– 5.53	3A	31.23–31.58	11
5.68– 5.89	3A	31.64–32.06	11
6.37– 6.50		32.46–32.90	12
6.70– 6.78	4	35.29–35.47	13
6.85– 7.28	4	35.54–35.87	13
7.35– 7.41	4	37.24–37.46	15
7.90– 8.21	4A	37.48–37.68	15
8.41– 8.50	4A	38.10–38.34	16
8.71– 8.80		38.50–38.79	16
8.92–10.42	5	38.83–39.24	16
10.54–10.59		39.53–40.43	17
11.03–11.09		40.50–40.70	17
11.55–11.73	5A	40.77–41.11	17
11.86–12.12	5A	41.29–41.73	18
12.46–12.49		41.80–42.23	18
12.58–12.62		42.30–42.73	18
12.83–13.01	5AA	43.60–44.06	19
13.20–13.46	5AB	44.66–46.17	20
13.69–14.08	5AC	48.75–50.34	21
14.20–14.66	5AD	51.95–52.62	22
14.87–14.96	5B	53.88–54.03	23
15.13–15.27	5B	54.09–54.70	23
16.22–16.52	5C	55.14–55.37	24
16.56–16.73	5C	55.66–56.14	24
16.80–16.98	5C	58.64–59.24	25
17.57–17.90	5D	60.21–60.75	26
18.12–18.14	5D	63.03–63.54	27
18.56–19.09	5E	64.29–65.12	28
19.35–20.45	6	66.50–66.17	29
20.88–21.16	6A	66.74–68.42	30
21.38–21.71	6A	68.52–69.40	31
21.90–22.06	6AA	71.37–71.65	32
22.25–22.35	6AA	71.91–73.55	32
22.57–22.97	6B	73.96–74.01	
23.27–23.44	6C	74.30–80.17	33
23.55–23.79	6C	84.00–118.00	34

chronological age by which we mean age estimates of epoch boundaries based on assessment of biostratigraphically controlled radiometric dates. The correlated positions of the epoch boundaries to the geomagnetic time-scale are extended by lines parallel to the biochronologic age axis; the biochronologic age estimate for each epoch boundary can then be plotted on its corresponding line assuming the boundary is correctly correlated to the geomagnetic sequence. The better the magnetochronologic and biochronologic age estimates for the boundaries agree, the closer will the points lie to a 45 degree trend intersecting the axis.

The solid symbols in Fig. 2 represent biochronologic age estimates we favour for the Paleogene epoch boundaries (a full discussion of these age estimates follows under appropriate headings). We find substantial agreement between these age estimates and the ages estimated on the basis of correlation to the revised geomagnetic reversal time-scale. The largest discrepancy is at the Paleocene–Eocene

boundary where an assessment of radiometric dates suggests an age of 56.5 Ma which is about 1 Ma younger than the magnetochronologic age estimate of 57.8 Ma. Respective age estimates for the Eocene–Oligocene boundary (37 Ma and 36.6 Ma) differ by 0.4 Ma, but in the opposite sense, while those for the Oligocene–Miocene boundary (23.5 Ma and 23.7 Ma) are in substantial agreement. There is some controversy concerning the age of the Cretaceous–Tertiary boundary, i.e. an age of about 63.5 Ma cited by Lerbekmo et al. (1979a, b) vs. about 66.5 Ma as estimated from recalculated dates in Obradovich & Cobban (1975). The magnetochronologic age estimate based on our revised geomagnetic reversal time-scale is 66.4 Ma which agrees well with the latter interpretation of the age of the Cretaceous–Tertiary boundary. Work is in progress to resolve the apparent discordance in dates relevant to this level. (J. Obradovich, pers. comm. 1982; see also discussion below in section on Cretaceous–Tertiary boundary).

We point out that incorporation of our preferred biochronologic age estimates for these Paleogene epoch boundaries as calibration tie-points would not appreciably alter the chronology we derive for the geomagnetic reversal time-scale. The high internal consistency of these data sets also supports the use of the geomagnetic reversal time-scale to estimate ages for other biostratigraphic boundaries correlated to the reversal sequence, for example, subdivisions of the epochs. Such age estimates can be read off the charts in Figs 3, 5, and 6.

Included in Fig. 2 for comparison are age estimates for boundaries of subdivisions of the Paleogene that have been suggested elsewhere. Plotted as open circles are the ages estimated by Odin & Curry (1981) and Curry & Odin (1982) which are based mostly on K-Ar dates on glauconites from NW Europe. While these ages are in reasonable agreement with our preferred estimated age for the younger (Oligocene–Miocene) limit of the Paleogene, they are appreciably younger for the remaining interval. For example, compare 53 Ma to our bio-(magneto)chronologic estimate of 56.5 Ma (57.8 Ma) for the Paleocene–Eocene boundary and 34 Ma to 36.6 Ma (37 Ma) for the Eocene–Oligocene boundary. The numerical age differences are largest in the Eocene, up to about 7 m.y. for the early–middle Eocene boundary (45 Ma against our magnetochronological estimate of 52 Ma for the base of the Lutetian). We suspect that these conflicting age estimates most likely reflect a geochemical problem, having to do with systematic errors in either the glauconite dates favoured by Odin or in the high temperature mineral dates which we use in calibration of the geomagnetic reversal time-scale and toward which our biochronologic age estimates are biased (see further discussion on this point in Appendix II). For reasons discussed below, we prefer the generally older set of age estimates for subdivisions of the Paleogene which are supported by high temperature mineral dates. We therefore consider the glauconite dates from NW Europe generally to be anomalously young. An age estimate of about 61 Ma (recalculated to about 62.5 Ma) for the Cretaceous–Tertiary boundary based on glauconite dates from the eastern coastal plain of North America (Owens & Sohl 1973) also appears problematically young.

Lastly, we show in Fig. 2 the calibration tie-point ages (open square symbols) used in the geomagnetic reversal time-scale of Lowrie & Alvarez (1981). We believe that the changes implied in sea-floor spreading rates are largely artifacts of inaccuracies in the closely-spaced calibration tie-
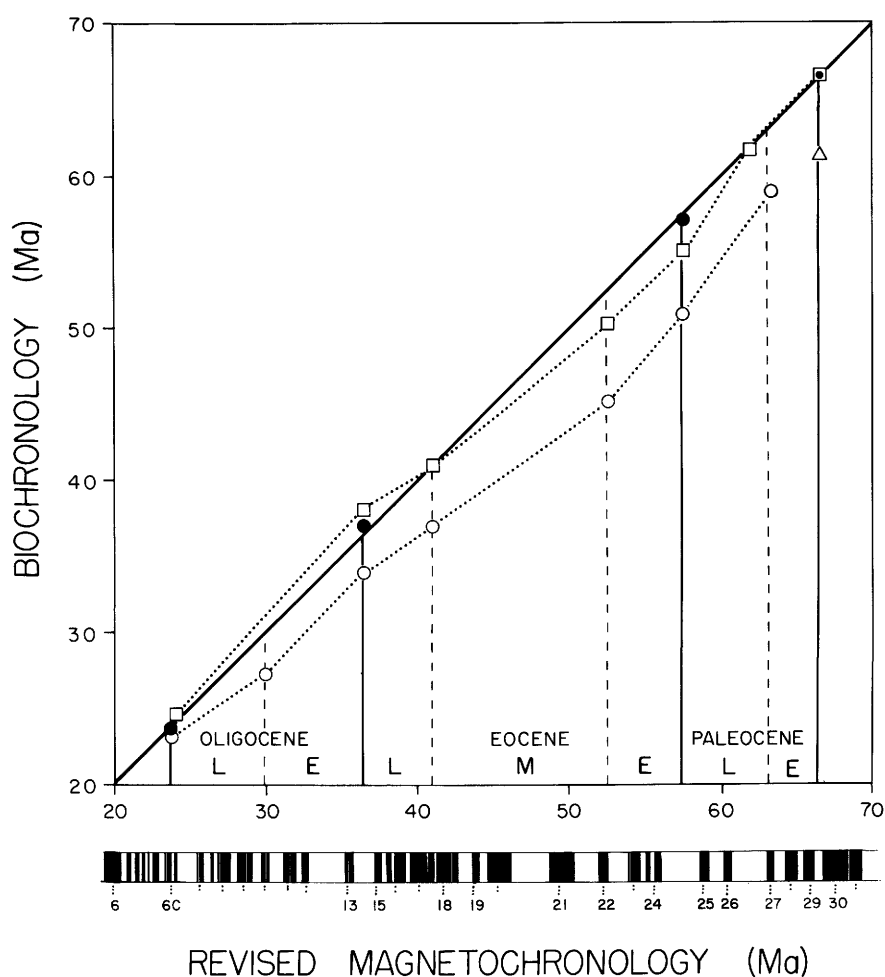
FIG. 2. Comparison of various biochronological estimates of Paleogene epoch and intra-epoch boundaries within magnetochronological framework provided by correlation to revised geomagnetic polarity time-scale. Solid circles: this paper. Open circles: from Odin & Curry (1981). Squares: recalculated from Hardenbol & Berggren (1978) by Ness *et al.* (1980) and used for calibration of geomagnetic reversal sequence by Lowrie & Alvarez (1981). Open triangle: from Owens & Sohl (1973). Anomaly numbers are indicated below bar graph of geomagnetic reversal sequence (filled for normal, open for reversed polarity).

point ages (now superceded) used in the Lowrie & Alvarez scale.

An assessment of magnetobiochronology for the Neogene is presented in the companion paper (Berggren *et al.*, this volume). To complete the analysis of the late Cretaceous to Recent interval incorporated in our revised geomagnetic reversal time-scale, magnetobiochronological data for the late Cretaceous is discussed in Appendix III.

The Cretaceous–Tertiary boundary

Until the end of the nineteenth century the Danian Stage remained, by almost universal consent, at the top of the Cretaceous. It was De Grossouvre (1897) who made the suggestion that the Mesozoic–Cenozoic boundary be placed at the upper stratigraphic limit (i.e. disappearance) of ammonites, rudistids, belemnites, inoceramids, dinosaurs, mosasaurs, plesiosaurs, and other characteristic Mesozoic animals. These faunal elements have since been shown to have disappeared at the top of the Maestrichtian Stage. In retrospect it is an interesting fact that the strata of the Danian Stage, although placed in the Upper Cretaceous by Desor (1847) (and correlated with the calcaire pisolithique of the Paris Basin, now regarded as Dano-Montian in age), were earlier considered to be of Tertiary age by Forchhammer (1825) who made the first systematic study of them. Recent palaeontologic and stratigraphic studies would appear to have

vindicated both Forchhammer and Desor.

Nevertheless the bio- and chronostratigraphic affinities and correlation of the Mesozoic–Cenozoic boundary and of the Danian Stage have continued to be debated by several workers. Two differing view-points have been summarized by Voigt (1960, 1979, 1981) and Eames & Savage (1975) who favour including the Danian within the terminal Cretaceous and by Berggren (1964, 1971) who favours including the Danian at the base of the Cenozoic. The arguments of the former are based primarily upon similarities between various components of the marine benthic faunas in Maestrichtian and Danian strata (although the argument of Eames is weakened by the appeal to similarities in lithologic facies of strata of both ages in some regions, some inaccurate bio-stratigraphic data and a failure to acknowledge the essential contemporaneity of the Tuffeau de Ciply (Mons Basin) and the Danstekalk (Denmark). The argument presented by Berggren was based predominantly upon the global extinction of marine microplankton and nekton at the end of the Maestrichtian Age and the repopulation and radiation which occurred in strata referable to the Danian Stage. The majority of stratigraphers now appear to have adopted the latter interpretation.

The Cretaceous–Tertiary boundary has recently become the focus of renewed interest (Christensen & Birkelund 1979; Silver & Schultz 1982). Recent work on the biostratigraphy and palaeomagnetic stratigraphy of marine deposits from Europe, including the boundary stratotype at Stevns Klint,

Denmark (Mörner 1982), and the deep ocean basins (Alvarez *et al.* 1977; Alvarez & Lowrie 1978; Alvarez *et al.* 1980; Hsü *et al.* 1982) indicates that the Cretaceous–Tertiary boundary (recognized by planktonic microfossil events) occurs within the reversed polarity interval preceding Anomaly 29 time (C29R). Analysis of sedimentation rates in the Gubbio section (Apennines) indicates that the faunal turnover at the boundary was rapid, possibly 10 000 yrs or less (Kent 1977). Indeed, Smit (1982), on the basis of a preliminary palaeomagnetic study of the Gredero section in SE Spain, has argued for a scenario in which the mass extinction event may have occurred within 50 yrs and a new stable planktonic fauna established within 35 000 yrs. Anomalously high iridium values in marine sediments in Italy, Denmark and New Zealand, among other places, have been reported at the biostratigraphically determined boundary between the Maestrichtian and Danian Stages (Alvarez *et al.* 1979, 1980). This iridium anomaly has since been reported in Spain (Smit & Hertogen 1980; Smit 1982) and Tunisia (Smit, pers. comm. 1982) within an expanded stratigraphic section that exhibits a distinct and rapid replacement of Cretaceous planktonic foraminiferal taxa by small forms which diversify into recognizable elements of basal Danian Age (Smit 1977, 1982; Smit & Hertogen 1980). This anomaly has also been reported recently at several DSDP sites in the Atlantic and Pacific oceans.

It has been suggested that the iridium anomaly (and seemingly related abrupt extinction of marine microfauna) was the result of an asteroid (with dimensions of approximately 10 ± 4 km; Alvarez *et al.* 1979, 1980) or cometary (Hsü 1980) impact that would have had catastrophic consequences upon marine and terrestrial biotas (references above; Emiliani 1980; Emiliani *et al.* 1981; Hsü *et al.* 1982; Hsü 1980, 1983; O'Keefe & Ahrens 1982; but see Kent 1981; Reid 1981; Gartner & McGuirk 1979; various papers in Silver & Schultz 1982; Officer & Drake 1983, *i. al.* for alternate viewpoints and interpretations).

McLean (1981a, b) has questioned the catastrophic theory of terminal Cretaceous extinctions and suggested that they may be hiatus controlled illusions of an incomplete stratigraphic record. However, current magnetobiostratigraphic studies on several DSDP cores and correlation with marine sections on land suggest the simultaneity and abrupt nature of the extinction event in the oceans at a level within magnetochron C29R. In a comprehensive review of the terminal Cretaceous extinctions within fossil plankton, Thierstein (1982) has reviewed the evidence in support of the catastrophic mass extinction hypothesis at the end of the Cretaceous due to a bolide impact but notes that ultimate verification of this scenario awaits higher stratigraphic resolution and a better knowledge of noble element geochemistry than is presently available. Finally Alvarez *et al.* (1984a, b) have reviewed the published invertebrate fossil record and mineralogic data which they believe indicates that the Cretaceous–Tertiary boundary event was instantaneous and synchronous at various boundary localities.

Analyses of the Cretaceous–Tertiary boundary in terrestrial sections have provided a conflicting portrayal of the timing and nature of the Cretaceous extinctions. The Cretaceous–Tertiary boundary in terrestrial sections is frequently recognized at the highest stratigraphic occurrence of dinosaurs. Recent biostratigraphic and magnetostratigraphic studies of this boundary in the San Juan Basin, New Mexico (Butler *et al.* 1977, 1981a; Lindsay *et al.* 1978, 1979a,

b and c, 1981, 1982) have located the Cretaceous–Tertiary boundary (based on dinosaurs) within a reversed polarity zone correlated with Chron C28R (or possibly within the underlying normal polarity interval correlated with C29N). This conflicts with the position of the Cretaceous–Tertiary boundary recognized in marine sections, where it is placed in a reversed polarity zone correlated with Chron C29R. These results indicate a non-synchronous Cretaceous–Tertiary boundary that differs in age from 0.5–1.5 million years between terrestrial and marine realms. If this conclusion is correct, a catastrophic extinction event at the Cretaceous–Tertiary boundary is unlikely.

A number of studies (Clemens & Archibald 1980; Archibald 1981; Clemens 1981; McLean 1981a, b; Schopf 1981; Archibald & Clemens 1982) support a non-catastrophic extinction and faunal replacement of terrestrial vertebrates throughout the late Cretaceous and across the Cretaceous–Tertiary boundary. Clemens & Archibald (1980), Clemens (1981), McLean (1981) and Clemens *et al.* (1981) supported a diachronous terrestrial Cretaceous–Tertiary boundary, based on extinction patterns of land vertebrates and floras and marine invertebrates. These were purely biostratigraphic conclusions, as no radiometric or magnetostratigraphic data were used in these studies.

Several authors have provided alternative results or have questioned the conclusions drawn by workers in the San Juan Basin. Lerbekmo *et al.* (1979a, b) located the Cretaceous–Tertiary boundary (based on both dinosaurs and palynoflora) in a reversed polarity zone that they correlated with Chron C29R. The palynofloral Cretaceous–Tertiary boundary occurs slightly higher than, but still within the same reversed polarity interval as, the boundary recognized by the highest stratigraphic occurrence of dinosaurs. Alvarez & Vann (1978), Fassett (1979), Lucas & Rigby (1979) and Lucas & Schoch (1982) have criticized various aspects of the San Juan Basin magnetostratigraphic and biostratigraphic correlations. Several potential problems are mentioned, such as incorrect or contradictory biostratigraphic age assignments and correlations, major depositional hiatuses and unconformities, and incorrect correlation between the observed magnetostratigraphy and the standard marine magnetic anomaly sequence. In particular, Alvarez & Vann (1979), Lucas & Ribgy (1979), and Lucas & Schoch (1982) stressed the possibility that the published San Juan Basin magnetostratigraphy is incorrect. Alternatively, they propose that the Cretaceous–Tertiary boundary in these sections *might* lie between normal polarity intervals correlative with anomalies 29 and 30, which would be consistent with the location of this boundary in marine sections.

Archibald *et al.* (1982) recently described a terrestrial sequence from Montana containing the Cretaceous–Tertiary boundary. They located this boundary, and the highest stratigraphic occurrence of dinosaurs, within an interval of reversed polarity, Polarity Interval B– (although the boundary may fall in the underlying normal polarity interval, A+, in one section). On the basis of biostratigraphy, Archibald *et al.* (1982) correlated the normal polarity interval, A+, with a normal zone in the Alberta, Canada section of Lerbekmo *et al.* (1979a; in which this zone was correlated with Chron C30N) and the San Juan Basin sections of Butler *et al.* (1977; and other later papers; in which this zone was correlated with Chron C29N). In all three sections the Cretaceous–Tertiary boundary recognized by dinosaurian, mammalian and palynologic biostratigraphy lies

within the upper reversed polarity interval (or possibly the underlying normal polarity zone in the San Juan Basin and one Montana section). However, Archibald *et al.* (1982, p. 159) specifically avoided correlation of the magnetostratigraphies (and the location of the Cretaceous–Tertiary boundary) in these three sections with the standard polarity time-scale, stating: 'Again, we stress that until the current controversy regarding correlation of the magnetic polarity sequence in the San Juan Basin is resolved, or other pertinent data become available, the magnetic polarity zones recorded in these terrestrial sections in Alberta, Montana, and New Mexico cannot be securely correlated with the magnetic polarity time scale.' It seems, therefore, that the degree of synchroneity between the Cretaceous–Tertiary boundary in terrestrial and marine sequences cannot be resolved by the presently available magnetostratigraphic data.

Floral evidence has also been used to recognize the Cretaceous–Tertiary boundary in terrestrial sections. Lerbekmo *et al.* (1979a, b) used palynoflora to locate this boundary just above the last occurrence of dinosaurs in their sections. Both of these events lie within a reversed polarity zone that they correlated with Chron C29R. However, their magnetostratigraphic section cannot be uniquely correlated to the magnetic polarity time-scale (see above, and references cited). However, Lerbekmo *et al.* (1980, in response to comments by Butler & Lindsay 1980) reasonably argue that the palynomorphic change they use to recognize the Cretaceous–Tertiary boundary in Alberta also occurs in Montana, Wyoming and North Dakota. In North Dakota this boundary is overlain by marine strata containing a *Globigerina edita* zone foraminiferal fauna. The *G. edita* Zone is equated with the early Paleocene *Globorotalia pseudobulloides* and *Globigerina eugubina* zones, and the Cannonball Formation (the base of which is at least 20 m above the Cretaceous–Tertiary boundary) spans the *G. pseudobulloides* Zone. At Gubbio, Italy the *G. pseudobulloides* Zone spans an interval correlated with part of Chron C28N to part of C29R. This evidence supports the original magnetostratigraphic correlations of Lerbekmo *et al.* (1979a), and the placement of the Cretaceous–Tertiary boundary within Chron C29R in both the terrestrial and marine realms. Other floral biostratigraphy studies of the Cretaceous–Tertiary boundary have not been directly associated with magnetostratigraphic data.

Orth *et al.* (1981a, b) used palynologic events to recognize the Cretaceous–Tertiary boundary in the Raton Basin, Colorado. This boundary lies at the base of a thin coal bed in association with an iridium anomaly. If this iridium anomaly is correlative with the iridium anomaly found at the Cretaceous–Tertiary boundary in marine sections, it would support synchrony of this boundary between terrestrial and marine realms, and an extraterrestrial cause for the extinctions marking this boundary.

The palaeobotanical work of Fassett (1981), Hickey (1981a, b; 1984) and Clemens *et al.* (1981) conflicts with a catastrophic, instantaneous terminal Cretaceous extinction. Fassett (1981) located the palynologic Cretaceous–Tertiary boundary *below* the boundary recognized by the last occurrence of dinosaurs in the San Juan Basin. Hickey (1981a, b) invoked a non-catastrophic climatic deterioration to explain the gradual, geographically variable extinction pattern he observed for land plants in the late Cretaceous and across the Cretaceous–Tertiary boundary. Further, Hickey (1981a) cited three areas where this boundary was diachronous; in all three sections latest Cretaceous floras

persisted several metres or more above the highest occurrence of dinosaurs. Clemens *et al.* (1981) emphasized the points made by Hickey (1981a, b), and concluded that the terminal Cretaceous extinctions were gradual and may have occurred over a period of time ranging from several years to hundreds of thousands of years.

Based on the available evidence, we place the Cretaceous–Tertiary boundary within the reversed polarity interval between anomalies 29 and 30 (i.e. Chron C29R). We believe that further work will show that this boundary, as recognized in marine and terrestrial realms, is synchronous. The validity of an instantaneous, catastrophic cause for the terminal Cretaceous extinctions is uncertain.

The most recent reviews covering the age of the Cretaceous–Tertiary boundary are Curry & Odin (1982) and Harland *et al.* (1982). Both agree that an age of 65 Ma would be a reasonable estimate given the lack of definitive data below and above the boundary in marine strata. With regard to continental strata where the boundary has been placed to coincide with the disappearance of dinosaurs, a major extinction in pollen (*Aquilapollenites*), and the first appearance of Puercan (Paleocene) mammals, a discrepancy in the age of the boundary has arisen. The Denver Formation, near Golden, Colorado (at a level 22 m above the boundary) has been dated at 65.8 ± 0.7 Ma (new constants; Obradovich & Cobban 1975) but further north in eastern Montana and southern Alberta Lerbekmo *et al.* (1980) have dated bentonites 1 metre above the boundary at 63 ± 2 Ma indicating that the boundary as so recognized in continental strata might be a diachronous horizon. However, recent work covering the same stratigraphic interval reveals that this boundary may indeed be closer to 66 Ma (Obradovich 1984).

The Paleocene

The Paleocene is here considered to consist of two stages, the Danian and the Thanetian (Hardenbol & Berggren 1978), although various other terms (e.g. Montian, Landenian, Selandian, Sparnacian, *i. al.*) are also used in various combinations by some authors (Curry *et al.* 1978). We shall not enter into a comprehensive review here of the applicability of these terms (see, rather, the discussion in the two references cited above as well as Cavelier & Roger 1980; Pomerol 1981).

The Danian Stage, as recently redefined with the type area extended from east Sjaelland (= Zealand) to include all of Denmark, and the boundary stratotype designated at Nye Klov (Jutland) rather than Stevns Klint (Zealand) (Thomsen 1981), corresponds essentially to planktonic foraminiferal Zone P1 and calcareous nannoplankton zones NP1–NP3 (?NP4 *partim*). The unconformity bounded Danian Stage is sandwiched between two eustatic sea-level regressions (Vail *et al.* 1977) and corresponds to the first transgressive cycle of the Cenozoic. The Danian s.s. can be correlated with the Tuffeau de Ciply (= lower Montian) of Belgium (Rasmussen 1964, 1965; Berggren 1964; Meijer 1969). However, the upper or type Montian (Calcaire de Mons) is younger than any Danian sediments exposed in Denmark and older than subsequent deposits of the Selandian Stage. The Montian s.s. can be correlated with post-Danian and pre-Thanetian limestones (with similar molluscan faunas) in the Crimea which can, in turn, be traced into the subsurface into beds containing planktonic foraminiferal faunas referable to the *Morozovella uncinata* (P2) Zone (Berggren 1964; see also

Curry *et al.* 1978: 39). In the interest of parsimony, the concept of the Danian has been extended upward to include the Montian s.s. as an expanded Danian s.l. (Berggren 1964, 1971; Hardenbol & Berggren 1978)

The actual temporal extent of the Danian Stage (as estimated by magnetobiostratigraphic cross correlation; see Fig. 3) has been derived in the following manner.

The Danian s.s. would appear to be bracketed (below) by the LAD's of *Micula murus* and *Lithraphidites quadratus* and the globotruncanids (younger part of Chron C29R) and (above) by the FAD's (or concurrent ranges) of *Ellipsolithus macellus, Neochiastozygus modestus, N. saepes, Prinsius martinii* and *Heliorthus concinnus* (with a zeugoid rim and central X) and *Planorotalites compressus* and *Subbotina trinidadensis* (within Chrons C27R to C28N; see Appendix IV, tables 3 and 4, and discussion below).

The extent of the Danian s.l. (as correlated here by the FAD of *Morozovella angulata*) is more problematic, owing to problems in magnetobiostratigraphic correlation in this part of the record.

There are three different interpretations of the magnetic polarity stratigraphy in DSDP Hole 527 over the 20 m interval of 258 m − 278 m involving anomaly correlatives 27–29. They are as follows:

1. Chave (1984; 529) suggests that the long normal interval between 267.41 m and 278.02 m represents an expanded anomaly 29 correlative. He then notes the FAD's of *E. macellus* (= NP4) at 258 m and *Fasciculithus tympaniformis* (NP5) at 249.78 m above an incompletely recovered normal event (258.75 m 260.77m) which he identifies as (part of) anomaly 28 correlative. He suggests that anomaly 27 correlative is not present (but should lie) between the FAD's of *F. tympaniformis* (NP5) at 249.28 m and *H. kleinpelli* (NP6) at 245.43 m. This interpretation may have been based upon early, unpublished interpretations of the magnetostratigraphy of DSDP Leg 73, and in particular Hole 524. However, it is now well established that the FAD's of *F. tympaniformis* and *H. kleinpelli* occur within the mid-part of Chron C26R, well *above* anomaly 27 correlative (see Appendix IV, Table 4).

2. Boersma (1984; 513) suggests that anomaly correlatives 27–29 are compressed in the predominantly normal interval between 268.278 m in Hole 527 and that anomaly correlative 27 lies close to 269 m and 28 close to 272 m. Boersma also identifies the younger normal event at 258–260 m with anomaly 27 correlative. The FAD's of *Morozovella angulata* and *Planorotalites compressus* were said (*op. cit.*: p. 513, Table 6; cf. Fig. 3, p. 510) to occur near 269 m associated with anomaly correlative 27 (although this is shown as 28 on Fig. 3). A cross-check of the barrel sheet data and the stratigraphic range chart (*op. cit.*: p. 512, Table 4) shows that the FAD of *M. angulata* is associated with the younger normal event identified with anomaly 28 by Chave (1984) and 27 by Boersma (1984). This record of *M. angulata* associated with anomaly 27 correlative is consistent with records from Gubbio, although it has been reported earlier elsewhere (see Appendix IV, Table 4).

3. Shackleton *et al.* (1984: 622) suggests that anomaly correlatives 28 and 29 are present in the predominantly normal polarity interval between 268–278 m and that the younger normal (258–260 m) is anomaly 27 correlative. They further note (*op. cit.*: p. 625) that the position of anomaly correlatives 27–30 is quite unambiguous and cite Chave's work in support of this statement. But Chave (1984) has suggested a different interpretation of the magnetic polarity sequence as we have seen above.

The FAD of *E. macellus* has been generally recorded within the lower part of Chron C26R at several DSDP sites but has recently been recorded from the Bottacione section, Gubbio (Italy) in Chron C27R (see Appendix IV, Table 4) as is the case with the FAD of *M. angulata*. Thus the magneto-biostratigraphic correlations to date give little support one way or the other in terms of the interpretation of the younger normal polarity event at 258–260 m in Hole 527 as either anomaly 27 or 28 correlatives.

However, we are reasonably safe in stating that the FAD of *Ellipsolithus macellus* predates that of *Morozovella angulata* in the stratigraphic record. This, added to the fact that *E. macellus* is known to be a solution susceptible taxon, suggests that the FAD of *E. macellus* (at a level correlative with the upper Danian s.s.) is probably associated with Chron C27R, whereas the top of the Danian s.l. is to be associated with the FAD of *M. angulata*, within the lower part of C26R (see Appendix IV, Table 4).

The temporal extent (i.e. numerical values) of the sea floor anomalies 27–29 (see Table 1) is such that the interpretation of Shackleton *et al.* (1984) is preferred here based on the assumption of a uniform and slow rate of sedimentation. Thus we show the Danian Stage s.l. extending from Chron C29R to C26R (approximately 66.4–62.3 Ma) with a duration of about 4 Ma (see Fig. 3).

The succeeding Thanetian Stage corresponds predominantly to Zone NP8 (Curry *et al.* 1978; Curry 1981; Aubry, 1983), although it may extend into NP9 at the top (Curry 1981; Hamilton & Hojjatzadeh 1982; but see discussion below). Its lower part (Pegwell Marls and subjacent, essentially non-calcareous clays and conglomerates) may be somewhat older than NP8 (Curry *et al.* 1978). The Thanet Beds rest disconformably upon Coniacian or Santonian chalk and are overlain by the Woolwich Beds (= Sparnacian). Thus, there is a demonstrable biostratigraphic gap between an extended Danian s.l. (the top of which is within zones P2 and NP3) and the Thanetian (whose base is within NP8–?NP7 = within P4), which led Curry (1981: 263) to admit that the Thanetian Stage, based on the Thanet Beds, is 'only a moderately satisfactory concept.' If we accept the concept that the 'base defines stage', the Thanetian is seen to rest well above the Danian; nor can the concept of the Danian be satisfactorily extended upwards to include the intervening interval (corresponding to Zone P3 and NP4–6; ?NP7). The intervening interval spans about 2–3 m. y. and, indeed, represents about a quarter to a third of Paleocene time (as revised herein). There are two alternatives: (1) insert a stage representative of this time-stratigraphic interval; (2) replace the term Thanetian with a time-stratigraphic unit which spans the interval from top Danian to base Ypresian.

There are two stage names which come to mind immediately: the Landenian (including Heersian) of Belgium (Dumont 1839, 1849; Laga 1981) and Selandian of Denmark (Rosenkrantz 1924; Perch-Nielsen & Hansen 1981). It is beyond the scope of this paper to enter into a detailed historical discussion of these two stages (see discussions presented by the authors cited above). Suffice to say that both units are essentially equivalent to the Thanetian in their upper part, the lower part of the Landenian s.l. (= Heersian = Orp-le-Grand sands) is only questionably slightly older than the basal Thanet Beds (both are within the *Cyprina morrisi* Zone; Curry et al. 1978), but the basal Selandian is demonstrably older than either of the above. It is for this reason that we would suggest insertion of (or replacement by) the Selandian as a standard Paleocene stage.

The Selandian Stage consists of a lower (Lellinge Greensand), middle (Kerteminde Clay) and upper (grey unfossiliferous clay) unit. The Selandian contains a typical Midway benthic foraminiferal fauna, and corresponds to dinocyst zones *Deflandrea speciosa* (= Lellinge Greensand and Kerteminde Clay) and the (lower) *Apectodinium hyperacanthum* Zone (grey unfossiliferous clay); to calcareous nannoplankton zones NP4 and 5 (= Lellinge Greensand and Kerteminde Marl; Perch-Nielsen 1979). The discovery of *Morozovella angulata* in the lower part of the Selandian (Hansen 1968) indicates correlation (at least of that part) with planktonic foraminiferal Zone P3. The Selandian is overlain by the ash-bearing series, the Mo Clay Formation which belongs to the middle to upper part of the *A. hyperacanthum* Zone (Hansen 1979; Heilmann-Clausen 1982), which provides direct, first order correlation with equivalent stratigraphies in England and continental Europe.

Thus the Selandian Stage is seen to span the entire post-Danian Paleocene and corresponds in its middle to upper part with the Thanetian Stage of England and to the Woolwich-Reading Beds = Sparnacian of France (see below). The Selandian Stage could be conveniently inserted in the Paleocene chronostratigraphic hagiography between the top of the Danian (= P2) and the base Thanetian (= NP7/8) (Selandian, restricted sense) or extended to include the upper (Thanetian) part of the Paleocene (= NP8–NP9, ? lower part of NP10) (Selandian, *sensu stricto*) (see Fig. 3). We leave this question open for the moment but would point out that the latter procedure would have the advantage of having the stratotype area (and concomitant sections) of two successive time-stratigraphic units lying in temporal and spatial continuity (i.e. in Denmark). The uppermost part of the Mo Clay ash-series lies within the *oebisfeldensis* Acme-subzone of the *Apectodinium hyperacanthum* Zone and provides direct correlation with the locally developed Division 1A (Harwich Member) of the London Clay Formation and which contains the youngest ash beds in southern England and also lies within the *oebisfeldensis* Acme-subzone (see discussion below under Paleocene/Eocene boundary).

The Thanet beds have been shown above to correspond essentially to zone NP8 and questionably to a part of NP9. Potassium-argon dates on glauconites from two levels within the Thanet Beds at Herne Bay have been presented by Fitch *et al.* (1978). The data, and our micropalaeontological correlation are presented below.

Location	Biostratigraphic age	Apparent age (Ma)
1. 5 m below top of Reculver Sands, Bishop-stone Glen, Herne Bay, Kent	NP 8–9	58.2 ± 0.6
2. basal Thanet Beds, Pegwell Bay, Kent	no indigenous Paleocene nannoplankton present in Pegwell Marls (= lower Thanetian); by stratigraphic relationships with overlying Thanet sands (= NP8) this level cannot be younger than NP8; possibly slightly older (M. P. Aubry, pers. comm.)	60.9 ± 0.9

A third radiometric date from the Sables de Bracheux at Butte de Reneuil (France) has been previously cited in

Berggren *et al.* (1978). This is a Rb-Sr date on glauconite and has been recalculated to 59.2 Ma by Berggren *et al.* (1978). The biostratigraphic age of the Sables de Bracheux is late Paleocene, probably latest Thanetian and/or earliest 'Sparnacian' based on the following evidence:

1. Presence of *Wetzeliella parva* (restricted to the *hyperacanthum* Zone in both the Sables de Bracheux and 'argiles et lignites du Sparnacien' (Châteauneuf & Gruas-Cavagnetto, 1969: 132, 137).
2. Presence of *Discoaster multiradiatus* (= NP9) (Aubry 1983; see also Curry *et al.* 1978: 40).
3. Molluscan faunal links with the Woolwich Formation (*Pitharella arenaria, Corbicula cordata, Ostrea bellovacina*) (Curry 1967; Curry *et al.* 1978: 40).

The date of 59.2 Ma on the Sables de Bracheux at Butte de Reneuil is seen to lie intermediate between the two (glauconite) dates on the type Thanetian, although it is probably stratigraphically equivalent or only slightly younger than the youngest Thanetian exposed in England, i.e. it is stratigraphically equivalent or slightly younger than the sample dated 58.2 Ma near the top of the Reculver Sands.

The Sparnacian problem

The question of the Sparnacian 'Stage' is dealt with in more detail in the succeeding section dealing with the Paleocene–Eocene boundary. Suffice here to observe that the Sparnacian (Conglomérat de Meudon, Argiles et lignites du Soissonnais, Sables de Sinceny, Faluns à Cyrènes et à Huîtres) of the Paris Basin is considered to be the biostratigraphic correlative of the Woolwich-Reading Beds of England (Curry *et al.* 1978), belongs to the *Apectodinium hyperacanthum* (dinocyst) Zone (Costa & Downie 1976, 1978; Châteauneuf & Gruas-Cavagnetto 1978) (which is generally equivalent with calcareous nannoplankton Zone NP9). This would appear to be corroborated by the reported occurrence of *Discoaster multiradiatus* in the Reading Bottom Bed at Berkshire (Hamilton & Hojjatzadeh 1982) and in the topmost fossiliferous sample from the Thanet Sands at Reculver (Kent). However, this is somewhat difficult to reconcile with palaeomagnetic data (Hailwood, pers. comm. 1982) and recent integrated deep sea studies on magnetobiostratigraphy.

Magnetostratigraphic studies in SE England (Townsend 1982; Townsend & Hailwood, in press) have shown that the upper 85% of the Oldhaven Formation at Herne Bay is of normal polarity, whereas the underlying Woolwich Formation and all of the Thanet Formation at Herne Bay are of reverse polarity, and that a normal polarity interval is present in the lower part of the Thanet Formation at Pegwell Bay. While the simplest interpretation (and the one we have adopted, see below) would be to correlate the Oldhaven and Thanet magnetozones with Chrons C25N and C26N, respectively, Townsend & Hailwood (in press) have drawn attention to problems with this interpretation. The normal polarity zone in the ash-bearing Oldhaven Formation at Herne Bay corresponds with a similar normal polarity event in the Oldhaven unit at Harefield and the top of the ash-bearing Harwich Member of the London Clay Formation at Wrabness. The latter is equated with the upper part of the *A. hyperacanthum* Zone (Knox & Harland 1979). An underlying assumption here is the correlation of these ash-bearing horizons in SE England with the main North Sea 'Ash Marker' and with the distinct ash-bearing unit in DSDP Site

403 (Rockall Plateau). However, as Townsend & Hailwood (in press) point out, the ashes in DSDP Sites 403 and 550 are reversely magnetized throughout, precluding direct correlation of the *totality* of the Rockall ash beds with the normal polarity Oldhaven ash units. Townsend & Hailwood (in press) suggest that the Oldhaven magnetozone may represent an intermediate normal polarity interval between anomaly correlatives 24B and 25 and not identified at DSDP Sites 403 and 550 owing to low sedimentation rates. A poorly-defined short normal polarity interval has been identified below the dominantly reversed polarity ash series at DSDP Site 401 close to the NP9/NP10 boundary. If this short normal polarity event at DSDP Site 401 is correlative with the Oldhaven magnetozone, it would suggest that the ash beds at Site 401 and SE England are approximately contemporaneous. Townsend & Hailwood (in press) conclude that the Oldhaven normal magnetozone probably represents a short normal polarity interval (of early NP10 age) intermediate between anomaly correlatives 24B and 25.

Inasmuch as magnetobiostratigraphic studies in deep sea cores and continental marine sections have shown that Chron C25N essentially straddles the NP8/NP9 boundary, Townsend & Hailwood (in press) suggest that the lower Thanet magnetozone represents Chron C25N, or alternatively, an additional short normal polarity zone intermediate between Chron C25N and C26N. In the latter case, which they appear to favour, the position of Chron C25N would correspond to the stratigraphic hiatus between the Thanet and Woolwich—Reading formations.

In this paper we have preferred what we view as a more parsimonious interpretation (Fig. 3) in associating the Oldhaven magnetozone with Chron C25N and the Thanet magnetozone with Chron C26N in view of the fact that the identification of the intermediate normal polarity intervals between anomaly 24B and 25 correlatives, and 25 and 26 correlatives, as well as the recognition of the corresponding oceanic basement anomalies remain poorly documented.

Magnetobiostratigraphic studies on deep sea cores have failed to demonstrate the presence of *Discoaster multiradiatus*, nominate taxon of Zone NP9, older than Chron C25N (see Appendix 4, and Fig. 3). If we examine the data on calcareous nannoplankton from the Reading and Thanet formations of England (Hamilton & Hojjatzadeh 1982) we note the following (Aubry 1983):

1. *Discoaster multiradiatus* was not illustrated from either the Reading or Thanet levels.
2. The specimens of *multiradiatus* illustrated (Hamilton & Hojjatzadeh 1982, pl. 6.1, Figs 9, 10) from the Selsey Formation, Bracklesham Group (Middle Eocene), Selsey, Sussex are poorly preserved, and at least one, (Fig. 10) could be *D. barbadiensis*, a typical early—late Eocene taxon. The specimens illustrated on pl. 6.2, Figs 1, 2, from the same locality are of *D. bifax*, a typical middle Eocene taxon.
3. The range of several taxa (table 6.1, p. 140, 141) are anomalous, for instance, *H. riedeli*, to Zone NP18 (restricted to Zone NP8), *C. bidens* to NP18 (NP3—NP10), *D. multiradiatus* to NP15 (NP9—NP11), *D. kueppen* to NP15 (NP12—NP14).
4. The Woolwich—Reading Beds are dominantly alluvial/fluviatile, lagoonal and estuarine, and one would not normally expect to find marine microplankton in them. The record of '*Discoaster multiradiatus*' from the Reading Bottom Bed may represent reworking from the older (marine) Thanet Beds, but the magnetobiostratigraphic data discussed above

suggests that this taxon may have been misidentified.

Thus we consider the record of *Discoaster multiradiatus* in the Reading and Thanet Beds to remain undocumented and would correlate the Thanet and overlying Reading—Woolwich Beds to Zone NP8 (or its equivalent).

Indeed the only record in northern Europe of *Discoaster multiradiatus* (with a calcareous nanoflora assemblage typical of that Zone) with which we are familiar is from the Sables de Bracheux (Paris Basin) (in Curry *et al.* 1978; Aubry 1983). This suggests that the Sables de Bracheux (with an NP9 nanoflora) may be the equivalent of the Oldhaven Formation in England (Chron C25N, the oldest level from which *D. multiradiatus* has been reported to date in deep sea cores).

As we have noted above the Sparnacian is within the *Apectodinium hyperacanthum* Zone. Inasmuch as the succeeding dinocyst *Wetzeliella astra* Zone is found in the overlying basal Sables de Cuise s.l. (= Cuisian) and in correlative, basal layers of the Ieper Clay (Ypresian) and London Clay Formation, of earliest Eocene age, the Sparnacian is demonstrably of latest Paleocene age.

The Sparnacian has alternatively been interpreted as latest Paleocene or earliest Eocene in age by various workers. In actual fact it is a partially marine but predominantly brackish to non-marine marginal facies (with associate hiatuses), probably deposited during the interval of a (predominantly) terminal Paleocene regression associated with a brief global (relative) eustatic sea-level fall. Our concept of Paleocene geochronology is shown in Fig. 3.

Magnetostratigraphic studies of terrestrial Paleocene sequences have largely been confined to the San Juan Basin, New Mexico (Butler *et al.* 1977, 1981a; Lindsay *et al.* 1978, 1979a–c, 1981, 1982; Taylor & Butler 1980; and see Cretaceous—Tertiary boundary discussion, above). However, several studies from other areas have sampled the earliest Paleocene (Montana: Archibald *et al.* 1982; Alberta, Canada: Lerbekmo *et al.* 1979a, b, 1980), early to early middle Paleocene (Utah: Tomida & Butler 1980; Tomida 1981), and middle to late Paleocene and the Paleocene—Eocene boundary (Wyoming: Butler *et al.* 1981b; West Texas: Rapp *et al.* 1983).

The San Juan Basin sections extend from below the Cretaceous—Tertiary boundary to unfossiliferous horizons above Torrejonian land mammal faunas. The magnetostratigraphic sequence for this interval has been correlated to the magnetic polarity time-scale between the younger part of Chron C31N and just younger than Chron C25N (see for example Lindsay *et al.* 1981). In this area the stratigraphic range of Cretaceous dinosaurs extends into a normal polarity interval correlated with Chron C29N, Puercan (*Ectoconus* Zone and *Taeniolabis* Zone) mammals are restricted to a normal polarity interval correlated with Chron C28N, and Torrejonian (*Deltatherium* Zone and *Pantolambda* Zone) mammals range from low within a reversed polarity interval correlated with Chron C26R to near the top of a normal polarity interval correlated with Chron C26N. Lindsay *et al.* (1978, 1981) extend the range of Torrejonian mammals down into the upper part of a normal polarity interval correlated with Chron C27N, based on the occurrence of *Periptychus*, a common early Torrejonian genus. Based on this work in the San Juan Basin, Lindsay *et al.* (1981, p. 128) suggest general 'guidelines' for predicted boundary limits of the Puercan and Torrejonian Land Mammal Ages in North America. These predicted limits include the occurrence of Puercan mammals

between (but probably not including) Chrons C27N and C29N, and the occurrence of Torrejonian mammals in Chrons C26N to C27N.

Based on the work of Tomida & Butler (1980) in Utah, Tomida (1981) considers the 'Dragonian' as earliest Torrejonian in age, rather than as a distinct land mammal age. Tomida (1981, p. 237–238) proposes a new *Periptychus-Loxolophus* Zone for this portion of the earliest Torrejonian. This zone appears (Tomida 1981, Fig. 10. 3) to extend over a stratigraphic range from the middle to the top of (or slightly higher than) a normal polarity zone correlated with Chron C27N. In adding the *Periptychus-Loxolophus* Zone to the Torrejonian, Tomida (1981) has extended the earliest part of the temporal range of the Torrejonian down into the middle of an interval which he correlates with Chron C27N. Tomida & Butler (1980) also document the presence of the 'Wagonroad faunal level' in a normal polarity interval correlated with Chron C28N, and within the base of the immediately overlying reversed polarity interval. This position is temporally younger than, and presumably stratigraphically higher than, the position of the Puercan faunas in the San Juan Basin. If the fauna of the 'Wagonroad faunal level' of Utah comes to be considered Puercan in age, it would extend the top of the temporal range of the Peurcan up into the time of Chron C27R (as correlated by Tomida & Butler 1980).

Middle to late Paleocene and earliest Eocene terrestrial sediments have been sampled in the Clark's Fork Basin, Wyoming (Butler *et al.* 1981b) in sections containing Tiffanian, Clarkforkian and Wasatchian faunas. The lengthy Polecat Bench South Section (Butler *et al.* 1981b, Fig. 4) also contains a Torrejonian (Rock Bench Quarry) and a Puercan (Mantua Quarry) faunal horizon below the Tiffanian to early Wasatchian portion of the section. In the Clark's Fork Basin sequence Tiffanian faunas occur within strata deposited during a reversed polarity interval correlated with Chron C26R to strata deposited during a normal polarity interval correlated with Chron C25N. Clarkforkian faunas occur within strata deposited during a normal polarity interval correlated with Chron C25N. and the overlying reversed polarity interval. Early Wasatchian faunas occur to the local top of the section within reversely magnetized strata believed to be deposited during Chron C24R. The correlation of the Clark's Fork Basin magnetostratigraphy to the magnetic polarity time-scale appears to be very reliable and is supported by magneto- and bio- stratigraphic work on the position of the Paleocene–Eocene boundary (see next section).

This temporal correlation results, however, in a major temporal discordance between the Clark's Fork Basin and San Juan Basin sequences. In the San Juan Basin the Torrejonian extends into Chron C26N, while in the Clark's Fork Basin the Tiffanian begins somewhere within Chron C26R. These correlations yield a temporal overlap of at least 50% between two supposedly temporally successive, non-overlapping, mammalian temporal units. Assuming accurate magnetostratigraphic correlation of the San Juan Basin section, Butler *et al.* (1981b, p. 313–314) presented two tentative explanations for this temporal discrepancy. Both of these explanations suggest significant temporal equivalence and overlap between the Torrejonian and Tiffanian land mammal ages due to the effects of a north–south geographic separation of the areas sampled. Because of this temporal overlap, one of the authors (P. Gingerich) does not support the claim of the others (R. Butler and E. Lindsay) that the correlation of the San Juan Basin polarity sequence to the polarity time-scale is correct. An alternative explanation of the temporal discrepancy is that the correlation of the San Juan Basin sequence to the polarity time-scale is *not* correct, and the top of the Torrejonian in the San Juan Basin occurs within a normal polarity interval correlated with Chron C27N rather than Chron C26N. Lindsay *et al.* (1981, 1982) use mammalian biostratigraphic similarities between the Cretaceous San Juan Basin faunas and those from Wyoming and Canada as one of their key arguments for establishing the temporal continuity, age, and palaeomagnetic correlation for the early parts of their San Juan Basin sequence. It is surprising that they unconditionally accept the temporal equivalence (based partly on the negative evidence of the absence of certain taxa in the San Juan Basin sequence) of Cretaceous faunas and mammalian faunal ages from areas as widely separated as New Mexico and Canada, while Butler and Lindsay (in Butler *et al.* 1981b; see above) readily accept significant temporal overlap of Paleocene mammalian faunal ages between New Mexico and Wyoming.

Recent work by Rapp *et al.* (1983) on a Paleocene–Eocene sequence in the Big Bend National Park area, West Texas, supports the chronologic conclusions of Butler *et al.* (1981b) for the Clark's Fork Basin. Rapp *et al.* (1983) have sampled a greater than 160 m section through the Black Peaks Formation to the base of the overlying Hannold Hill Formation in this area to the south of the San Juan Basin. This section contains a poor Torrejonian/Tiffanian fauna near its base, and good Tiffanian, Clarkforkian and Wasatchian faunas higher in the section. Their magnetostratigraphy includes three normal polarity intervals correlated with Chrons C26N, C25N, and C24N. Tiffanian faunas occur in strata correlated with Chrons C26R to C25N a Clarkforkian fauna occurs within strata correlated to C24R, and Wasatchian faunas occur in unsampled strata overlying strata containing a normal polarity interval correlated with Chron C24N.

These results are consistent with those from the Clark's Fork Basin, and they conflict strongly with those from the San Juan Basin. They support temporal equivalence of Tiffanian faunas throughout the time represented by Chrons C26N to C25N across widely separated geographic intervals. The north–south geographic separation invoked by Butler *et al.* (1981b) to explain the supposed temporal overlap of the Tiffanian and Torrejonian mammal ages is invalidated by the presence of Tiffanian faunas in the Chron C26N to C25N time interval at more southerly latitudes than the San Juan Basin.

It is interesting to note that in a revised Paleocene and early Eocene magnetic polarity time-scale, Butler & Coney (1981) cite the work of Butler *et al.* (1981b) in the Clark's Fork Basin, but do not mention the extensive work of Butler, Lindsay and others in the San Juan Basin. The Clark's Fork Basin study is essential to their use of the Paleocene–Eocene boundary as a radiometric calibration point in their polarity time-scale. The other calibration point for their time-scale is the Cretaceous–Tertiary boundary, for which they use a terrestrially-derived radiometric age estimate of 66.7 Ma. This age estimate, however, is applied to a Cretaceous–Tertiary boundary point within Chron C29R (as it is located in marine sections), *rather than* Chron C28R (as the Cretaceous–Tertiary boundary is located in the terrestrial San Juan Basin sequence). It is unclear why Butler & Coney (1981) ignore the relevant San Juan Basin information.

Our placement of the boundaries of the Paleocene North

PALEOCENE TIME-SCALE

A large chart ("Paleocene Time-Scale") with columns for Geochronometric Scale in Ma, Magnetic Polarity (History, Anomaly, Chron), Plankton Zones (Foraminifera — Berggren 1969; Bolli 1957, 1966, Premoli-Silva & Bolli 1973, Stainforth et al. 1975; Calcareous Nannoplankton — Bukry 1973, 1975, Okada & Bukry 1980, Martini 1970, 1971), Radiolaria (Riedel & Sanfilippo 1978), Dinoflagellates (NE Atl. Costa & Downie 1976, NW Europe Costa & Downie 1976, Hansen 1977, 1979 a,b), Epochs, Standard Ages, Position of Stage Stratotypes, North American Land Mammal Ages, and Geochronometric Scale in Ma.

Ma	Anomaly	Chron	Foraminifera	Calc. Nanno (Bukry)	(Martini)	Radiolaria	Dinoflagellates	Epochs	Standard Ages	Stage Stratotypes	NALMA	Ma
55	23	C23	P8 Morozovella aragonensis	CP10	NP12	25	IV a / III Kisselovia coleothrypta	EARLY EOCENE	YPRESIAN	YPRESIAN	WASATCHIAN	55
56	24	C24	P7 Morozovella formosa formosa	—	b NP11	II Dracodinium similis / D. varielongi-tuda				CUISIAN		56
57			Morozovella subbotinae	CP9	a NP10	Ib W meckelfeldensis / W astra					1,4	57
58			P6 b Morozovella edgari / a			26					CLARKFORKIAN	1,3	58
59	25	C25	P5 Morozovella velascoensis	CP8 b / a	NP9	Ia Apectodinium hyperacanthum	LATE PALEOCENE	THANETIAN / SPARNACIAN	THANETIAN	TIFFANIAN		59
60	26		P4 Planorotalites pseudomenardii	CP7 / CP6	NP8 / NP7			SELANDIAN / UNNAMED	SELANDIAN / LANDENIAN (HEERSIAN)			60
61		C26		CP5	NP6		Deflandrea speciosa					61
62			P3 b Planorotalites pusilla pusilla / a Manguiata	CP4	NP5						? 1,3	62
63	27	C27	P2 Morozovella uncinata	CP3	NP4		Hafniasphaera cryptovesiculata				TORREJONIAN	63
64			c Subbotina trinididadensis	CP2	NP3				DANIAN	DANIAN / MONTIAN	? 1,2	64
65	28	C28	P1 b Subbotina pseudobulloides	CP1	NP2		Xenicodinium lubricum / Xenicodinium rugulatum	EARLY			PUERCAN	65
66	29	C29	a "Globigerina" eugubina		b / a NP1		Carpatella cornuta				1	66
67	30	C30	Abathomphalus mayaroensis				Palynodinium grallator	LATE CRETACEOUS	MAESTRICHTIAN			67
68												68

RADIOLARIAN ZONES
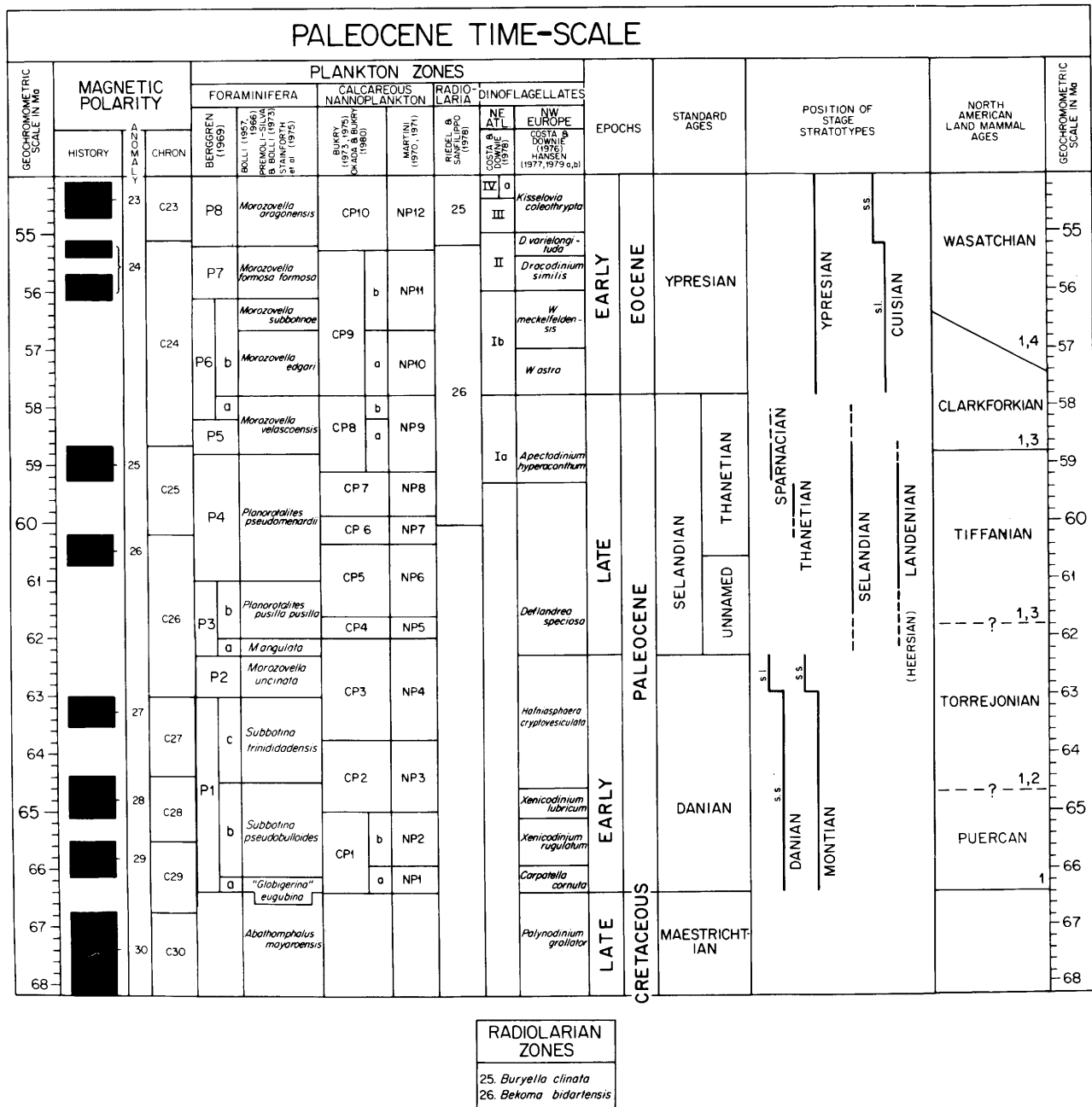25. *Buryella clinata*
26. *Bekoma bidartensis*

FIG. 3. Paleocene geochronology. The geochronologic scale at the margins of the figure is derived from the magnetic polarity chronology which is in turn derived from palaeontologically and/or palaeomagnetically controlled radiometrically dated calibration points in the late Neogene, early Oligocene, middle Eocene and late Cretaceous (see text for further explanation). The position of the calcareous plankton zonal boundaries is based, for the most part, upon direct (first order) correlation between biostratigraphic datum levels and palaeomagnetic polarity stratigraphy as determined in deep sea cores or continental marine sediments. In this way a true 'magnetobiochronology' is possible. The extent (duration) of standard time-stratigraphic units and their boundaries and the position of stage stratotypes are estimated on the basis of their relationship to standard plankton biostratigraphic zones.

Magnetobiochronology of Paleocene North American Land Mammal Ages is shown on the right (footnote numbers at boundaries refer to sources used in determining the temporal position of these boundaries). Boundaries shown as -- ? -- indicate our predicted boundaries in cases of conflicting evidence (see text); diagonal boundaries reflect uncertainty in precise relationship between boundary and magnetic polarity sequence or geochronometric scale.

Explanation of sources denoted by footnote numbers:
1) This paper — based on data and discussions presented in the text.
2) Adapted from Tomida & Butler 1980; Tomida 1981.
3) Butler *et al.* 1981a, b; Rapp *et al.* 1983.
4) Butler *et al.* 1981a, b; Rapp *et al.* 1983; Rose 1980; Gingerich 1976, 1980.
5) Radiometric dates and discussion in West *et al.*, in press.
6) Flynn 1983a, b.
7) Radiometric dates in Black 1969; McDowell *et al.* 1973.
8) Prothero *et al.* 1982, 1983 (supported by radiometric dates near the base of the Arikareean — R. H. Tedford, pers. comm.).
9) Radiometric dates in McDowell *et al.* 1973; Wilson 1980.

American Land Mammal Ages relative to the magnetic polarity time-scale is shown in Fig. 3. The placement of these boundaries for the middle and later Paleocene (Tiffanian to Clarkforkian) seems secure, based on the work of Butler *et al.* (1981b) and Rapp *et al.* (1983). However, for the middle and early Paleocene the location of these boundaries is more speculative. We tentatively place the base of the basal Tertiary Puercan within Chron C29R, and the base of the Torrejonian within the younger part of Chron C28N. These boundary placements are predictive, and are based on our belief that further detailed studies of terrestrial sequences will locate these boundaries in approximately the positions indicated in Fig. 3.

The consistent discrepancy of temporal correlations between the San Juan Basin sequence and those in other areas, at both the top (Butler *et al.* 1981b; Rapp *et al.* 1983; see above) and bottom (Lerbekmo *et al.* 1979 a and b, 1980; Alvarez & Vann 1979; Fassett 1979; Lucas & Rigby 1979; Orth *et al.* 1981 a and b; Lucas & Schoch 1982; see above) of the section, have forced us to re-evaluate the San Juan Basin magnetostratigraphic correlations. Faunal and magneto-stratigraphic correlations to the time-scale in both the upper and lower parts of the San Juan Basin sequence are younger than those from other areas. The Cretaceous–Tertiary boundary is placed within Chron C28R in the San Juan Basin, but it is located within Chron C29R elsewhere. Similarly, Chron C26N is associated with the Torrejonian in the San Juan Basin, while it is associated with the middle of the (younger) Tiffanian age elsewhere. We prefer to minimize the temporal discrepancies of correlations between other areas by placing the base of the Puercan (and the base of the Tertiary) within Chron C29R, the base of the Torrejonian within upper Chron C28N, and the base of the Tiffanian somewhere within Chron C26R. The base of the Tiffanian is at *least* as old as Chron C26R (based on Butler *et al.* 1981b; Rapp *et al.*, 1983), and the relative temporal durations of the Torrejonian and Puercan are approximately the same in this time-scale as in the temporal correlation proposed for the San Juan Basin sequence. Placing the base of the Torrejonian within the younger part of Chron C28N is an approximation based on the known location of the base of the Tiffanian, and on an assumption that the San Juan Basin magnetostratigraphic pattern is approximately correct, but that the temporal correlation of this pattern to the polarity time-scale is consistently (but in a complex manner) too young.

It is uncertain which (if any) of the explanations referenced above accounts for the presumed anomalous correlation of the San Juan Basin magnetostratigraphy. It is also possible that the magnetostratigraphic pattern would become more consistent with those from other areas under detailed thermal demagnetization treatment of samples from this section. At present the results from the San Juan Basin section are anomalous, but the discrepancies discussed above cannot be adequately explained. Resolution of these problems, and more precise refinement of the early Paleocene, terrestrial temporal framework await further detailed studies.

The Paleocene–Eocene boundary

Paleogene stratigraphy of NW Europe and the British Isles has been summarized most recently by Curry *et al.* (1978) and the lower Eocene London Clay and correlatives in NW Europe by King (1981).

The London Clay has been subdivided into components, formally designated lithostratigraphic units (King 1981). The Thames Group has been created with (a lower) Oldhaven Formation and (an upper) London Clay Formation. Five major transgressive and regressive cycles are recognized within the London Clay Formation which has been subdivided into five informal units (A-E; Fig. 4; see also Knox *et al.* 1983).

The Oldhaven Formation has not yielded a diagnostic microfauna or microflora but the base of the suprajacent, locally developed Division 1A (Harwich Member) of the London Clay Formation is in the *Apectodinium hyperacanthum* (dinocyst) Zone, which is present in the subjacent Woolwich Beds (= 'Sparnacian' = latest Paleocene). The top of the so-called 'ash series' in southern England is within the Harwich Member (and equivalents) and in the *A. hyperacanthum* Zone, whereas in the Central North Sea Basin ash beds extend into the *meckelfeldensis* Zone (see below).

The base of the *Wetzeliella astra* Zone has been found to lie approximately within 1 metre of the base of the overlying Walton Member (Division A2) of the London Clay and the succeeding *W. meckelfeldensis* Zone approximately 5 m above the base of the London Clay (Costa & Downie 1976; Denison 1977; Costal *et al.* 1978). Thus the *A. hyperacanthum–astra* zonal boundary lies within the basal part of the non-tuffaceous clays of Division A2.

The basal part of the Argile d'Ypres of Belgium and the Formation de Varengeville on the Normandy coast south of Dieppe (Seine-Maritime) are also placed within the *W. astra* Zone which has, in turn, been correlated with the *Tribrachiatus contortus* (NP10) Zone by Costa & Müller (1978). However, according to Aubry (1983), Zone NP10 has not been identified in any NW European marine sediments on land. It has been recorded recently from the Rockall area (Backmann, in Mortun *et al.*, 1983). The oldest early Eocene zone present is NP11. Calcareous nannoplankton are rather sporadically developed in the lower Eocene of Great Britain and NW continental Europe and are not found in the basal part of the London Clay (and correlative levels elsewhere). Indeed, the earliest appearance of calcareous nannoplankton in the early Eocene (Zone NP11) of NW Europe appears to be associated with an horizon rich in calcareous planktonic, and predominantly nodosariid benthic foraminifera which are within the *meckelfeldensis* and *similis* zones (King 1981). In the North Sea, Denmark and NW Germany this horizon is within the *similis-coleothrypta* Zone interval; see below. Characteristic elements of this horizon include (PF) *Subbotina patagonica* (= *Globigerina triloculinoides* auct.), *Acarinina triplex-coalingensis* gp., *Pseudohastigerina wilcoxensis* and (BF) *Nodosaria latejugata*, *Marginulina enbornensis*, *Clavulina anglica*, *Gaudryina hiltermanni*, *Anomalinoides grosserugosus*, *Turrilina brevispira*, *i. al.* (see also Williams 1982).

This basal, essentially calcareous plankton-free, interval of the London Clay and its correlatives in NW Europe would appear to span the time represented elsewhere by Zone NP10. If the shallow water, unconformity-bounded stratigraphic units of NW Europe reflect eustatic sea-level changes, we may well expect difficulties in precisely determining the age of basal sediments associated with each successive transgression-regression. The calcareous plankton appear to be present only during the transgressive peaks.

In a similar manner the calcareous nannoplankton suggest a hiatus between uppermost Paleocene and lowermost Eocene marine strata in the Gulf and Atlantic Coastal Plains of the United States. In the former region the Upper Paleocene Tuscahoma Sand is overlain by the Hatchetigbee Formation which has been assigned to the *Discoaster multiradiatus* Zone (Hay & Mohler 1967; Hay *et al.* 1967), which subsequently became Zone NP9 of Martini (1971; see also Siesser 1983: 27–29). The Hatchetigbee Formation has been assigned to Zone NP10 based on the occurrence of *Tribrachiatus bramlettei* (= *T. nunnii*) and *T. contortus* (Bybell 1980; Bybell in Reinhardt *et al.* 1980; Gibson & Bybell 1981; Gibson *et al.* 1982). However, the occurrence of *Discoaster binodosus* and *Chiasmolithus grandis* in this unit (*op. cit.*, faunal list) would indicate, if verified, an NP11 assignment. The floral list and illustrations provided by Siesser (1983: 27, 29) from the Hatchetigbee, on the other hand, would appear to support his NP9 assignment. In the subsurface Atlantic Coastal Plain the Aquia Formation (NP5-NP9) is separated from the overlying Nanjemoy Formation by an approximately 5–6 m thick non-calcareous unit, the Marlboro Clay. The Nanjemoy has been assigned to Zone NP10 (Bybell, in Gibson *et al.* 1980: 25) based on the listed occurrence of *Marthasterites tribrachiatus*. However, this taxon has its initial appearance in upper NP10 and *Discoaster binodosus*, which is recorded from near the base of the unit, appears in Zone NP11 which would appear to preclude assignment to Zone NP10. More recently Fredriksen *et al.* (1982) have reviewed the nannoplankton and sporomorph evidence in the Tuscahoma–Hatchetigbee sequence in the eastern Gulf Coast. They conclude that a (minor) hiatus (= paraconformity) exists between the Tuscahoma and Hatchetigbee units, spanning the time represented by latest NP9 and early NP10 zones and that the depositional patterns of coastal onlap followed by an abrupt regression at the Paleocene–Eocene boundary, followed by a rapid eustatic rise in sea-level in earliest Eocene time agrees well with the global coastal onlap curve of Vail & Mitchum (1979). It would seem that we are seeing an essentially contemporaneous and similar lithic expression of an essentially regional, global phenomenon, eustatic sea-level lowstand, on opposite sides of the North Atlantic. Biostratigraphic resolution in these marginal epicontinental facies precludes precise correlation, however, at the present time.

The Sparnacian Stage of the Paris Basin is within the upper part of the *Apectodinium hyperacanthum* Zone, equivalent to the acme of *Deflandrea oebisfeldensis* which characterizes the ash-series of the central North Sea, the ash series of East Anglia and the Danish Mo Clay (Bignot 1980; Costa *et al.* 1978; Knox and Harland 1979). Associated with the main ash episode in the North Sea, the lower Eocene of NW Germany, the Mo Clay of Denmark, and the basal London Clay of the Thames Estuary is an acme of (diatom) *Coscinodiscus* spp. The Woolwich Beds belong to the *hyperacanthum* Zone (based more on stratigraphic position than on definitive dinoflagellate evidence, however; Downie *et al.* 1971; Costa & Downie 1976) and are generally correlated with the Sparnacian. Although the Woolwich Marine Beds at Reculver are barren of calcareous nannofossils, they are most likely correlative with Zone NP8 (see discussion in preceding section).

The exact location of the Paleocene–Eocene boundary has been a subject of controversy since Schimper (1874) originally defined the Paleocene. Alternative placements have spanned the extremes of base Ilerdian (= base of *Nummulites deserti/fraasi* Zone) to base of Cuisian (= base *Nummulites planulatus* Zone) with intermediate positions including the base and top of the Sparnacian, base of the Ypresian, top of the Landenian, and others (see, for instance, Berggren 1971; Pomerol 1977; Costa *et al.* 1978; Curry *et al.* 1978; King 1981). Marine micropalaeontologists have drawn the Paleocene–Eocene boundary at various levels ranging from the *Planorotalites pseudomenardii*–*Morozovella velascoensis* (P4–P5) boundary to the *Morozovella formosa*–*M. aragonensis* (P7–P8) boundary, with intermediate positions including the P5–P6 boundary, the *Pseudohastigerina* Datum (within Zone P6), base zone NP9, base Zone NP10, middle of Zone NP10, base of the *Apectodinium hyperacanthum* Zone, base *W. astra* Zone, *i. al.*

King (1981) has drawn attention to the fact that current Paleocene–Eocene boundary definitions are inadequate because they propose to locate a major time-stratigraphic boundary at a lithologic discontinuity. If the base of the London Clay or base of the Cuisian ~ base Argile d'Ypres = Ypresian is chosen, the beds below are largely non-marine and contain few fossils of (regional) correlative value. A more appropriate procedure is to locate the boundary at a lithic level (with a 'golden spike') at which biostratigraphic criteria may serve to recognize, extend, and correlate this boundary elsewhere on a regional basis. King (1981) has followed a recommendation by a joint IGS/oil industry committee to locate the Paleocene–Eocene boundary at the *A. hyperacanthum*–*W. astra* zonal boundary. This level corresponds approximately to the P6a/b boundary of Berggren (1969), the NP9–10 boundary, and lies near the base of Division A2 of the London Clay Formation, and near the base of the Ieper Formation (Belgium) and the Cuisian *s.l.* (France).

If the Paleocene–Eocene boundary is drawn virtually at the base of the Ieper Formation, the Sparnacian is of terminal Paleocene age, although in its local (and regional) facies development it may span latest Paleocene–earliest Eocene time. The Sables de Sinceny, near the top of the Sparnacian, belong to the *hyperacanthum* Zone (Châteauneuf & Gruas-Cavagnetto 1978; Costa *et al.* 1978).

The late Paleocene–early Eocene dinoflagellate sequence has been recorded on Rockall Bank (DSDP Hole 117A) and the SW margin of Rockall Plateau (DSDP sites 403 and 404) (Costa & Downie 1979). The *hyperacanthum* Zone was recorded in the basal part (Cores 6–10) of Hole 117A and was said to be equivalent to the lower part of Zone Ia in the basal sediments of sites 403 and 404 (Costa & Downie 1978: 513, 522).

The *astra* Zone has been recognized in Hole 117A in section 1, Core 6 (Costa & Downie 1979: 522) at the NP9–10 boundary (Perch-Nielsen 1972: 1004). A somewhat different interpretation of the biostratigraphy of the basal sediments of DSDP Hole 117A has been presented by Morton *et al.* (1983). Calcareous nannoplankton suggest the presence of Zone NP10 from (at least) cores 7 to 4 and perhaps to the base of the hole (core 8 contains rare nannofossils and 9 and 10 are essentially barren). The presence of *Wetzeliella astra* in cores 4 to 8 indicates the presence of the *W. astra* (Ia1) Zone. The authors indicate that the correspondence between the base of Zone NP10 and the *W. astra* Zone remains unproved. The interval of upper Zone Ia (= *astra* Zone) and Ib (*meckelfeldensis* Zone) of sites 403 and 404 corresponds to zones NP10 and 11 (undifferentiated; Müller 1979: 182, 184,

but see Tables 13 and 15 on p. 603, 604, respectively). Volcaniclastic tuffs occur in the basal part of this sequence at sites 117A, 403 and 404 consistent with their stratigraphic occurrence in NW Europe (Costa & Downie 1976; Costa *et al.* 1978) and the North Sea (Jacqué & Thouvenin 1975; Knox & Harland 1979, 1983).

Determination of an age estimate for this boundary is not quite as straightforward. Ultimately it will depend upon an integration of biostratigraphic, radiometric and magneto-stratigraphic data. Recent age estimates for this boundary have ranged from 49–57 Ma (see Tarling & Mitchell 1976; Odin 1978; Odin *et al.* 1978; Hardenbol & Berggren 1978; Rubinstein & Gabunya 1978; Butler & Coney 1981; Butler *et al.* 1981b; Odin (ed.) 1982). Hardenbol & Berggren (1978) estimated the age of the Paleocene–Eocene boundary at 53.5 Ma following earlier work by Berggren (1969b, c, 1971a, 1972). In constructing their time-scale Ness *et al.* (1980), Lowrie & Alvarez (1981) and Lowrie *et al.* (1982) accepted this age estimate for the Paleocene–Eocene boundary, although all inappropriately 'recalibrated' this age estimate to 54.9 Ma by applying a correction for new K-Ar constants as if this subjective age estimate were an empirically determined radiometric date and even though one of the relevant controlling radiometric dates is based on the Rb-Sr system (see below). Hailwood *et al.* (1979) used the oldest value of the age range of 47–52 Ma (Hailwood *et al.* 1973) to estimate the age of the upper part of the East Greenland basalts and provide an approximation of the age of the beginning of Chron C24N. La Brecque *et al.* (1977) used van Eysinga's (1975) estimate of 55 Ma for the Paleocene–Eocene boundary which, in their magnetic polarity chronology, correlated with Chron C23N.

There are few available radiometric dates that are directly relevant to estimating the age of the Paleocene–Eocene boundary. Hardenbol & Berggren (1978) and Berggren (1971, 1972) relied heavily on two glauconite dates, a K-Ar date of 52.0 Ma from the Bashi Marl of the Gulf Coast of North America (stated to be basal Eocene = *Globorotalia rex* Zone of Bolli and assigned to the *Tribrachiatus contortus* (NP10) Zone (Bybell 1980; Gibson & Bybell 1981) but more probably assignable to Zone NP11; see discussion above) and a Rb-Sr date of 53.6 ± 2.5 Ma from presumed Thanetian (but probably 'Sparnacian' — see above — Sables de Bracheux; Pomerol 1973; Curry *et al.* 1978) sediments at Butte de Reneuil, France. Berggren *et al.* (1978: 74) recalculated the date from Butte de Reneuil to be 59.2 Ma based on a change in the presumed initial ratio of $^{87}Sr/^{86}Sr$ in early Tertiary sea water.

Several other dates must also be considered here. Four K-Ar dates on two glauconite horizons overlying fossiliferous horizons of the late Paleocene or early Eocene Ewekoro Formation, Nigeria, were originally (Adegoke *et al.* 1972) reported as an averaged date of 54.45 ± 2.7 Ma. Correcting for the 1976 IUGS K-Ar decay and abundance constants yields an average age for these four dates of 55.85 (= 55.9) Ma. A more precise assignment of the biostratigraphic age of these sediments is not possible at present, and therefore this date gives only an approximation of the age of the Paleocene–Eocene boundary.

Rubinstein & Gabunya (1978: 209) cite earlier studies of theirs as the basis for assigning an approximate age of 57 Ma (using the old Western decay constant of $\lambda_K = 0.584 \times 10^{-10}$ yr^{-1}) to the Paleocene–Eocene boundary. This age estimate was originally accepted 'with a considerable degree of un-

certainty' (Rubinstein & Gabunya 1978: 209). If this age estimate is a correct approximation for the Paleocene–Eocene boundary, recalibration of the 57 Ma age would yield an estimate of 58–58.5 Ma.

The widely developed ash series in the central part of the North Sea and parts of NW Europe is related to the extensive volcanism (= episode 7 of 'enhanced magmatic activity' of Fitch *et al.* 1978) in East Greenland around 57–54 Ma (ICC; cf. Soper *et al.* 1976a, b). Dates on the Blosseville Group volcanics in East Greenland may provide the best approximation of the age of the Paleocene–Eocene boundary. Sediments containing dinoflagellate floras bracket the radiometrically dated extrusives and can be directly correlated with the standard early Tertiary sections in NW Europe.

The *Dracodinium varielongituda* Zone sediments of the Kap Dalton Formation are not directly intercalated with, but rather, overlie and are separated from the top of the Blosseville Group, by an unknown temporal hiatus. This zone can therefore only represent the extreme maximum constraint on the younger age limit of the basalts.

Beckinsale *et al.* (1970) originally dated the Blosseville extrusives and tentatively concluded that they were between 55–50 million years old. However, the whole-rock K-Ar dates on the basalts ranged from 33–60 Ma and many samples showed evidence of alteration and presumed argon loss. Preliminary K-Ar determinations on the East Greenland Tertiary basalts by Hailwood *et al.* (1973) ranged between 47–52 Ma. Seven sampling sites on fresh material throughout the 2800 m basalt sequence were dated with good repeatability, but no experimental data were presented. Soper *et al.* (1976a) reported a refinement (based on Dr J. G. Mitchell, pers. comm.) of the original basalt data range of 47–52 Ma of Hailwood *et al.* (1973) to 48–49 Ma. This is consistent with an age reflecting regional thermal overprinting at 49–50 Ma as proposed by Fitch *et al.* (1978).

Fitch *et al.* (1978) re-evaluated the data of Beckinsale *et al.* (1970) by use of K-Ar correlation diagrams (regression analysis on plots of $^{40}Ar/^{36}Ar$ vs. $^{40}K/^{36}Ar$). Their (Fitch *et al.* 1978) correlation diagram of all the conventional K-Ar data of Beckinsale *et al.* (1970) on the East Greenland basalts showed a best-fit regression apparent age of 51 ± 3 Ma (= 52.3 ± 3 Ma) for these data. The scatter of the data around the regression line and the low $^{40}Ar/^{36}Ar$ intercept value for this line was interpreted as evidence of argon loss in these samples reflecting a regional thermal overprinting event at around 50–49 (52–50) Ma. Analysis of various subsamples of the Beckinsale *et al.* (1970) data produced approximately the same apparent age. However, correlation diagram analysis of data from the fine-grained upper and lower margins of a single basalt flow at Kap Brewster differed in having a best-fit regression line age of 54.5 ± 1.0 Ma (= 55.9 ± 1.0 Ma). Fitch *et al.* (1978) only used the data from samples EG 7147, 7150, and 7151 of Beckinsale *et al.* (1970, Table 1) at the outer margin of the flow. These samples were believed to give good approximations to the true age of the extrusion because they were relatively unaltered and unaffected by argon loss. The K-Ar dates for these samples given by Beckinsale *et al.* (1970) were 55.4 ± 3.1, 60.1 ± 2.8 (EG 7147); 57.8 ± 2.2 (EG 7150) and 56.1 ± 1.6 (EG 7151) Ma. But the correlation diagram of Fitch *et al.* (1978, Fig. 4) indicated a younger age of 54.5 (55.9) Ma due to the presence of previously unrecognized initial argon in these samples. A re-evaluation of the analysis made by Fitch *et al.* (1978) suggests an age of 56.5 Ma for the Kap Brewster flow (see

Appendix 1). The concordance of an apatite fission track age of 58.0 ± 2.8 Ma and a hornblende K-Ar age of 54.9 ± 1.6 Ma (both dates cited in Gleadow & Brooks 1979) on nepheline syenite from Nagtivit in the mouth of Sermilik Fjord, Angmagssalik district are further evidence for an early age for Tertiary igneous activity in East Greenland.

To what biostratigraphic and palaeomagnetic intervals is this age estimate applicable? Part of the answer has been suggested above and we shall now pursue the problem further.

Beckinsale et al. (1970: 31) state that '...it has not been possible to evaluate directly the palaeontological evidence from Kap Brewster and Kap Dalton, since faults of unknown displacement separate the fossiliferous sediments from the main basalt areas where suitable samples (for K-Ar dating) were found. However, both the dated basalts and the sediments are believed to represent very nearly the top of the pile...'. More detailed and refined studies by later investigators (Soper et al. 1976b) have shown that there are two important floral horizons associated with the basalt pile of the Blosseville Group. The lower horizon is within thin tuffaceous shales about 100 m above the base of the approximately 520 m thick Vandfaldsdalen Formation, in the Ryberg Fjord (Kangerdluggsuaq area), located near the base of, and within, the main body of the Blosseville Group. This horizon contains a small dinoflagellate flora referable to the *Apectodinium hyperacanthum* Zone (Soper *et al.* 1976b), which we have seen above spans the latest Paleocene—earliest Eocene of NW Europe and is equivalent to the later part of Zone NP9.

The upper fossiliferous horizon is within a shale about 300 m from the top of the Blosseville Group basalt pile at Kap Dalton with a rich dinocyst flora indicative of the *W. meckelfeldensis* Zone (Soper *et al.* 1976b). This zone also occurs in the lower part of the London Clay Formation (~ 5–18 m above the base of the London Clay at Herne Bay, London Basin), in the basal Ypresian of Belgium (~ 3 m above the base of the Argile d'Ypres) and at the top of lower Eocene 1 and the lower part of the lower Eocene 2 of NW Germany and is equivalent to Zone NP11 (at least in part; see above).

Although the dated basalt from Kap Brewster does not directly underlie or overlie sediments containing these assemblages, the *A. hyperacanthum* Zone sediments are found within, and just above the base of the main body of the Blosseville Group basalts and clearly would provide a lower biostratigraphic (and maximum age) limit for the basalt data of *c.*56.5 Ma cited above. The *W. meckelfeldensis* Zone horizon at Kap Dalton lies within 300 m of the top of the Blosseville Group basalts and is probably contemporaneous with, or younger than, the dated Kap Brewster basalt from 'very nearly the top of the pile ...' (Beckinsale *et al.* 1970: 31). However, the precise determination of the stratigraphic position of the *W. meckelfeldensis* Zone horizon and the Kap Brewster basalt does not preclude the possibility that the basalt is slightly younger than the *W. meckelfeldensis* Zone sediments.

Soper *et al.* (1976b) discuss a dinoflagellate flora from the Kap Dalton Formation, which concordantly overlies the Blosseville Group basalt pile. They assigned the assemblage to the *D. varielongituda* Zone (Ypresian) based on the presence of *Deflandrea wardenensis* and *Wetzeliella lunaris*. The *D. varielongituda* Zone is also found in the upper part of the London Clay, the upper part of the Argile d'Ypres (King

suggests an age of 56.5 Ma for the Kap Brewster flow (see 1981), and in the upper part of the Formation de Varengeville (Châteauneuf & Gruas-Cavagnetto 1978) and possibly in the lower part of the lower Eocene 3 (Costa & Downie 1976) and is correlative with Zone NP12 (Costa & Müller 1978). Sediments of the *D. varielongituda* Zone in NW Europe are uniformly situated above beds with tuffaceous layers in them. The *D. varielongituda* Zone flora overlying the basalts provides an upper biostratigraphic age limit for the 56.5 Ma basalt dates, and for the cessation of extrusive igneous activity in this area. Soper *et al.* (1976b) suggested a correlation of the beginning of the basalt eruption in East Greenland with the *A. hyperacanthum* Zone (correlated here with zones P5, P6a and lower part of P6b) and the end of the extrusion of the basalt pile during the *W. meckelfeldensis* Zone (correlated with Zone P6b). Together these two zones constrain the duration of volcanism to zones NP9 (*partim*) and NP10 essentially and latest P5–P6. Hailwood *et al.* (1973) preferred to use the broader constraints implied by the presence of the *D. varielongituda* Zone overlying the basalts to extend the end of volcanism to possibly as young as the end of NP11.

Available palaeomagnetic evidence supports the assignment of the Blosseville Group basalts to an NP9–10 correlative. Nielsen *et al.* (1981) have summarized palaeomagnetic results from East Greenland and presented detailed studies of previously unsampled sections of the Blosseville Group. These results indicate that all of the sampled intervals of the Blosseville Group were deposited during a reversed polarity field. However, the studies to date have sampled only the basal parts (at Kangerdluggsuaq) and the (presumed) upper parts (in the Scoresby Sound area) of the basalt sequence. Based on the presence of anomaly 24 as the oldest marine magnetic anomaly off the south-west coast of Greenland, previous workers (Hailwood *et al.* 1979) have assumed that the basalt pile on East Greenland must have predated rifting and therefore be older than anomaly 24 time. The thick reversed polarity sequence observed in East Greenland has been interpreted as representing the single reversed polarity interval Chron C24R (e.g. Soper *et al.* 1976a; Faller 1975) or possibly Chrons C24R, C25N, and C25R, should the unsampled middle portion of the Blosseville basalts contain normally magnetized rocks (e.g. Nielsen *et al.* 1981). In either case the radiometrically dated basalt from the upper part of the Blosseville Group must lie within the reversed polarity interval of Chron C24R.

Additional data on the relationship between the estimated age of the East Greenland volcanism and the biostratigraphic position of the Paleocene—Eocene boundary comes from a consideration of regional palaeomagnetic and biostratigraphic studies from the nearby Deep Sea Drilling Project (DSDP) sites in the NE Atlantic, North Sea and adjacent NW Europen stratigraphic sections and Mediterranean region.

The thick tuffaceous layers intercalated in the basal marine sediments (in dinocyst zones Ia (= *hyperacanthum*) and Ib (= *astra* and *meckelfeldensis*) at DSDP sites 403 and 404 are a direct reflection of the major magmatic event which resulted from the reversely magnetized flood basalts of East Greenland which lay immediately adjacent to the margin of Rockall Plateau (pre-anomaly 24) prior to the initiation of the seafloor spreading in the NE Atlantic (Hailwood 1979: 329). The entire basalt pile was probably erupted within a maximum time interval of *c.*3 Ma (~ 54–57 Ma with a best estimate of about 56.5 Ma for the Blosseville Group at

Kap Brewster). The basalts are bracketed by the *hyperacanthum* Zone (near the base), the *meckelfeldensis* Zone (within, near the top) and the *varielongituda* Zone (above the basalts) as we have seen above.

The main tuff falls in NW Europe are within the later part of the *hyperacanthum* Zone (= Sparnacian; Knox & Morton 1983). The volcanism would appear to have terminated within the early Ypresian inasmuch as red shales in North Sea wells containing the youngest ash levels are characterized by an acme occurrence of *Subbotina patagonica* and small acarininids (= the planktonic foraminiferal horizon seen in NW European mid-Ypresian sections; see above) and an NP11 nannoflora (Berggren & Aubry, pers. obs.) equivalent to the ash-bearing lower Rosnaes Clay of Denmark with a P7 planktonic foraminiferal fauna (Berggren 1960) and an NP11 flora (Thiede *et al.* 1980). The same relationship has been seen on SW Rockall Bank (sites 403, 404; Müller 1979). Early Eocene volcanic tuffs reported recently from dredge hauls in Rockall Trough were said to reflect explosive volcanicity in the vicinity of the Wyville-Thompson Ridge (Faeroese Province) (Jones & Ramsay 1982). The volcanic tuffs were dated as belonging to the *Marthasterites tribrachiatus* Zone (= NP12) but the floral list and the absence of *Discoaster lodoensis* do not preclude assignment to Zone NP11 (M-P. Aubry, pers. comm.).

Normally magnetized sediments have been correlated to the time represented by anomaly 24A at DSDP site 404 (SW Rockall Plateau) at a stratigraphic level near the base of Zone NP12 and dinocyst Zone II (= *W. varielongituda* Zone of NE Europe) (Hailwood 1979: 329; Hailwood *et al.* 1979: 1130, 1131).

Additional calibration comes from the Contessa Road section (Gubbio, Italy) where the FAD of *Discoaster lodoensis* (= NP12) is located at the base of Anomaly 24A correlative and the FAD of *Tribrachiatus orthostylus* (~ NP11) is located just below the base of Anomaly 24B correlative (Perch-Nielsen *In*: Lowrie *et al.* 1982). Since the East Greenland basalt pile is assumed to be older than Chron C24N, the base of NP11 may be the minimum age bracket for the top of these basalts. Finally, the LAD of *Fasciculithus*, which occurs within Zone NP10 (Romein 1979: 77) but is generally considered to represent the NP9–NP10 boundary and the Paleocene–Eocene boundary by most micropalaeontologists has been identified at a stratigraphic position approximately midway in the long reversely magnetized interval of Chron C24R in the South Atlantic (DSDP sites 527 and 528) and in the Contessa Road section (Perch-Nielsen *In*: Lowrie *et al.* 1982; Shackleton 1983).

Where, then, are we to locate the Paleocene–Eocene boundary. The data summarized above suggest that it lies:
1. between the *hyperacanthum* (Sparnacian) and *astra* (Ypresian) dinocyst Zone
2. between calcareous nannoplankton zones NP9 (Thanetian) and NP10 (Ypresian) or within Zone NP10
3. between zones P6a and P6b (planktonic foraminifera)

Further, the Paleocene–Eocene boundary lies within the lower part of Chron C24R and is seen to be situated biostratigraphically within the tuff series of the North Sea and NW Europe (and thus within the basalt extrusion of the Blosseville Group) which essentially spans the *hyperacanthum-astra* zonal intervals and (locally) extends into the *meckelfeldensis* Zone. The best age estimates for this boundary (based on the East Greenland dates) would appear to be about 56.5 Ma and our estimate of 57.8 Ma (based on our newly constructed palaeomagnetic chronology) is seen to lie reasonably close to the radiometric dates.

Finally, if the unconformity between the Oldhaven and London Clay formations is an expression of a global relative eustatic sea-level fall, this unconformity is seen to lie within the latest part of the *hyperacanthum* Zone and to have been of very short duration (< 1 m.y.).

Deposition of the Bracklesham Formation in the Hampshire Basin has been interpreted in terms of five transgressive cycles which were thought to be the small scale (epicontinental) reflection of eustatic sea-level rises superimposed upon a major transgressive cycle that began after a major regression at the top of the London Clay (Plint 1983). This latter regression was correlated, in turn, with the major late early Eocene eustatic sea-level fall (TE1.2/TE2.1) of Vail *et al.* (1977). We note in passing that this regression is probably incorrectly correlated with the Vail *et al.* (1977) cycle sequence, because that boundary occurs within Zone NP13, whereas the London Clay–Bracklesham boundary occurs within the upper part of Zone NP11 or lower part of Zone NP12 (Aubry 1983). It is more likely that the regression that Plint (1983: 647) is referring to at the top of the London Clay corresponds to the TE1.1–TE1.2 cycle boundary of Vail *et al.* (1977) which occurs within Zone P7 and NP12. The major eustatic cycle (TE1.2–TE2.1) boundary of the late early Eocene has been suggested to lie within the upper part of the Wittering division and to represent a marine/non-marine paraconformity at Whitecliff Bay (Aubry 1983) which can be seen to correspond to cycle 2 of Plint (1983: 645, Fig. 15).

Butler *et al.* (1981b) and Rapp *et al.* (1983) have constructed magnetic polarity stratigraphies for Tiffanian to Wasatchian strata in the Clark's Fork Basin, Wyoming and Big Bend area, Texas, respectively (see discussion above). The Clarkforkian extends from the lower or middle part of a normal polarity interval that Butler *et al.* (1981b) correlate with Chron C25N to approximately the middle of a reversed polarity interval correlated with Chron C24R (Butler *et al.* 1981b; Rapp *et al.* 1983).

The studies by Butler *et al.* (1981b) and Rapp *et al.* (1983) raise an interesting problem with the placement of the Paleocene–Eocene boundary relative to biochronologic and magnetic polarity zonations. Butler *et al.* (1981b) follow Gingerich (1976, 1980) and Rose (1980), placing the Paleocene–Eocene boundary at the base of the European Sparnacian (represented by the fauna of the Conglomérat de Meudon, France). Basal Sparnacian is equated with the *Plesiadapis cookei* biochron (Cf₂) of Rose (1980), and therefore the Paleocene–Eocene boundary is located between lower and middle Clarkforkian (Cf₁/Cf₂) boundary. Butler *et al.* (1981b) place the Paleocene–Eocene boundary within the basal portion of Chron C24R (*contra* the younger placement of this boundary within the anomaly time-scales of LaBrecque *et al.* 1977, and Ness *et al.* 1980). The precise placement of the Paleocene–Eocene boundary relative to various biostratigraphic zonations has been a subject of much recent controversy (see above). As mentioned above, Costa *et al.* (1978) and King (1981) place the Paleocene–Eocene boundary at the *Apectodinium hyperacanthum–Wetzeliella* astra dinoflagellate zonal boundary. Accepting this definition of the Paleocene–Eocene boundary results in most, or all, of the Sparnacian being late Paleocene rather than early Eocene (see above). This boundary in North America then falls later in the Clarkforkian than has previously been recognized
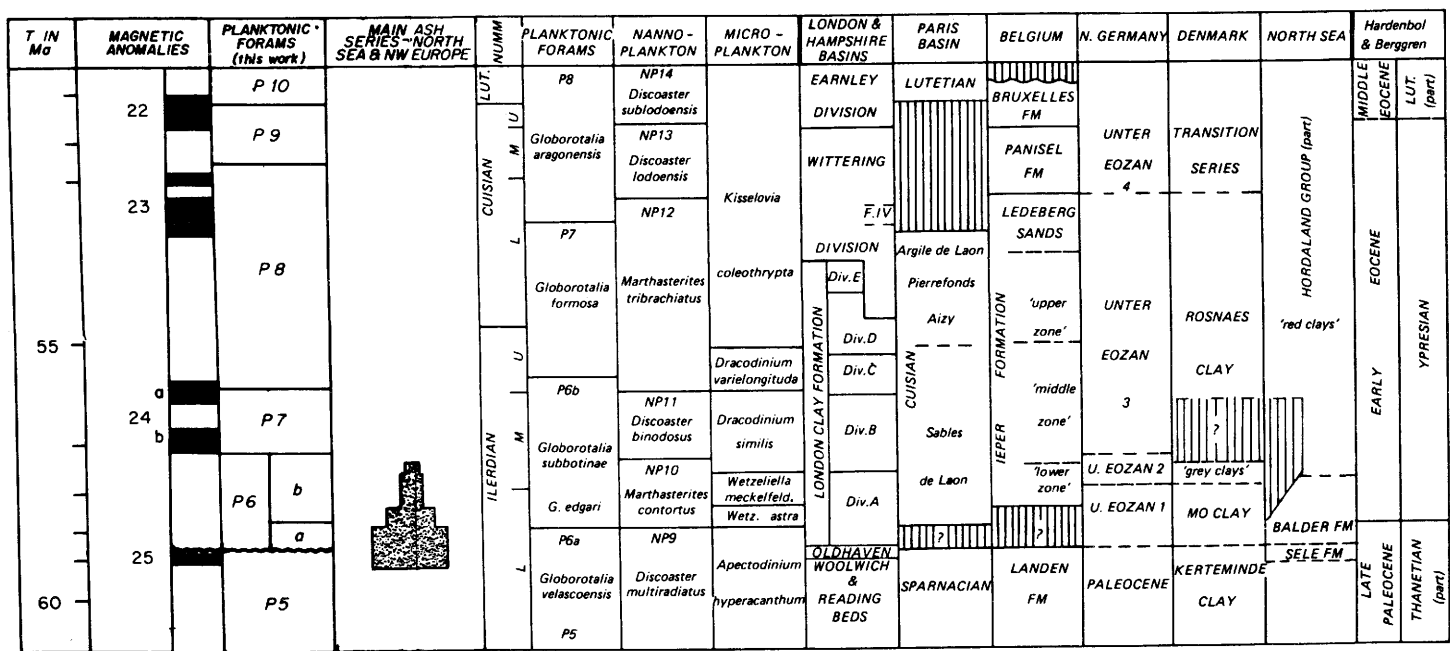
T IN Ma	MAGNETIC ANOMALIES	PLANKTONIC FORAMS (this work)	MAIN ASH SERIES—NORTH SEA & NW EUROPE	NUMM	PLANKTONIC FORAMS	NANNO-PLANKTON	MICRO-PLANKTON	LONDON & HAMPSHIRE BASINS	PARIS BASIN	BELGIUM	N. GERMANY	DENMARK	NORTH SEA	Hardenbol & Berggren
22		P 10		LUT. U	P8	NP14 Discoaster sublodoensis		EARNLEY DIVISION	LUTETIAN	BRUXELLES FM			HORDALAND GROUP (part)	MIDDLE EOCENE / LUT. (part)
		P 9		CUISIAN M	Globorotalia aragonensis	NP13 Discoaster lodoensis	Kisselovia	WITTERING		PANISEL FM	UNTER EOZAN 4	TRANSITION SERIES		EOCENE
23		P 8		CUISIAN L	P7 Globorotalia formosa	NP12 Marthasterites tribrachiatus	coleothrypta	F.IV DIVISION	Argile de Laon / Pierrefonds	LEDEBERG SANDS				YPRESIAN
55								Div.E	Aizy	'upper zone' (IEPER FORMATION)	UNTER EOZAN	ROSNAES CLAY	'red clays'	EARLY
	a			CUISIAN U		Dracodinium varielongituda		Div.D / Div.C						
24	b	P 7		ILERDIAN M	P6b Globorotalia subbotinae	NP11 Discoaster binodosus	Dracodinium similis	Div.B (LONDON CLAY FORMATION)	Sables de Laon	'middle zone'	UNTER EOZAN 3		'grey clays'	
		P6 b			NP10 Marthasterites contortus	Wetzeliella meckelfeld. / Wetz. astra		Div.A		'lower zone'	U. EOZAN 2	MO CLAY	BALDER FM	
25		a		ILERDIAN L	G. edgari						U. EOZAN 1		SELE FM	LATE PALEOCENE
		P6a a			P6a Globorotalia velascoensis	NP9 Discoaster multiradiatus	Apectodinium hyperacanthum	OLDHAVEN WOOLWICH & READING BEDS	SPARNACIAN	LANDEN FM	PALEOCENE	KERTEMINDE CLAY		THANETIAN (part)
60		P 5			P5									

FIG. 4. Correlation of uppermost Paleocene and lowermost Eocene stratigraphic succession in northwest Europe (modified from King 1981, text-Fig. 52). In the left hand side of the figure we show the relationship between planktonic foraminiferal zones to the magnetobiochronologic scale developed in this paper and the main ash series of NW Europe is placed in its biostratigraphic and chronologic framework. The numerical scale and the magnetic polarity anomaly scale are not plotted in a linear manner because they are calibrated to the chronostratigraphic columns shown in the middle (NUMM.) and on the right (Hardenbol & Berggren). A hiatus is shown just above anomaly 25 correlative, corresponding to that which separates the Oldhaven and London Clay formations. Thus the Paleocene–Eocene boundary is interpreted to lie at some distance (temporally and spatially) above anomaly 25 time and sediments.

(Butler *et al.* 1981b; Gingerich 1976, 1980; Rose 1980), or even within the early Wasatchian.

Marine micropalaeontologists and palaeomagnetists locate the Paleocene–Eocene boundary within the reversed interval of Chron C24R. Identification of the normal polarity intervals in the Clark's Fork Basin sequence (Butler *et al.* 1981b) as correlatives of Chrons C25N and C26N, and in the Big Bend area (Rapp *et al.* 1983) section as correlatives of Chrons C24N, C25N and C26N is consistent with placement of the Paleocene–Eocene boundary in the later Clarkforkian or early Wasatchian. This recognition of the Paleocene–Eocene boundary higher within Chron C24R than is indicated by Butler *et al.* (1981b) is also consistent with the placement of this boundary in marine stratigraphic sections.

The Clarkforkian–Wasatchian boundary in the Clark's Fork Basin (Butler *et al.* 1981b) and Big Bend area (Rapp *et al.* 1983) would therefore lie within Chron C24R, and early Wasatchian faunas would correlate with the later portion of this reversed interval (see Figs 3 and 5).

The Eocene

Early Eocene geochronology and chronostratigraphy have been discussed in the preceding section on the Paleocene–Eocene boundary. We begin our discussion of the Eocene with the Middle Eocene.

The early–middle Eocene (Ypresian–Lutetian Age) boundary is recognized (i.e. correlated) by most marine micropalaeontologists at the base of the *Hantkenina aragonensis* (P10) planktonic foraminiferal zone. This boundary has been variably correlated with either the base (Hardenbol & Berggren 1978; Poore 1980) or the middle (Kleinpell *et al.*

1980) of the *Discoaster sublodoensis* (NP14) calcareous nannofossil zone. Current investigations by Aubry (1983) on the Paleogene stratotype sections of the Paris and Hampshire–London Basins are pertinent in this connection. The base of the stratotype Lutetian Stage is within Zone NP14 (based on the presence of *Discoaster sublodoensis*) and extends upward into levels within Zone NP15 (with *Nannotetrina alata*). The upper part of the Lutetian is characterized by shallow water (brackish to lacustrine) limestones in which nannoplankton are absent. The uppermost Lutetian (= shallow water calcarenites with *Discorinopsis kerfornei* and *Linderina* sp. = Biarritzian facies) contains a nannofossil flora comparable to that found in the uppermost Bracklesham Beds of the Hampshire Basin in England, assigned to Zone NP16. Thus the Lutetian, as stratotypified in the Paris Basin, essentially spans the NP14–NP16 (*partim*) zonal interval. These results differ significantly from those of Bigg (1982). However, Aubry (1983) presents a detailed critique of Bigg's results which would appear to be due primarily to an overestimation of the role of reworking and inadequate evaluation of total nanofloral associations.

The succeeding Bartonian Stage (Barton Beds of England) is equivalent to Zone NP16 (*partim*) and NP17 (Cavelier & Pomerol 1976; Hardenbol & Berggren 1978; Aubry 1983).

If the unconformity which marks the Cuisian = Ypresian/Lutetian boundary in the Belgian-Paris basin(s) corresponds to, and is a reflection of, the eustatic sea-level lowering (between cycles TE 1.2 and TE 2.1) which lies within Zone P9 and NP13 (Vail *et al.* 1977) and has been recognized at correlative levels in California (Berggren & Aubert, 1983) and Cyrenaica, Libya (Barr & Berggren 1981), then it would appear that the base of the Lutetian probably lies close to the NP13–14 boundary (it has not yet been

possible to recognize the lithostratigraphic-palaeooceano-
graphic expression of the eustatic sea-level fall in deep sea
(i.e. bathyal) deposits and trace this level into the uncon-
formity as expressed in outcrop) and this is probably the best
estimate that can be made for the biostratigraphic position of
the Ypresian–Lutetian boundary, essentially supporting
earlier correlations suggested by Hardenbol & Berggren
(1978).

Studies at Gubbio, Italy (Lowrie & Alvarez 1981; Lowrie *et
al.* 1982), in which planktonic foraminifera and calcareous
nannoplankton zonations are directly associated with the
magnetic polarity stratigraphy, locate the early–middle
Eocene boundary, as determined by the FAD of *Hantkenina*,
just below the top of Chron C22N. The *Hantkenina
aragonensis* (P10) Zone spans the interval represented by the
very youngest part of Chron C22N to all but the latest part of
Chron C21N.

The placement of Zone NP14 is less precisely controlled at
Gubbio, and indeed, elsewhere. A survey of published data
reveals some variation in the placement of zonal boundaries
(and resulting correlations) over the interval of Zones P9–11
and NP13–15. The FAD of *Hantkenina* has been variously
placed within Zone NP14 (DSDP Sites 366, 405, 506), at the
NP14–15 boundary (DSDP Sites 384, 401) or even within
Zone NP15 (DSDP Site 356). It has been even recorded in
the later part of Zone NP15, near the NP15–16 boundary
(Toumarkine & Bolli 1975; Proto-Decima *et al.* 1975) in the
Possagno section of northern Italy, but this is clearly a
delayed entry and not a true FAD.

The FAD of *Hantkenina* has been shown to occur within
the youngest part of Chron C22N in the Contessa Highway
section, near Gubbio, Italy (Lowrie *et al.* 1982), whereas
recent, as yet unpublished data from the North Atlantic
indicate that the NP13–14 boundary is located in the earliest
part of Chron C22N.

Thus the FAD of *Hantkenina* occurs within Zone NP14,
approximately 1 m.y. later than the NP13–14 boundary. The
early–middle Eocene boundary, as determined by the FAD
of *Hantkenina*, is located in the latest part of Chron C22N
with an estimated (magnetochronologic) age of 52 Ma.

Evidence from DSDP Leg 73 (Poore *et al.* 1983) supports
these biostratigraphic–magnetostratigraphic associations.
Nannoplankton floral zonations from Site 523 associate Zone
NP15 with most, or all, of Chron C20. Much of Chron C20R
appears to be represented at Site 523 (although neither the
top of the anomaly 21 correlative nor the base of Zone NP15
is present because of missing section at the bottom of this
site), and it is associated only with Zone NP15. Therefore,
the top of Zone NP14 is probably closely associated with
Chron C21N.

This is supported by the record of *Nannotetrina fulgens* (*N.
quadrata = N. alata*), nominate taxon of Zone NP15, as low
as the lower part of Chron C20N in the Contessa Quarry
section, Gubbio, Italy (Lowrie *et al.* 1982) and of *Nanno-
tetrina* sp. as low as the lower part of anomaly 21 cor-
relative in the Contessa Road section (*op. cit.*) which sug-
gested tentative correlation of the base of Zone NP15 at least
as low as lower Chron C21N (Lowrie *et al.* 1982).

Berggren *et al.* (1978) summarized evidence that the
Ardath Shale of the La Jolla Group, San Diego, California
contains calcareous nannoplankton floras assignable to the
Rhabdosphaera inflata Subzone of the *Discoaster sub-
lodoensis* (NP14) Zone (Bukry & Kennedy 1969; Bukry 1973;
Bukry 1980) and planktonic foraminiferal faunas correlative

with the *Hantkenina aragonensis* (P10) and/or *Globigerapsis
kugleri* (P11) Zone (Gibson 1971; Steineck & Gibson 1971;
Gibson & Steineck 1972; Steineck *et al.* 1972).

Based on benthonic foraminiferal assemblages Phillips
(1972) and Mallory (1959) interpreted the Ulatisian–Narizian
Stage boundary (of the California Paleogene benthonic
foraminiferal zonation) to fall within the Rose Canyon Shale
of Milow & Ennis (1961; equivalent to the Ardath Shale of
Kennedy & Moore 1971, and Kennedy & Peterson 1975).
However, Gibson (1971) and Gibson & Steineck (1972,
p. 2226) assigned the Ardath Shale entirely to the *Amphi-
morphina californica* Zone of the Ulatisian Stage. Poore
(1980) has shown that the Ulatisian–Narizian boundary
approximately coincides with the *Discoaster sublodoensis–
Nannotetrina quadrata* (NP14–15) Zone boundary. The
Ulatisian–Narizian boundary and NP14–15 boundary would
therefore lie within, or slightly above the top of, the Ardath
Shale.

Since the Ardath Shale contains floras assigned to upper
NP14 and faunas correlated with P10 and/or P11, the
early–middle Eocene boundary (base of P10) *must* lie within
or below the Ardath Shale in this area.

The Friars Formation of the La Jolla Group and Mission
Valley Formation of the Poway Group contain an abundant
early Uintan land mammal fauna (Golz 1973; Golz &
Lillegraven 1977). Based on stratigraphic relationships the
Friars Formation is partly time correlative with, and younger
than, the Ardath Shale, while the Mission Valley Formation
is entirely younger than the Ardath Shale. The early Uintan
fauna is possibly partly contemporaneous with, but most
likely entirely younger than, the Ardath Shale marine faunas
and floras.

Golz & Lillegraven (1977) and Berggren *et al.* (1978) have
assigned the mammalian faunas from the Friars and Mission
Valley Formations to an early Uintan age. Golz & Lillegraven
(1977, p. 44) stated that the San Diego fauna was more
primitive than other Uintan faunas from California (Laguna
Riviera, Camp San Onofre and Ventura County localities)
and '... from most of the Rocky Mountain Uintan sites'. The
Friars and Mission Valley Formation fauna was believed to
be younger than standard Bridgerian faunas and older than
standard Uintan faunas previously described from the Rocky
Mountain region, occupying a temporal position somewhere
intermediate between previously defined Bridgerian and
Uintan. Faunas from the type section of the Tepee Trail
Formation, Fremont Co., Wyoming and 'Pruett Tuff', Agua
Fria Area, Brewster Co., Texas are likely temporal cor-
relatives of this earliest Uintan San Diego fauna (see Golz &
Lillegraven 1977; Berggren *et al.* 1978; Wilson 1980).
Microfaunal localities recently collected by W. Turnbull in
the Adobe Town Member, Washakie Formation, Sweetwater
Co., Wyoming lie stratigraphically between well known
Bridgerian and Uintan faunas, and may occupy a temporal
position similar to the faunas mentioned above.

Studies of the palaeomagnetic stratigraphy of the
Bridgerian to Uintan sections in these four areas are currently
in progress (Flynn 1983a, b), and some of these results will be
preliminarily discussed here. Palaeomagnetic results from the
La Jolla Group, San Diego indicate that the entire type
section of the Ardath Shale was deposited in a normal
polarity interval (except for a thin reversed polarity horizon
in the upper third of the section). The Delmar Formation is
one of the oldest units in the La Jolla Group and it generally
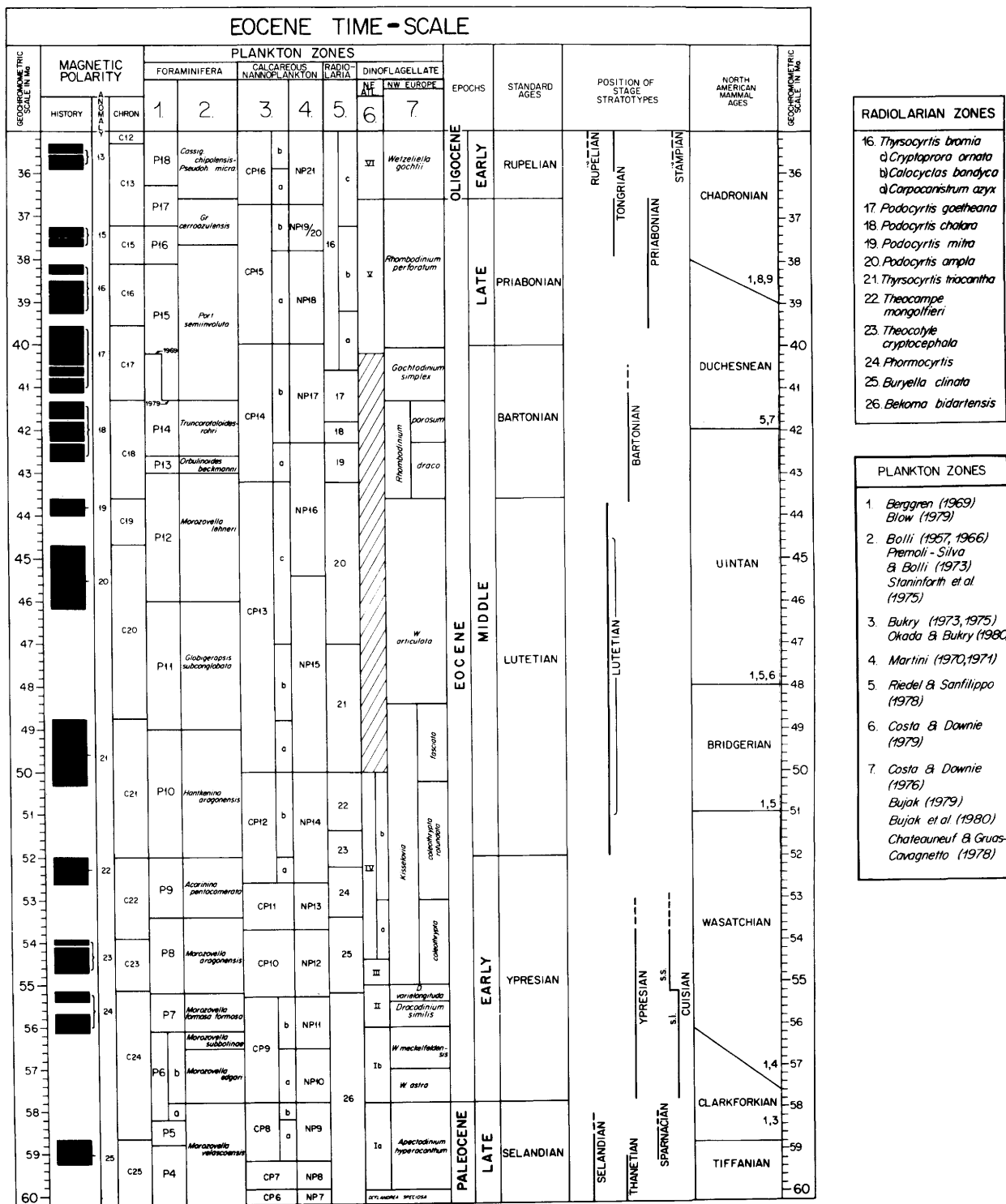lies stratigraphically below both the Torrey Sandstone and

FIG. 5.   Eocene geochronology (explanation as in Fig. 3).

Ardath Shale. At its type section the Delmar Formation was deposited during an interval of reversed polarity, while near the gradational contact with the Torrey Sandstone above, the sediments were deposited in a normal polarity field. A section of mollusc-bearing Scripps Formation and overlying early Uintan mammal-bearing Friars Formation produced a palaeomagnetic pattern of normal polarity at the top of the Scripps Formation and base of the Friars Formation, and reversed polarity to the local top of the Friars Formation.

Based on the biostratigraphic correlation of the P10 and/or P11 and upper NP14 (possibly close to the NP14–15 boundary) Zones in the Ardath Shale to the Gubbio and DSDP Leg 73 sections the normal polarity interval in the Ardath Shale represents Chron C21N. The Delmar Formation reversed section correlates with Chron C21R and the Friars Formation reversed sequence represents Chron C20R. Since the base of Zone P10 only barely falls within the top of Chron C22N, and the Ardath Shale biostratigraphic information indicates an age younger than the extreme base of P10 for the entire normally magnetized thickness of the Ardath Shale type section, it is almost certain that this normal polarity interval can only be correlated with Chron C21N. The early Uintan mammal fauna of the Friars Formation would lie within the reversed interval just older

than Chron C20N and early Uintan would therefore be temporally correlative with at least part of Chron C20R (Fig. 5).

In NW Wyoming the upper part of the type section of the Tepee Trail Formation preserves a diverse early Uintan mammal fauna (Berggren *et al.* 1978; McKenna 1980), while the underlying Aycross Formation contains a Bridgerian mammal fauna. The Tepee Trail Formation is almost completely reversely magnetized, except for a relatively thin normal polarity interval at the base of the section. The Aycross Formation in its type area consists of a polarity sequence of reversed at its top, a long normal, a long reversed, a normal, and a reversed at its base. The Aycross and Tepee Trail Formations appear to be partial temporal equivalents (as is frequently encountered laterally in volcaniclastic terrains of this area, see Smedes & Prostka 1972), in which the base of the Tepee Trail Formation is correlative with the top portion of the Aycross Formation. The early Uintan Tepee Trail Formation fauna lies within a thick reversed interval, as does the temporally correlative early Uintan fauna from San Diego. Based on this correlation, the normal polarity interval at the base of the Tepee Trail Formation and near the top of the Aycross Formation must represent Chron C21N. The correlation of the normal polarity interval lower in the Aycross section is equivocal; the entire Aycross Formation normal intervals could represent Chron C21N with a very expanded short duration reversed event preserved between them (note the short reversed interval preserved in Chron C21N of the Ardath Shale, and Contessa Highway section of Lowrie *et al.* 1982), the lower normal could represent the preservation of a short normal event in Chron C21R, or the two normal polarity intervals could represent Chrons C21N and C22N. Although no definitive conclusion is presently possible, we believe the available palaeomagnetic pattern data in this section and radiometric data on Bridgerian sediments argue against interpreting the lower normal polarity interval as an anomaly 22 correlative. The Bridgerian–Uintan 'Land Mammal Age' boundary therefore lies within Chron C20R (Fig. 5).

Four published radiometric dates from these two sections (Smedes & Prostka 1972; Love *et al.* 1976) bracket the recognized polarity interval boundaries. A date of 50.5 ± 0.5 Ma lies within the Aycross Formation palaeomagnetic section, approximately 950′ below the top of the normal polarity interval correlated with Chron C21N. Dates of 47.9 ± 1.5 Ma and 48.3 ± 1.3 Ma (Mean = 48.1 Ma) lie within the Wiggins Formation, at a single horizon 500–600′ above the top of the Tepee Trail Formation palaeomagnetic section. Another date of 45.7 ± 1.2 Ma has been determined on a sample 650–750′ above the top of the Tepee Trail Formation in the same section. These horizons are approximately 1650′ and 1800′ above the top of the normal interval correlated to Chron C21N. Using a simple linear interpolation of age versus stratigraphic thickness between the mean 48.1 Ma and single 50.5 Ma dates results in an age estimate of 49.57 Ma for the top of Chron C21N correlative. An alternative linear interpolation between the 50.5 Ma date as one endpoint and the midpoint of the overlap in the error bars between the Wiggins Formation dates as the other endpoint, results in an age of 49.2 Ma for the top of Chron C21N correlative. An age range of between 49.2 and 49.6 Ma is therefore indicated for the younger boundary of anomaly 21 time. We favour an age estimate of approximately 49.5 Ma for this boundary.

Eight other high temperature K-Ar dates on sediments of

certain Bridgerian age from western Wyoming range from 49.0 to 50.3 Ma, supporting the age estimates for the top of Chron C21N (and late Bridgerian age) given above. Three other dates on samples from latest Wasatchian or early Bridgerian sediments range from 50.5 to 50.6 Ma, while five dates from sediments of Bridgerian or early Uintan age range from 46.6 (or 47.3) to 50.6 Ma. Interpretation of the Aycross Formation section as representing part of Chron C20R, all of Chron C21N, all of Chron C21R, all of Chron C22N, and part of Chron C22R (an interval of at least 4.5 m.y.) is difficult to reconcile with the short temporal duration indicated by the radiometric dates for Bridgerian time.

The Washakie Basin (Washakie Formation) palaeomagnetic section further supports the correlation of earliest Uintan, and the Bridgerian–Uintan boundary, within Chron C20R; classic Bridgerian and Uintan faunas fall within a long reversed interval and an overlying long normal interval of Chron C20, respectively.

Correlation of earliest Uintan faunas and the Bridgerian–Uintan boundary within a reversed polarity interval is consistently found in the Wyoming and California sections. Marine biostratigraphic correlations of the San Diego sections to standard sections at Gubbio, Italy and in the deep sea South Atlantic indicate that this reversed interval is correlative with Chron C20R, while the immediately underlying normal polarity interval represents Chron C21N. Radiometric dates bracketing the top of the Chron C21N correlative boundary in Wyoming provide an age estimate of 49.5 Ma for this boundary. This results in an age estimate of approximately 52.7 Ma for the top of Chron C22N (assuming a difference of approximately 3.2 Ma between the end of anomaly 22 and the end of anomaly 21, as is indicated in the spacings of the anomaly boundaries in the magnetic anomaly time-scales of LaBrecque *et al.* 1977, and Ness *et al.* 1980). As the work of Lowrie & Alvarez (1981) and Lowrie *et al.* (1982) indicates an association of the top of Chron C22N with the base of P10 (and, by assumed correlation, the early–middle Eocene boundary), an age estimate of 52.7 Ma can be made for the early–middle Eocene boundary, which is close to our magnetochronologic age estimate of 52 Ma and brings us full circle to the discussion at the beginning of this section.

Uintan faunas are well known from several areas of the United States (see West *et al.*, in press), but to date there have not been any magnetostratigraphic studies of middle to late Uintan strata, and isotopic dates from this interval are rare (see West *et al.*, in press). However, the magnetostratigraphy of strata of Bridgerian (or early Uintan) to Chadronian age from the 'Pruett' Formation, western Texas is currently under investigation by J. Flynn. Our tentative placement of the Uintan–Duchesnean boundary presently is based only on high temperature isotopic data from strata of Uintan and Duchesnean age (see West *et al.*, in press; Black 1969; McDowell *et al.* 1973). Correlation of this boundary, and all Eocene North American Land Mammal Ages, to the magnetic polarity time-scale is shown in Fig. 5.

The middle–late Eocene boundary (Bartonian–Priabonian Age boundary) is traditionally correlated with the P14–15 (*sensu* Blow 1969) and NP17–18 boundary by planktonic foraminiferal and calcareous nannoplankton biostratigraphers, respectively. However, attention is drawn to the fact that Blow (1979: 290–293) has emended the definition of his (1969) Zone P14 (*Truncorotaloides rohri–Globigerinita howei* Partial-range Zone) and renamed it the *Globorotalia (Morozovella) spinulosa spinulosa* Partial-range

Zone and emended the definition of his (renamed) Zone P15 (*Porticulasphaera semiinvoluta* Partial-range Zone).

The change in the nominate taxon for Zone P14 was made to emphasize the virtually simultaneous LAD and FAD of *Morozovella spinulosa* and *Porticulasphaera semiinvoluta*, respectively. The extinction of the *Truncorotaloides rohri* group (previously used to denote the P14–15 boundary) occurs within the range of *P. semiinvoluta*. In choosing what is generally regarded as an easily recognizable taxon with an apparently abrupt termination Blow (1979) has effectively shortened Zone P14 at the expense of P15 (see Blow 1979; Figs 58–61 for the relationship and historical changes of various zonal schemes during this interval).

Recent magnetobiostratigraphic studies on deep sea cores (Poore *et al.* 1982, 1983; Pujol 1983) and the Contessa (Lowrie *et al.* 1982) and Gubbio (Napoleone *et al.* 1983) sections in Italy have placed some constraints on the position of the middle–late Eocene boundary. The LAD of *Acarinina* and *Truncorotaloides* is associated with mid-Chron C17N (Poore *et al.* 1982, 1983; Napoleone *et al.* 1983), the LAD of *Morozovella spinulosa* is associated with the chron C17–C18 boundary (Pujol 1983) as is the FAD of *Porticulasphaera semiinvoltua* (Lowrie *et al.* 1982).

In terms of calcareous nannoplankton the NP17–18 boundary is traditionally placed at the FAD of *Chiasmolithus oamaruensis* or the LAD of *Chiasmolithus grandis*. Proto-Decima *et al.* (1975) have suggested a correlation of the *Chiasmolithus oamaruensis* (NP18) Zone with the *Truncorotaloides rohri* (approximately P14) Zone based on a study of the Possagno section, northern Italy, as well as comparative studies on samples from Trinidad and the Blake Plateau. However, this correlation leads to difficulties *vis a vis* magnetobiostratigraphic correlations and, indeed, the stratigraphic distribution of the calcareous nannoplankton in the Possagno section (Proto-Decima *et al.* 1975, Figs 1 and 2) indicate considerable reworking throughout the Eocene. Correlation of the *oamaruensis* Zone with Zone P15 is shown in DSDP Sites 363, 401, and 402, whereas at Sites 359 and 360 the *oamaruensis* Zone is correlated with the P15–16 interval. At DSDP Site 95 Zone P15 is correlated with Zone NP17 (*Discoaster barbadiensis*). In a recent study Verhallen & Romein (1983) suggest that the type Priabonian probably corresponds to the upper part of the *Isthmolithus recurvus* and *Sphenolithus pseudoradians* (*partim*) zones based on a study of the calcareous nannoplankton flora.

We point out here that the top of Bolli's (1966) *Truncorotaloides rohri* Zone was defined on the basis of the LAD of the nominate taxon, supposedly contiguous with the FAD of the nominate taxon of his succeeding (total range Zone) 'Globigerapsis semiinvoluta'. However, these two taxa overlap in deep sea sequences and the *Truncorotaloides rohri–Globigerapsis semiinvoluta* zonal boundary (*sensu* Bolli 1966, based on the LAD of *T. rohri* = P14–15 boundary of Blow 1969) would fall within Chron C17N, close to the level of the LAD of *Chiasmolithus grandis* (= NP17–18 boundary) as recorded by Poore *et al.* (1983). Thus the identification of the *C. oamaruensis* (NP18) Zone with (at least a part of) the *T. rohri* Zone by Proto-Decima *et al.* (1975) is understandable. That it probably does not correspond to the entire, or even a major part of, *T. rohri* Zone, however, is seen by the following.

The LAD of *Chiasmolithus solitus* (= NP16–17 boundary) occurs in the lower part of Chron C18N (Poore *et al.* 1983), a short distance above the LAD of *Porticulasphaera*

beckmanni (nominate taxon of Zone P13) in basal Chron C18N (Lowrie *et al.* 1982). The NP16–17 zonal boundary is thus within Zone P14. The LAD of *Chiasmolithus grandis*, which is commonly used to denote the NP17–18 boundary, is recorded in the later part of Chron C18N (Lowrie *et al.* 1982; Monechi & Thierstein, in press) but (together with the FAD of *Chiasmolithus oamaruensis*) in the later part of Chron C17N by Poore *et al.* (1983). If the former interpretation is accepted as definitive, it would have the effect of placing the NP17–18 boundary within the upper part of Zone P14 (and well down within the later part of the range of the nominate taxon *T. rohri*). If the latter interpretation is accepted as definitive the NP17–18 boundary is essentially correlative with the P14–15 boundary (*sensu* Blow 1969 = LAD *T. rohri*; i.e. within Zone P15 *sensu* Blow 1979).

We have chosen the latter interpretation and place the middle–late Eocene (= Bartonian–Priabonian) boundary at a level within the later part of Chron C17 = c.40.0 Ma (Fig. 5).

The Eocene–Oligocene boundary

There are a number of major changes that have long been recognized in marine and terrestrial faunas and floras at levels that coincide approximately with the classical position of the Eocene–Oligocene boundary (Cavelier *et al.* 1981; Van Couvering *et al.* 1981). These include:

1. the 'Grande Coupure' ('Big Break') in terrestrial vertebrate faunas (Stehlin 1909) between the late Eocene (Gypse de Montmartre in the Paris Basin) and the early Oligocene (e.g. Ronzon in the Haute-Loire, Soumailles in Lot-et-Garonne) which manifests itself in the relatively rapid but demonstrably time-transgressive appearance of some 10–13 new mammalian families that occurred when palaeogeographic conditions allowed North American and Asian mammals to cross shallow barriers (e.g. the Turgai Straits, south of the Urals and perhaps also the Beringia lowlands) into Europe. In England, this faunal break occurs between the Bembridge Limestone (*Ectropomys exiguus* Zone) and Hamstead Beds (*Eucricetodon atarus* Zone) in the Hampshire Basin;

2. a number of extinctions in the large benthic foraminifera (*i.al.*, *Nummulites*, *Discocyclina*, *Asterocyclina*, *Orbitolites*) at levels which can be shown to occur within the biostratigraphic limits of the Priabonian Stage (Upper Eocene);

3. a number of biostratigraphic events in the calcareous nannoplankton (LAD of all rosette-shaped discoasters, *i.al.* *Discoaster barbadiensis*, *D. saipanensis*; LAD of *Reticulofenestra reticulata*) and planktonic foraminifera (LAD *Globorotalia centralis* gp., *G. cerroazulensis* gp., *Hantkenina*, *Globigerapsis*) which can be shown to occur within the biostratigraphic limits of the Priabonian Stage (Upper Eocene);

4. major changes in molluscan faunas in Europe and the Soviet Union between units of late Eocene and early Oligocene age;

5. Major floral changes in Europe and the Mediterranean area. These changes involve a replacement of angiosperms by gymnosperms, in terms of dominance, and an increase in 'Arcto-Tertiary' elements reflecting increased aridity, relief and cooling;

6. major palaeobotanical changes in mid- to high latitudes of the Pacific north-west. These changes include replacement of

broad-leaved evergreen forests by temperate broad-leaved deciduous forests representing a decline in mean annual temperature of 12°–13°C at latitude 60°N and 10°–11°C at latitude 45°N, and a change in the mean annual temperature range of from 3°–5°C in middle Eocene to 21°–25°C in the Oligocene (Wolfe 1978);

7. in the oceans and on continental margins dramatic changes in oceanic conditions. These include global drop in the CCD (Berger 1973; van Andel 1975; Ramsay 1977) coinciding with the basin-shelf fractionation change that occurred between the Eocene and the Oligocene; i.e. extensive carbonate precipitation on broad, warm, shallow shelves (leading to widespread development of nummulitic limestones) in the Eocene was replaced by terrigenous sedimentation on reduced shelf areas and a large scale transfer of carbonate to the deep sea; global lowering of palaeotemperature of about 3°–5°C (Kennett & Shackleton 1976; Keigwin 1980); global eustatic sea-level fall (Vail *et al.* 1977) which has its expression in the essentially global regression seen in passive continental margin stratigraphic sequences around the world.

These changes should be viewed as a sequence of step-like events which occurred over an interval of time spanning several million years in response to major changes in ocean-continent geometry, and attendant palaeoclimatic (predominantly high latitude cooling) and palaeo-oceanographic (development of vigorous deep water circulation) changes. The Eocene–Oligocene boundary itself may be viewed as coinciding approximately with a 'threshold' event whereby the earth appears to have entered into an irreversible climatic phase characterized by a thermospherically derived deep water circulation pattern (Corliss *et al.* 1984).

A precise definition of the Eocene–Oligocene boundary remains controversial, stemming in no small part from continued controversy surrounding biostratigraphic correlation of the various stages used for Upper Eocene–Lower Oligocene strata, lamentably, but historically unavoidably located in the shallow water basins of northern Europe.

The Eocene–Oligocene boundary is traditionally placed at the lithic and faunal discontinuity between strata assigned to the Priabonian (Mediterrean region) or Ludian (Paris Basin) Stage and the Lattorfian (North German Basin), Stampian (Paris Basin) or Rupelian (Belgian Basin) Stage. The Oligocene, as originally defined by Beyrich (1854) was created for a series of rocks in northern Europe believed to represent a major transgression. Its uppermost part included rocks equivalent to the lowest part of Lyell's Miocene series, i.e. based on the 'Apennine Marls'. As its lowest fossiliferous unit the Oligocene included the sands of Magdeburg and Egeln in Germany. The historical modifications to the term Oligocene, particularly as a result of the expansion of the concept of the lower unit, the Lattorfian Stage, has resulted in considerable problems in arriving at agreement on appropriate time-stratigraphic terminology. The molluscan fauna of the Lattorfian Stage *s.l.* has been shown to range from late middle Eocene to early Oligocene in age, whereas the (long since inaccessible) stratotype locality may be of latest Eocene or earliest Oligocene age (see below). In any case it is inappropriate as a standard chronostratigraphic term.

Much of the current controversy around the Eocene–Oligocene boundary centres on the biostratigraphic position of the stratotype Lattorfian. Martini & Ritzkowski (1968) have interpreted it as being equivalent to the *Ericsonia? subdisticha* (NP21) Zone and proposed a redefinition of the Lattorfian and base of the Oligocene at the base of Zone

NP21. This suggestion is hardly practical nor does it represent correct stratigraphic procedure. The latitudinally diachronous extinction of rosette shaped discoasters (*Discoaster saipanensis* and *D. barbadiensis*) during the late Eocene (Cavelier 1972, 1979; Aubry, pers. comm. 1982) results in distinctly time-transgressive biostratigraphic correlations. The redefinition of the base Lattorfian = base Oligocene by Martini & Ritzkowski (1968) leads to a situation in which a biostratigraphic definition (base Zone NP21) for a chronostratigraphic unit will lead to demonstrably time-transgressive correlations elsewhere. Furthermore, proper stratigraphic procedure requires that palaeontological criteria, although definitive for regional correlation (i.e. recognition) beyond the stratotype region, should not be a part of the definition itself (Hedberg (ed.) 1976).

Stratigraphic harmony would best be served, we believe, by abandoning the term Lattorfian as a standard stage unit (see discussion below), and using the terms Priabonian and Rupelian for late Eocene and early Oligocene stages, respectively.

As the discussion below shows it is not entirely clear whether the base of the Rupelian is contiguous with the top of the Eocene (= Priabonian). A possible solution to the problem of early Oligocene chronostratigraphy may be the substitution of a different unit. In the Gulf Coastal Plain of the United States there are neritic marine sediments that span the Eocene–Oligocene boundary in accessible outcrops. These belong to the classic Jacksonian (= late Eocene) and Vicksburgian (early Oligocene) stages. It might be possible to use the Vicksburgian in its present sense, a stage which essentially spans the interval from the top of the Eocene (Priabonian) to the base of the Chattian (= NP23–24 boundary; see below). Alternatively, the term Vicksburgian could be used in a more restricted sense to include that interval between the top of the Eocene and the base of the Boom Clay in Belgium = Middle Rupelian, but the lowest level which can be unequivocally dated biostratigraphically is Zone NP23 (see further discussion below). Studies are currently underway in the Gulf Coast sections and we may expect definitive data on this problem in the near future. An alternative, or supplementary choice, would be the bathyal deposits of the Contessa section(s) in the Apennines (Lowrie *et al.* 1982) in which integrated magnetobiostratigraphic studies have already been done, and in which radiometric studies are being made (Montanari *et al.* 1983). Suffice to say that it would appear that these sections have the requisite characteristics for a more precise delineation of early Oligocene chronostratigraphy and/or of boundary stratotype(s) for the Eocene–Oligocene boundary.

The recent integration of biostratigraphy and magnetostratigraphy in the Mediterranean (Lowrie *et al.* 1982) and the South Atlantic (Poore *et al.* 1982, 1983; LaBrecque *et al.* 1983) have gone a long way towards clarifying the problem of the relative sequence of biostratigraphic events associated with the Eocene–Oligocene boundary. The boundary, as recognized on the basis of the virtually simultaneous, yet discretely separated, LAD's of the *Globorotalia cerroazulensis* and *cocoaensis* groups and *Hantkenina* and *Discoaster saipanensis* and *D. barbadiensis*, falls approximately midway in Chron C13R.

Current age estimates of the Eocene–Oligocene boundary vary from 32 Ma (Armentrout 1981; Wolfe 1981; Glass & Crosbie 1982) to 33–34 Ma (Odin *et al.* 1978; Odin & Curry 1981; Curry & Odin 1982; Odin, (ed.) 1982; Harris 1979; Harris & Zullo 1980) to approximately 34–36 Ma (Odin 1978; Pomerol

1978) to about 37–38 Ma (Hardenbol & Berggren 1978; Rubinstein & Gabunya 1978) based on assessment of various (predominantly glauconite) radiometric dates and palaeontological control of varying reliability and quality. Several lines of evidence now point to an age estimate which is within these limits but which, at the same time, allows rejection of the estimates at both extremes:

1. The younger limits of Chrons C12 and C13 have (high temperature) K-Ar dates of 32.4 Ma and 34.6 Ma, respectively, in the White River Group (containing Chadronian mammalian faunas) at Flagstaff Rim, Wyoming (Prothero *et al.* 1982, 1983).

2. The Bracks Rhyolite occurs in the basal part of a predominantly reversed polarity interval (interpreted as Chron C12R by Testarmata & Gose 1979) in the Vieja Group (Chadronian mammal 'age') of SW Texas. This interval has been reinterpreted (Prothero *et al.* 1982, 1983) as Chron C13R correlative, but might also correspond to Chron C15R correlative. K-Ar data on the Bracks Rhyolite of 37.4 Ma and 37.7 Ma provides limiting dates for the late Eocene (Prothero *et al.* 1982, 1983) if the reversed interval of the Vieja is a Chron C13R or C15R correlative and because the Eocene–Oligocene boundary is biostratigraphically linked with Chron C13R in the deep sea.

3. Upper Eocene strata at Polanyi, Poland belonging to Zone NP19 and the *Rhombodinium perforatum* (dinoflagellate) Zone have sequentially consistent fission track dates of 39.8 ± 1.6 Ma and 41.7 ± 1.7 Ma (Naeser *In:* Van Couvering *et al.* 1981). These dates stand in marked contrast to the 34–35 Ma fission track dates on supposed late Eocene North American strewn field microtektites (Glass *et al.* 1973; Glass & Zwart 1979) in North America and the Caribbean which have led Glass & Zwart (1977) to suggest an age of less than 35 Ma and more recently Glass & Crosbie (1982) an age of 32 ± 1 Ma for the Eocene–Oligocene boundary (see below).

4. Ghosh (1972) has obtained K-Ar (glauconite) dates of 37.6 Ma on the Pachuta Member (Jackson Formation), 37.9 Ma on the Shubuta Member (Jackson Formation), 38.2 Ma on the Moodys Branch Formation, and 39 Ma and 39.4 Ma on the Yazoo Formation — all of which are of late Eocene (Priabonian) age. The Shubuta and Pachuta Members of the Jackson Formation contain a latest Eocene P16–P17 fauna and NP19–20 flora. The dates of Ghosh (1972) are similar to those obtained on Lattorfian strata in NW Germany (see below) and the age estimate of 37 Ma made for the Eocene–Oligocene boundary by Hardenbol & Berggren (1978). In fact it was primarily on the basis of Ghosh's (1972) determinations that Hardenbol & Berggren (1978: 228, Fig. 6) chose the value of 37.0 Ma in estimating the age of this boundary.

5. Sequentially consistent K-Ar dates of 34.9 ± 1.6 Ma and 31.5 ± 1.5 Ma on basalt flows overlain by sediments with an early Oligocene Zone NP23 calcareous nanoflora at DSDP Site 448A in the Philippine Sea (Sutter & Snee 1981; see also Van Couvering *et al.* 1981) are consistent with radiometric calibrations of early Oligocene magnetobiostratigraphy (point 1 above; see also discussion below). The younger limit of Chron C12N has a date of 32.4 Ma; the slightly younger date of 31.5 ± 1.5 Ma on a basalt flow at Site 448A overlain by Zone NP23 suggests correlation with a magnetostratigraphic level close to Chron C11N. Zone NP23 actually extends up to the base of Chron C10N (see below). The data cited here and in point 4 (above) are difficult to reconcile with suggestions of an age of less than 35 Ma for the

Eocene–Oligocene boundary.

6. The Eocene–Oligocene boundary, in terms of marine biostratigraphy (LAD's *Hantkenina alabamensis, Globorotalia cerroazulensis, Discoaster barbadiensis, D. saipanensis*) occurs at a level approximately half way in Chron C13R correlative in the Contessa section(s), Gubbio, Umbria, Italy (Lowrie *et al.* 1982) and at DSDP Site 522 (Poore *et al.* 1982, 1983).

We shall now consider the basis for some of (what we view to be) the anomalously young age estimates of the Eocene–Oligocene boundary. One of the younger estimates for the Eocene–Oligocene boundary, *c.*32 Ma (Wolfe 1981) is based on (high temperature) K-Ar dates of 34.0 ± 1.2 Ma (Fischer 1976) and 30.3 ± 3.0 Ma (Laursen & Hammond 1974) on Goshen-type floral assemblages in the Stevens Ridge and overlying Fifes Peak Formations in the Cascade Mountains, Washington. A number of dates centring on 33 Ma mark the initiation of volcanism in the Sierra Nevada. Thus Wolfe (1981: 43) suggests that the marine Wheatland Formation of the Sacramento Valley, California, with a reported Refugian benthic foraminiferal fauna (Kleinpell 1938) and abundant volcaniclastic, including rhyolitic, debris, must be as young as 33 Ma. If the Refugian, in turn, is entirely of late Eocene age, the Eocene–Oligocene boundary must be younger than 33 Ma. This chain of correlation(s) depends upon a number of tenuous assumptions:

1. the assumed contemporaneity of the volcaniclastic tuffs in the Wheatland Formation and the dated tuffs in the Cascade Mountains. A radiometric age of 53.5 Ma (based on a composite of 35 to 40 andesite pebbles from the basal part of the Wheatland Formation, is cited in a footnote by Wolfe (1981: 43) which, as he indicates, merely indicates that the Wheatland is younger than 53.5 Ma and that one of the sources of the pebbles is older than 53.5 Ma. However, by the same token this does not necessarily prove the contemporaneity of the Cascade Mountains volcanism (*c.*33 Ma) and the pyroclastic debris in the Wheatland Formation;

2. the assumed biostratigraphic accuracy of a Refugian benthic foraminiferal fauna. Paleogene benthic foraminiferal 'stages' of California have been shown to be distinctly time-transgressive (Steineck & Gibson 1971; Poore 1976; Bukry *et al.* 1977), although it would appear that the Refugian is probably of latest Eocene age (Tipton 1976, 1980). Kleinpell *et al.* (1980) considers that it also includes lowermost Oligocene.

3. the assumption of the reliability of the radiometric dates. Averaging dates made on two stratigraphically distinct lithostratigraphic units is a dangerous procedure. As we have seen above, however, the radiometric dates on the Chron C12 to C15 series and marine correlation with marine magnetobiostratigraphy, suggests that the Goshen-flora if reliably dated, is of early Oligocene age.

The radiometric data from the Cascade Mountains and the suggested correlations by Wolfe (1981) were accepted by Armentrout (1981, p. 140, item 15) in his compilation of Pacific North-west biostratigraphic units and their correlation with a global chronostratigraphic and geochronologic scale. High temperature K-Ar dates of 37.5 ± 3.6 Ma and 38.5 ± 1.6 Ma on basalt intercalated with 'Narizian' and 'Refugian' foraminifera, respectively, in the Pacific North-west (Armentrout 1981, p. 140, item 5), led Armentrout (1981, Figs 2, 3, p. 143, 145) to then estimate an age of 32 Ma for the Eocene–Oligocene boundary and ages for the base of the Refugian and the Priabonian of 39 and 40 Ma, respectively, resulting in an anomalously long (7–8 Ma) and

numerically anomalous (32–39 or 40 Ma) late Eocene. Armentrout (1981, p. 138) is then led to conclude that 'the fact that other workers are also proposing younger ages for the Eocene–Oligocene boundary (Odin 1978; Pomerol 1978; Wolfe, this volume; Harris, 1979; Fullagar *et al.*, 1980) suggests that the Oregon–Washington and European time-scales are accurately calibrated.' This conclusion is scarcely justified; the scientific validity of an argument, to say nothing of that elusive chimera we call 'truth', is not guaranteed by majority opinion. It requires careful assessment of empirical data from a variety of sources.

Glass & Crosbie (1982) have recently estimated the age of the Eocene–Oligocene boundary to be about 32.3 ± 0.9 Ma based on upward extrapolation of sedimentation rates in several DSDP cores from a microtektite layer with which are associated the termination or reduction in abundances of several radiolarian taxa.

The microtektite layer was reported from a Caribbean piston core (RC9–58) by Glass *et al.* (1973) and shown to have a fission-track date of 34.6 ± 4.2 Ma. It was related (i.e. correlated with) to the North American tektite strewn field for which K-Ar and fission-track dating methods have yielded apparently concordant dates of about 34–35 Ma.

In presenting an analysis of these data it is important to distinguish between the biostratigraphy of the *Thyrsocyrtis bromia* Zone and the biostratigraphy of the tektite layer and the associated termination or reduction in abundance of various radiolarian taxa.

The *Thyrsocyrtis bromia* Zone has been generally regarded to be of late Eocene age and its boundary with the overlying *Theocyrtis tuberosa* Zone to coincide with the Eocene–Oligocene boundary (Riedel & Sanfilippo 1978; Glass & Crosbie 1982). Indeed, Glass & Crosbie (1982: 472, 473) query why Hardenbol & Berggren (1978) showed the *T. bromia* Zone extending into the lower Oligocene. This extension was based on the correlations presented by Hays *et al.* (1972: 88, 89) in which the *T. bromia–T. tuberosa* zonal boundary (as defined by the FAD of *Lithocyclia angusta*) is shown to lie (in core 49B, DSDP Hole 77B) within the *Coccolithus bisectus–Helicopontosphaera compacta* Subzone and the *Pseudohastigerina barbadoensis* Zone of early Oligocene age (see also Goll 1972; 947 who observed that the *T. bromia* Zone spans the Eocene–Oligocene boundary and that its top 'must lie' within the *Cassigerinella chipolensis-Hastigerina micra* Zone of Bolli based on DSDP Leg 77 studies). Studies on several other DSDP sites (*i.al.* 162, 216, 366, 462) indicate that the *T. bromia* Zone straddles the Eocene–Oligocene boundary and that its top lies in calcareous nannoplankton Zone NP21. Indeed, in a study based on DSDP Site 462 (Nauru Basin, western central Pacific) Sanfilippo *et al.* (1981: 550, 501) show that the *T. bromia/T. tuberosa* zonal boundary lies within Zones NP23 and P20. A compilation of published data from DSDP legs 1–50 led the same authors to suggest a correlation of the *T. bromia–T. tuberosa* boundary with a level near (but below) the NP21–22 boundary (Martini 1971) which is correlative with a level within the *C. formosus* (CP16b) Subzone of the *Helicosphaera reticulata* (CP16) Zone (Bukry 1973; Okada & Bukry 1980). Finally the *T. bromia–T. tuberosa* boundary (based on the FAD of *L. angusta*) at the Bath Cliffs section, Barbados occurs (Sanfilippo, pers. comm. 1982) at a level about 12–13 m above the Eocene–Oligocene boundary as denoted by the LAD of the *Globorotalia cerroazulensis* group, *Hantkenina* spp., *Nuttallides truempyi*, and the

rosette-shaped discoasters (Aubry, pers. comm. 1982). The biostratigraphic and chronostratigraphic position of the *T. bromia–T. tuberosa* zonal boundary would appear to be reliably established near the NP21–22 boundary and in the early Oligocene.

The biostratigraphy of the tektite layer(s) is a separate problem. Glass & Crosbie (1982) have shown that at least 4 taxa (*Thyrsocyrtis bromia, T. triacantha, T. tetracantha,* and *Calocyclas turris*) became extinct or experienced a reduction in abundance at the microtektite layer in several DSDP sites. At several sites this level is within late Eocene planktonic foraminiferal (P15 or P16) or calcareous nannoplankton (*Discoaster barbadiensis, Isthmolithus recurvus* or *Sphenolithus pseudoradians*) zones.

The biostratigraphic ranges of these (and associated taxa) are not unequivocal and this has led to difficulties on the part of those attempting biostratigraphic syntheses. For instance, the ranges of these four taxa are shown to extend into early (Johnson 1977) to middle (Sanfilippo *et al.* 1981) Oligocene levels, at DSDP Sites 366 (South Atlantic) and 462 (western central Pacific), respectively. However, in a recent study of the Bath Cliff section, Barbados, the termination of three of the species mentioned above has been shown to coincide with the (late Eocene) tektite layer and that of the fourth taxon (*T. triacantha*) occurred only slightly prior to this (Sanfilippo, pers. comm. 1982). The stratigraphic sequence at Bath Cliff, Barbados, is believed to be more complete (i.e. continuous) than those observed heretofore in DSDP sites (Sanfilippo, pers. comm. 1982) and the extension of these taxa into lower Oligocene levels in DSDP cores is now considered to be due to reworking.

In view of the amount of reworking that is seen in the calcareous nannoplankton in the Bath Cliff section and the considerable tectonic disturbance to which the island has been subjected compared to the general stability which has characterized most DSDP sites (including Sites 77, 366, 462, *i.al.*), this interpretation is at least debatable.

Let us now look at the question of the biostratigraphic age of the microtektite layer(s) and the age of the Eocene–Oligocene boundary as proposed by Glass & Crosbie (1982). The age estimate of 32.3 ± 0.9 Ma for the Eocene–Oligocene boundary by Glass & Crosbie (1982) is based on the following data and line of reasoning:

1. Donelly & Chao (1973: 1031) found microtektites which they thought were closest in petrographic properties ('but are not necessarily identical') to the bediasites from Texas (i.e. part of the North American strewn field) which had been dated at about 34 Ma, in the core catcher of core 31 from DSDP Site 149 in the Caribbean Sea. This level is within the *Thyrsocyrtis bromia* Zone (which, as we have seen above, spans the Eocene–Oligocene boundary; cf. Glass & Crosbie 1981: 471, who state that the 'microtektites occurred in sediments of late Eocene age (*Thyrsocyrtis bromia* Zone)'). No independent biostratigraphic age determination was possible on this level at Site 149 because of the scarcity of calcareous microfossils.

2. Glass *et al.* (1973) reported the occurrence of micro-tektites in a piston core (RC9–58) from the Caribbean Sea with a fission-track date of 34.6 ± 4.2 Ma. The microtektites in this piston core were said to be genetically related to those of the North American strewn field based on general appearance, petrography, chemistry, age concordance and geographic propinquity.

3. Biostratigraphic data on this core were presented,

subsequently, by Maurrasse & Glass (1976). They show that several species of radiolaria (see above) were sharply reduced in quantity at the level of the microtektites which they placed in the latest Eocene based on the occurrence of calcareous nannoplankton assemblages in RC9−58 (no stratigraphic data were presented) referable to the upper part of the *Discoaster barbadiensis* Zone or *Cyclicargolithus reticulatus* Subzone according to K. Geitzenauer (pers. comm.). The *Cryptoprora ornata* Zone (defined by Maurrasse 1973) was used to denote the biostratigraphic interval between the 'extinction' of *T. bromia* and associated taxa and the initial evolutionary appearance of *Lithocyclia angusta* from *Lithocyclia aristotelis* (which has been shown above to have occurred in lower Oligocene levels). This taxon was not recorded in RC9−58, however. Extensive evidence for reworking is apparent in this core, however, as indeed the authors acknowledge. For instance, such taxa as *Lithochytris archaea, Lamptomium fabaeoforme* s.s. and *Podocyrtis chalara* (LAD's within the *T. mongolfieri* Zone of middle Eocene age), *Podocyrtis goetheana* (LAD within the middle part of the *T. bromia* Zone), are recorded as 'mingled with younger fossils clearly indicating upward reworking, while apparently also affecting the youngest underlying levels' (Maurrasse & Glass 1972: 207). It is not clear to what 'the youngest underlying level' refers (the level immediately below the microtektites?).

Let us look at the radiolarian evidence more specifically. The age of core RC9−58 is interpreted as latest Eocene based on the supposed restriction of *Cryptoprora ornata* to the late Eocene and the belief that the tektite horizon is a true extinction horizon for *T. bromia, T. tetracantha* and associated forms, despite the fact that they continue as rare faunal components above this supposed extinction level. If the ranges of taxa *above* the tektite level are real and not due to reworking, it could be suggested that the age of the core is older, i.e. within the mid-part of the late Eocene. However, the initial appearance of taxa are of greater reliability in some instances in age determination than (supposed) extinctions; yet these criteria are apparently lacking or at least elusive in the late Eocene.

There are some data which may shed light on the subject and allow an alternative conclusion about the age of this core. *Dorcadospyris* aff. *spinosa* and *D. ateuchus* both occur near the top of core RC9−58 (Maurrasse & Glass 1973, Fig. 2), the former having its lowest occurrence at about 30−40 cm; and the latter at about 75 cm. Both occur within the upper metre of the core, in other words, and about 2 m above the microtektite layer. *Dorcadospyris ateuchus* and *Cryptoprora ornata* are thus shown to range concurrently over the upper 75 cm of the core. The following points are pertinent:
(a) The base of *D. spinosa* occurs between cores 30/31 at DSDP Site 149 (Riedel & Sanfilippo 1973: 724, Table 9) which is within the *Sphenolithus predistentus* (NP23) Zone (Hay & Beaudry 1973: 654, Table 11) and within the *Theocyrtis tuberosa* Zone (Riedel & Sanfilippo 1973: 707, 710). The top of *D. spinosa* occurs between cores 29 and 30 at Site 149 (Riedel & Sanfilippo 1973: 724) within the *T. tuberosa* Zone (Riedel & Sanfilippo 1973: 710) and at the *Sphenolithus predistentus−S. distentus* (NP23−NP24) boundary (Hay & Beaudry 1973: 654, Table 11). A similar relationship between the stratigraphic occurrence of *D. spinosa* and other zonations based on calcareous and siliceous plankton has been shown at several other DSDP sites (e.g. in the equatorial Pacific where Moore 1971: 728, Fig. 1) shows *D. spinosa* to range within the upper part of the

T. tuberosa Zone which is within the interval of the *Discoaster tani ornatus-Sphenolithus predistentus* (calcareous nannoplankton) and P19−20 (planktonic foraminiferal) zones which are of early−middle Oligocene age.
(b) The base of *Dorcadospyris ateuchus* (morphotype) occurs between sections 2 and 3 of core 29 at DSDP Site 149 (Riedel & Sanfilippo 1973: 724) which is within the *T. tuberosa* Zone and at the *Sphenolithus distentus−S. ciperoensis* (NP25−NP24) zonal boundary (Hay & Beaudry 1973: 654) and within the *Globorotalia opima opima* (P22) Zone (Bolli & Premoli Silva 1973: 487). The evolutionary first occurrence of *D. ateuchus* occurs, on the other hand, between section 4 of core 28 and section 2 of core 29 which is near the *S. ciperoensis−Triquetrorhabdulus carinatus* (NP25−NN1) zonal boundary and near the *Globorotalia opima opima−G. kugleri* boundary (see references above). A similar relationship has been demonstrated at several DSDP sites (86, 94, 95, 96) in the Caribbean, the South Atlantic (366, 369), and the equatorial Pacific (70−73, 462, among others).

In summary, a perusal of DSDP data on the stratigraphic range of *D. spinosa* and *D. ateuchus* indicates that both taxa make their initial occurrence at or near the lower−upper Oligocene boundary, within the interval of the *S. predistentus−S. distentus* (calcareous nannoplankton) zones, within the interval of planktonic foraminiferal zones P19−20/21 (approximately the interval of the *ampliapertura-opima* zones) and within the interval of the upper part of the *T. tuberosa* — lower *T. annosa* (radiolarian) zones.

Yet Maurrasse & Glass (1976) show these two taxa, which are elsewhere regarded as reliable biostratigraphic markers for a mid-late Oligocene age, occuring together over a short interval interpreted as late Eocene in age. We are faced with a dilemma of the following nature:
1. If the stratigraphic ranges of the radiolarians shown in core RC9−58 (Maurrasse & Glass 1976, Fig. 2) are taken at face value, two taxa, previously regarded as reliable indicators of mid−late Oligocene age, are documented to range down into upper Eocene levels.
2. An alternative explanation is that if the initial appearance of *D.* aff. *spinosa* and *D. ateuchus* indicates a mid-Oligocene age for the upper 1 m of core RC9−58, then an unconformity (? paraconformity) at or just below the tektite level may be present which may account for the abrupt disappearance (reduction in abundance) of several radiolarian taxa. The continued presence of *T. tuberosa* above the microtektite level in RC9−58 may represent its normal stratigraphic range within the lower Oligocene. The fission-track date of 34.6 ± 4.2 Ma may then represent an early Oligocene date within the upper part of the *T. tuberosa* Zone, somewhat below the base of the *S. predistentus* Zone and within Zone P19−20 (by correlation with other DSDP cores).
3. Glass & Crosbie (1982) believe that the occurrence of microtektites in other DSDP cores are stratigraphically equivalent in age to that found in RC9−58 either because they occur in the *T. bromia* Zone (which they assume to be restricted to the late Eocene in the absence of corroborating data from calcareous plankton) or because they occur at levels which, in certain cores, can be shown, on the basis of calcareous plankton, to lie within the late Eocene. They further believe that characteristic chemical and petrographic 'fingerprinting' allow identification and correlation of microtektite specimens. They then proceed a step further and conclude that since the fission-track date of 34.6 ± 4.2 Ma on

the RC9–58 tektite is similar to the concordant set of dates obtained by both K-Ar and fission-track methods (approximately 34–35 Ma) on North American tektites, these separate microtektites are all the unique expression of a single contemporaneous event of late Eocene age.

But there are several problems with this interpretation. The North American strewn field apparently yields concordant dates of 34–35 Ma by both fission-track and K-Ar methods which would seem to indicate their consistency, if not reliability. Yet there is no definitive evidence for their stratigraphic position. The bediasites of Texas occur in secondary position, only a single *in situ* sampling having been reported (King 1968: 160) in 'bedded Jackson Group rocks'. (These are non-marine and their relationship with the marine Jackson of Alabama-Mississippi is unknown.) They were said to occur in 'close association with outcrops of Oligocene sandstone, from which they are presumed to have been derived' (McCall 1973: 281) but no reference for this age determination was cited. The Georgia tektites occur in Pliocene–Pleistocene deposits. In short we have no definitive evidence of the stratigraphic position of the North American tektites.

Finally, the spectre of multiple microtektite strewn fields during a 3–8 Ma timespan (38 Ma, 34 Ma, and 30 Ma) has been raised by the 40-Ar — 39-Ar dating of North American tektites (bediasites) and two impact craters in Canada (Lakes Wanapitei and Mistastin; Bottomley *et al.* 1979). The presence of multiple microtektite strewn fields is suggested by current investigations of microtektite occurrences in several DSDP sites in the Pacific (Sites 167, 292), Atlantic (Site 363), Caribbean (149, RC9–58), Gulf of Mexico (Site 94, E67–128), Indian (Site 242) Oceans and St. Stephen's Quarry, Alabama (Keller 1983; Keller *et al.* 1983). Microtektites from five levels ranging in age from late middle Eocene to mid-Oligocene have been recovered, and at least five of the occurrences (in Sites E67–128, 94, 167, 242, and 292) have been shown to be coeval and of late Eocene (P15–P16 boundary) age. A second microtektite level is shown to lie at a level correlative with P15 or near the P14–P15 boundary (including the occurrence in Site 149 and RC9–158 discussed above). Both of these intervals are associated with hiatuses and carbonate dissolution. Keller *et al.* (1983) observe that (1) the sediments underlying the microtektite horizon in these two cores are of late middle Eocene age; (2) the dissolution interval containing the microtektite horizon is latest middle Eocene or late Eocene age; (3) the late early Oligocene *S. predistentus* Zone (CP12) overlies the dissolved interval in Site 149 suggesting that a hiatus spans the latest Eocene–earliest Oligocene. Further, Keller (written communication 1983) points out that microtektites are scattered throughout a 1 m interval with two abundance peaks in RC9–158 and the sediments between these two peaks have reworked late Eocene to early Oligocene calcareous nannoplankton assemblages so that dating of specific levels within this reworked interval is precluded. Keller *et al.* (1983) indicate that it is possible that these tektites may, in fact, be identical (correlative) with those in the other, well dated, later Eocene (P15–P16) level. The record of microtektite horizons associated with the P13–P14, P17–P18, and P20–P21 boundaries (Keller *et al.* 1983) will require further documentation in the form of chemical and petrographic analyses. Evidence for a late Eocene–early Oligocene hiatus remains ambiguous. For instance, the gap of 12–13 m at DSDP Site 149 between the core containing the micro-

tektite layer and the overlying core belonging to the *Sphenolithus predistentus* (NP22) Zone precludes a definite determination that the basal Oligocene *Helicosphaera reticulata* (NP22) zone is missing. The latter zone is extremely short (0.5 m.y.) and is only slightly subsequent to Chron C13N (see Fig. 6) and it (or at least sediment representative of the time to which this zone corresponds) could be present in the intervening coring gap. In a similar manner the evidence for a hiatus in RC9–158 is equivocal (see discussion above). Until adequate quantitative and petrographic data are presented on the vertical distribution of 'microtektites' in deep sea cores, the possibility remains that at least some of the occurrences may be due to concentration by erosion and redeposition and/or downhole admixture. However, the identification and correlation of multiple tektite layers in DSDP cores and land sections with the North American strewn field(s) may be complicated beyond the point of radiometric resolution. (4) Finally, Glass & Crosbie (1982) have estimated an age of 32.3 ± 0.9 Ma (given as 32.5 ± 0.9 Ma in the text) for the Eocene–Oligocene boundary (as denoted by the calcareous plankton) in nine DSDP cores (three of which were considered reliable, two additional to be useful) by upward extrapolation of sedimentation rates as provided in the DSDP Initial Reports. The age difference between the tektite and the Eocene–Oligocene boundary was believed to range between 1.6 and approximately 2.2 Ma. The correlation of the late Eocene microtektite horizon with the P15–P16 zonal boundary by Keller *et al.* (1983) suggests its association with a level approximately correlative with the top of Chron C16N, which would indicate an age of about 1.5 m.y. older than the Eocene–Oligocene boundary according to the magneto-chronologic scale presented here.

In summary, there appear to be two alternatives to the age estimates for the Eocene–Oligocene boundary based on fission track dates:

1. The radiometric dates on the North American strewn field and those on the microtektites in RC9–58 are reliable and reflect a single impact event of late Eocene age. In this case the current age estimates of the Eocene–Oligocene boundary need to be revised accordingly. This interpretation conflicts with other radiometric and palaeomagnetic data presented in this paper which suggests that the Eocene–Oligocene boundary lies within the span of 36–38 Ma.

2. The radiometric dates on the North American strewn field and RC9–58 may reflect a late Eocene event but we would view the dates as anomalously young. Indeed, Keller *et al.* (1983) have suggested that the difference in age estimate for the late Eocene microtektite horizon based on magneto-chronology and radiochronology may be partially explained by the bias towards younger dates of the fission track method (see also Odin, (ed.) 1982). The only available date on North American strewn field microtektites is 34.6 ± 4.2 Ma (Glass & Crosbie 1982). Fission track dates on tektites range from 34.5–36.4 Ma with error bars of ± 1.5 Ma to ± 8.3 Ma (*op. cit.*). The older range of these dates is well within the chronologic framework of most palaeomagnetic time-scales, including the one presented here as well as the radiochronology presented by Ghosh (1972) based on glauconites.

A resolution of the conflicting age estimates for the Eocene–Oligocene boundary may eventually come from additional high temperature dating of magnetobiostratigraphically controlled horizons associated with the boundary. Preliminary K-Ar (biotite) dates on the top of Chron C13N and C16N correlatives in the Contessa Road section, near

Gubbio, of 35.2 ± 0.5 Ma and 36.1 ± 0.5 Ma (Montanari et al. 1983; 1984) yield an age estimate of 35.6 ± 0.5 Ma for the biostratigraphically determined Eocene–Oliogocene boundary (Lowrie et al. 1982). These dates may be contrasted with the age estimate of 32–34 Ma for this boundary cited above and below.

The age estimate of less than 34 Ma for the Eocene–Oligocene boundary by Harris (1979), Fullagar et al. (1980), Harris & Zullo (1980), is based on a Rb-Sr glauconite isochron date of 34.8 ± 1 Ma on the Castle Hayne Formation of New Hanover County, North Carolina, at a stratigraphic level interpreted as belonging to calcareous nannoplankton zones NP19 and NP20 (= late Eocene, Priabonian Stage; Turco et al. 1979; Worsley & Turco 1979). We have dealt with this set of data elsewhere (Berggren & Aubry 1983) and will not consider it further here beyond pointing out that an analysis of the stratigraphic section from which the radiometric date was made has shown that it is of middle Eocene age (Claibornian Age = late Lutetian to early Bartonian Age), and belongs to planktonic foraminiferal Zone P12–P13 and calcareous nannoplankton Zones NP16–17 (most likely to NP17). In short, the radiometric date of 34.8 ± 1 Ma refers to a late middle Eocene stratigraphic level and is of no value in estimating the age of the Eocene–Oligocene boundary.

A third source of the younger age estimate (approximately 33 Ma) for the Eocene–Oligocene boundary is the series of K-Ar (glauconite) dates presented by Odin (1978), Odin et al. (1978), Odin & Curry (1981) from NW Europe. Indeed, when the age estimates of Odin for various stratigraphic levels within the Paleogene are plotted against the magnetochronology derived in this paper (Fig. 2) a systematic deviation is seen to occur with maximum extension in the Eocene. We are at a loss to explain this discrepancy except to suggest that some glauconites appear to be unreliable chronometers.

In another vein Odin et al. (1978: 487) prefer the revised (30.9 ± 1.7 Ma) rather than the original (37.5 Ma) date on the latest Eocene Neerrepen Sands of Belgium over the 37.5 ± 0.7 Ma date on the essentially contemporaneous, or only slightly stratigraphically younger, Silberberg Beds of NW Germany as an indicator of the age of the Eocene–Oligocene boundary. The reason for this appears to be that accepting the date on the Silberberg Beds would result in the compression of the duration of calcareous nannoplankton zones NP16–21 into an interval of less than 3 m.y. (37–39 Ma), while the Oligocene zones NP21–25 would span about 14 m.y. (23–37 Ma). But this, in turn, is due to the acceptance of a radiometric date of 39 Ma on the base of the Bartonian. They place the middle–upper Eocene boundary at the base of the Bartonian; we have placed this boundary at the top of the Bartonian. Our own (magnetochronologic) estimate for the base of the Bartonian (= P12 = NP16) would be about 44–45 Ma. Odin et al. (1978) suggest that if an age of 31 Ma is accepted for the Neerrepen Sands, then 33 Ma would be a reasonable estimate for the Silberberg Beds, in which case zones NP16–21 would span 7 Ma and NP21–25 would span 10 Ma. This, they state, 'is a much more reasonable proposition' (Odin et al. 1978: 488).

We fail to understand the reasoning behind this statement. There is no inherent reason for biostratigraphic zones to be of equal (or even comparable) duration. Most biostratigraphic (and virtually all calcareous nannoplankton) zones currently in use are based on the first or last appearance of various (often unrelated) taxa. The relative duration of some biostratigraphic zones is more often a reflection of palaeo-oceanographic-palaeoclimatic factors. Thus, major palaeoclimatic changes may induce an acceleration in evolutionary turnover. This would result in an accelerated number of biostratigraphic events leading to greater biostratigraphic resolution over a short interval of time, such as in the Pliocene (Berggren 1973; 1977a, b). However, one cannot assign an average length of time to biostratigraphic zones, or assume a priori a similarity in duration, and use this as a means of manipulating age estimates of biostratigraphic, let alone time stratigraphic, boundaries.

Finally, we note that in defence of this estimate of c.33 Ma for the Eocene–Oligocene boundary, Odin et al. (1978: 490) observe that the various high temperature dates of Evernden et al. (1964) that suggested a 37.5 Ma date for the Duchesnean–Chadronian land-mammal-age boundary would seem to be about 10% high and suggest that the marine and continental chronostratigraphic units are incorrectly 'correlated. The magnetobiostratigraphic studies of Prothero et al. (1982; 1983) have shown, however, that it is possible to correlate the land mammal units directly with oceanic (Poore et al. 1982, 1983) and continental marine (Lowrie et al. 1982) magnetobiostratigraphy. The high temperature dates on the magnetic anomaly 12–13–15 correlative sequence stand in marked contrast to those suggested by Odin and colleagues for late Eocene – early Oligocene horizons in NW Europe. In a similar manner the dates on the Paleocene–Eocene basalts of East Greenland contrast sharply with the various late Paleocene – early Eocene glauconite dates of NW Europe. The resolution of these radiometric date discrepancies appears to us to be a geochemical problem since they tend to be beyond the typical range of analytical errors as well as current uncertainties in stratigraphic correlation.

We have presented sufficient evidence above to show that the Eocene–Oligocene boundary lies within a relatively brief interval which has limiting dates of approximately 36.1–37.4 Ma (= base Chron C13N — top Chron C15N). It terms of biostratigraphic correlations discussed above a numerical estimate of 36.5–37 Ma appears reasonable.

Finally, we note that uncritical acceptance of these younger age estimates for the Eocene–Oligocene boundary have led to what we consider to be a premature misreading of the geohistorical record (Ganapathy 1982; Alvarez et al. 1982). These authors have suggested a cause and effect relationship between a bolide impact (c.34 Ma), the termination of five 'major' (sic !) radiolarian species (Ganapathy 1982: 885) (which were said to constitute over 70% of the total Radiolaria) and an iridium anomaly at supposedly correlative levels in DSDP Site 149 and RC9–58 in the Caribbean. These events are believed to have occurred near the Eocene–Oligocene boundary. We have shown above, however, that these conclusions are unjustified, and that multiple bolide impacts may be involved here. At any rate, there is no evidence in the deep sea of an abrupt change in microfauna or microflora (planktonic or benthic) during the late Eocene or associated with this boundary. The record is rather of a sequential change in various faunal and floral elements (with extinctions generally exceeding new forms) beginning in the late middle Eocene (Corliss et al., 1984).

The interpretation of the geohistoric record at the Eocene–Oligocene boundary in a framework of 'catastrophism' (Ganapathy 1982; Alvarez et al. 1982) similar to that postulated for the Cretaceous–Tertiary boundary (Alvarez et al. 1979; 1980) is quite unwarranted by presently available data.

The Oligocene

A threefold subdivision of the Oligocene Epoch (Beyrich 1854) is generally accepted by many stratigraphers: Lattorfian (Mayer-Eymar 1893), Rupelian (Dumont 1849) and Chattian (Fuchs 1893). The term Stampian (d'Orbigny 1852) is generally used in France for the lower and middle subdivision of the Oligocene (Fig. 6).

For the past decade it has been suggested that the Lattorfian Stage spans the time interval represented by late middle Eocene (NP15–16) to earliest Oligocene (NP21) (Cavelier 1972, 1979; Hardenbol & Berggren 1978) and that the Oligocene is adequately served by a two-fold subdivision into Rupelian (lower) and Chattian (upper) stages. Calcareous nannoplankton extracted from gastropods of the von Koenen collection have yielded a stratigraphically undefinitive nanofloral assemblage assigned to the *Ericsonia* ? *subdisticha* (NP21) Zone (Martini & Ritzkowski 1968; Martini 1969) based primarily on the absence of rosette-shaped

discoasters (*Discoaster barbadiensis*, *D. saipanensis*). Ritzkowski (1981) has emphasized that, although the molluscan faunas from various North German localities grouped together by von Koenen and which have today resulted in an extended concept of Lattorfian *s.l.*, span the time interval of late middle Eocene (NP15) to earliest Oligocene (NP21), the stratotype Lattorfian is of NP21 (= early Oligocene) age. However, we would reject the use of the Lattorfian Stage as a standard chronostratigraphic unit in mid-Cenozoic stratigraphy for the following reasons:

1. The Lattorfian nanoflora is not definitive for age assignment. The absence of a typically late Eocene assemblage of discoasters is not definitive for assignment in as much as it is now well known that these taxa disappear earlier than *Hantkenina* and *Globorotalia cerroazulensis* in low-latitudes (Hardenbol & Bergren 1978; Lowrie *et al.* 1982; Poore *et al.* 1982, 1983; various Deep Sea Drilling Initial Reports) and progressively earlier in mid- to high latitudes
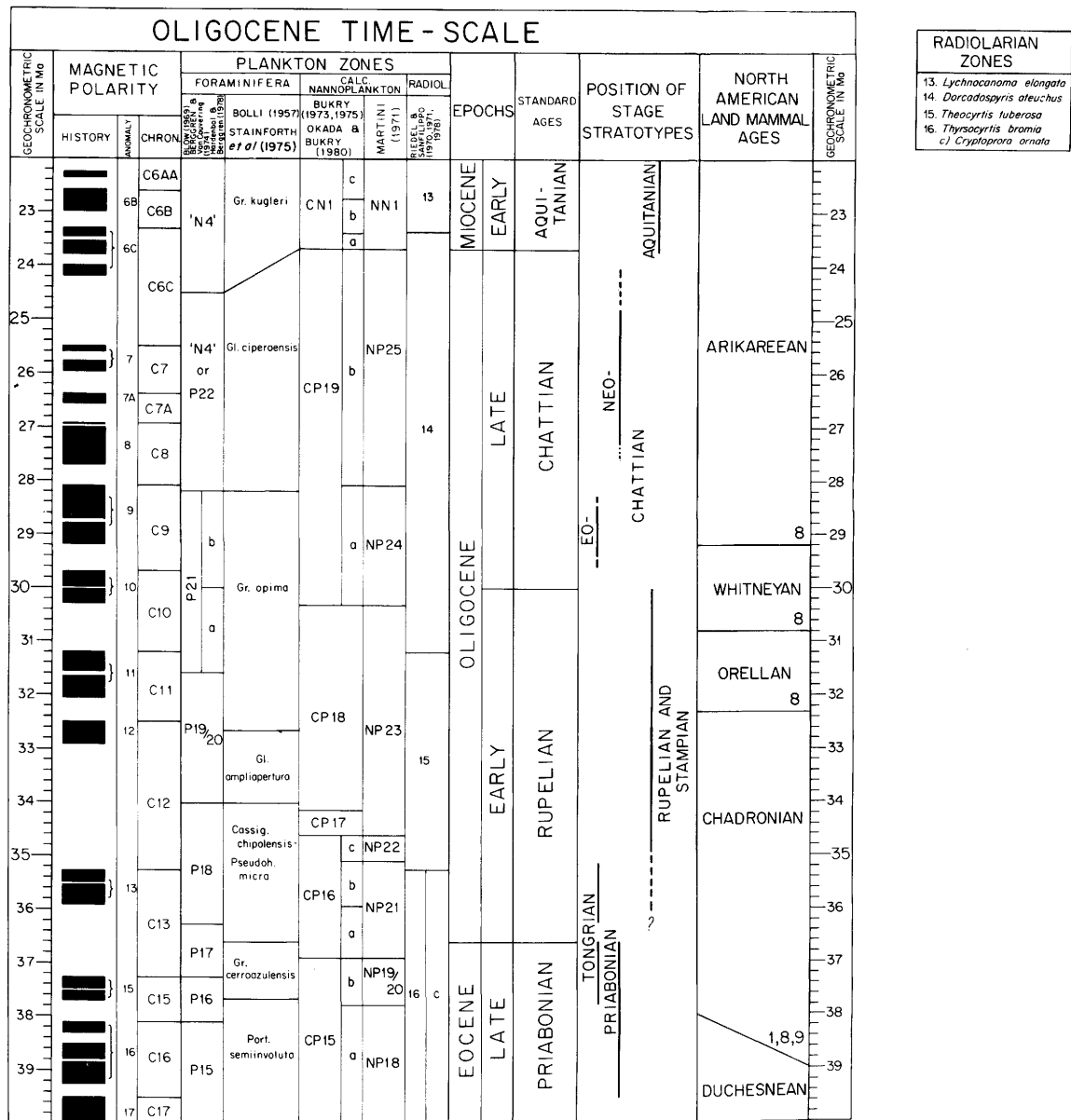
FIG. 6.   Oligocene geochronology (explanation as in Fig. 3).

(Cavelier 1972, 1979; Aubry, pers. comm. 1982).

2. The Lattorfian stratotype, located in an abandoned lignite open cast mine in East Germany, has not been accessible for over 70 years (Ritzkowski 1981), scarcely a commendable attribute for a standard chronostratigraphic unit.

Recently Benedek & Müller (1976) have proposed a modified (extended) biostratigraphic correlation of the Lattorfian as exposed at the former Piepenhagen brickworks at Doberg, near Bunde (Westphalia) where, in addition to Brandhorst and Vahrenkamp, beds, assigned to the Lattorfian, have been recently found. They have extended the Lattorfian (= lower Oligocene) to include the *Helicosphaera reticulata* (NP22) Zone and redefined the Lattorfian–Rupelian boundary to coincide with the *H. reticulata* (NP22) – *Sphenolithus predistentus* (NP23) boundary. On the other hand, sediment scraped from a specimen of *Fusus elongatus* from the von Koenen collection derived from layer five in the stratotype lignite mine Carl near Latdorf and considered middle Oligocene by von Koenen himself, has yielded an NP22 calcareous nanoflora (Martini, pers. comm. 1979, *In*: Ritzkowski 1981). Ritzkowski (1981: 158) observes that these various data essentially result in a Lattorfian Stage whose time span does not correspond to the early Oligocene: the Lattorfian represents but a part of the early Oligocene of Beyrich (1854), whereas the redefinition, based on the Piepenhagen section at Doberg (Benedek & Müller 1976) extends the Lattorfian to include Zone NP22 (and NP21). The middle Oligocene would follow the lower Oligocene without a stratigraphic break, whereas the Lattorfian s.s. is separated from the Rupelian by Zone NP22.

A more extensive Tertiary sequence is exposed in the lignite open cast mine and clay pits near Helmstedt, and at Lehrte (east of Hannover). A potassium-argon (glauconite) date of 37.5 ± 0.7 Ma has been reported (Graman *et al.* 1975) from the basal part of the Silberberg Beds (= NP21; Martini 1969; Martini & Ritzkowski 1968, 1969, 1970; and with a planktonic foraminiferal fauna 'more related to the Eocene than to the Oligocene (in the sense of the Rupelian . . .'); Marks & von Vessem 1971: 64, 65). Four potassium-argon (glauconite) dates with an average value of 38.6 ± 0.7 Ma have been reported from the Gehlberg Beds (Gramann *et al.* 1975) whose biostratigraphic position has not been determined, although they lie stratigraphically between the Annenberg Beds below (= NP15–NP16) and the Silberberg Beds above (= NP21). A potassium-argon (glauconite) date of 36.4 ± 0.7 Ma and two of 39.4 ± 0.9 Ma and 39.6 ± 0.6 Ma have been determined for the upper and lower parts, respectively, of the *Ostrea queteleti* Sands near Lehrte which are correlated with Zone NP21 (Martini 1969; Haq 1972). Odin *et al.* (1978) have criticized the Silberberg date (37.5 Ma) based on their incompatibility with a date on the Sands of Neerrepen (31 Ma = Tongrian) in Belgium of presumed equivalence with the Silberberg Beds, as well as on the basis of some circular reasoning that attempts to prejudge the appropriate ('more reasonable') time span of late Eocene–Oligocene planktonic foraminiferal zones (see discussion below).

Correlation of the Silberberg Beds of Helmstedt with the glauconitic sand of the stratotype Lattorfian can be made on the basis of the extensive molluscan fauna (i.e. independent of the imprecise, yet probably correct, determination based on calcareous nannoplankton) and the date of 37.5 Ma is viewed here as a reasonable determination on a stratigraphic level close to the Eocene–Oligocene boundary (Hardenbol &

Berggren 1978). More definitive data are seen in the form of biostratigraphically well controlled (P16–17, NP19–20), latest Eocene, K-Ar (glauconite) dates of 36.7 Ma and 37.0 Ma on the uppermost Gulf Coast Jackson Formation (Hardenbol & Berggren 1978). These data support the relationship between bio- and magnetostratigraphy and radiochronology of the latest Eocene–early Oligocene in deep sea and continental sections discussed above.

The Rupelian = Stampian Stage represents the first post-Eocene transgression of NW Europe; their upper limits are sharply demarcated by the distinct regression (= eustatic sea-level fall) of the overlying Chattian Stage. Their biostratigraphic limits, particularly their lower boundaries, have proved difficult to determine because of the paucity of definitive faunal and/or floral data important in regional correlation.

The Boom Clay (the main unit of the Rupelian) and the Sables de Fontainebleau (the main unit of the Stampian), both situated in the middle of their stages, belong to the *Sphenolithus predistentus* (NP23) Zone (Martini 1971; Benedek & Müller 1974; Aubry, pers. comm. 1982) and in NW Germany the uppermost part of the Rupelian (= Rupel 4) and the succeeding Eochattian (= Beds 1–25 of the Doberg section) probably belong to the *Sphenolithus distentus* (NP24) Zone although the zonal markers for this zone were not found (Martini 1971; Benedek & Müller 1974, 1976; Martini & Müller 1975). The major part of the Chattian Stage appears to belong to the *Sphenolithus ciperoensis* (NP25) Zone, although again the definitive zonal taxa were not found here (Martini & Müller 1975). In France the basal part of the Stampian Stage (the so-called Sannoisian 'facies') probably belongs to Zone NP22 (Aubry, pers. comm. 1982) and thus corresponds to the lower part of the Boom Clay and subjacent lithostratigraphic units included in the Rupelian Stage in Belgium (see below).

The LAD of *Pseudohastigerina* in the middle part of the Rupelian (= Rupel 3) and of *Chiloguembelina* at the top of the Rupelian (= Rupel 4) has led Ritzkowski (1982) to suggest that the Rupelian–Chattian boundary should be more appropriately placed at the biostratigraphic position of the latter, rather than the former datum (cf. Hardenbol & Berggren 1978; Fig. 4). This is an important point with which we concur and it is all the more important in the light of recent magnetobiostratigraphic correlations in HPC (hydraulic piston cores) taken by the Deep Sea Drilling Project and in the Contessa section(s) at Gubbio, Italy (Lowrie *et al.* 1982).

A synthesis of recent magnetobiostratigraphic data (Poore *et al.* 1982, 1983; Pujol 1983; Lowrie *et al.* 1982; Miller *et al.*, in press; see Table 3 in Appendix IV) indicates the following:

1. The LAD of *Pseudohastigerina* occurs at a level virtually equivalent to the NP22–NP23 boundary somewhat below the mid-point of Chron C12R.

2. The NP21–NP22 boundary occurs only slightly above the top of Chron C13N.

3. Zone NP22 is thus extremely short and confined to the basal part of Chron C12R.

4. The LAD of *Globigerina ampliapertura* (P19/20–P21) boundary) and the FAD of *Globorotalia opima* s.s. are associated with Chron C12N.

5. The LAD of *Globigerina angiporoides* is associated with Chron C11N.

6. The NP23–NP24 boundary occurs just below Chron C10N, virtually coincident with the LAP of *Chiloguembelina*.

7. The LAD of *Globorotalia opima* s.s. is associated with

Chron C9N (smaller, atypical specimens appear to range higher to levels equivalent to Chron C8, and even C7.).

8. The FAD of *Globorotalia kugleri* is associated with Chron C6CN, or a somewhat older level in Chron C6CR, and that of *Reticulofenestra bisecta* with Chron C6CN.

The above data leads to the following observations:

1. Previous correlations of the Rupelian–Chattian boundary with the LAD of *Pseudohastigerina* (Berggren 1971, 1972; Hardenbol & Berggren 1978) have been in error. They were based on the general assumption that the sporadic occurrence of *Pseudohastigerina* in the Rupelian (and its absence in the Chattian) indicated the persistence of the genus to the boundary between the two units. In fact the association of the LAD of *Pseudohastigerina* with the NP22–23 boundary just above anomaly 13 correlative in deep sea deposits and within the middle part of the Rupelian (= Rupel 3) in NW Europe, suggests that the lower Rupelian extends downward to older levels that are biostratigraphically equivalent to Zone NP22 and (in view of the short interval of time represented by this zone) perhaps to Zone NP21 itself, which essentially spans the Eocene–Oligocene boundary (Hardenbol & Berggren 1978). In short a two-fold subdivision of the Oligocene into Chattian (above) and Rupelian (below) appears justified by recent magnetobiostratigraphic correlations. Alternatively a three-fold subdivision of the Oligocene may be justified, in which case a new, lower stage should be inserted whose base corresponds to the Eocene–Oligocene boundary which is biostratigraphically linked to a level between Chron C13N and C15N and whose top would be limited only by a clear biostratigraphic identification of an unequivocally defined lithostratigraphic level (= 'golden spike') in the beds historically assigned to the Rupelian. The suitability of the Gulf Coast Vicksburgian and/or the Contessa section(s) of the Apennines, northern Italy, in this connection has been alluded to in the section above.

2. The LAD of *Globigerina angiporoides* (present throughout most of the Rupelian; Berggren 1969; Blow 1969: 315) at Chron C11N indicates that the Rupelian–Chattian boundary is at least as young as Chron C11N.

3. The LAD of *Chiloguembelina* near the Rupelian–Chattian boundary and the suggested correlation of the uppermost Rupelian and basal Chattian with a level within Zone NP24 suggests that this boundary is closely linked with Chron C10N. In actual fact, deep sea magnetobiostratigraphic correlations support correlation of the LAD of *Chiloguembelina* with a level low in Zone NP24 and we would agree that the Rupelian–Chattian boundary is closely linked with the LAD of *Chiloguembelina* and the NP23–24 boundary.

4. The association of the LAD of *Globigerina ampliapertura* with Chron C12N indicates that the top of Zone P19/20 (Blow 1969, 1979) is well within the Rupelian stage (cf. Hardenbol & Berggren 1978, Fig. 4 where the top of Zone P20 was estimated to lie within Chron C10R).

5. The LAD of *Chiloguembelina* which occurs within the stratigraphic range of *Globorotalia opima* s.s. and forms the basis of a two-fold subdivision of Zone P21 (Jenkins & Orr 1972), near the Rupelian–Chattian boundary, suggests that the *Globigerina angulisuturalis/Globorotalia opima* (P21) Concurrent-range Zone extends into the Rupelian, and that the base of Zone P21 (= FAD *G. angulisuturalis*) is situated in the upper part of Chron C11N (Fig. 6; cf. Hardenbol & Berggren 1978, Fig. 4 where the P20–21 boundary is suggested to be correlative with Chron C10N). However, the LAD of

Globigerina ampliapertura in Chron C12N (see point 4 above) if reinforced by additional studies, indicates a biostratigraphic gap between the top of Zone P19/20 (= lower Chron C12N) and the base of Zone P21 (= top of Chron C11N).

6. Ritzkowski (1981, 1982) places the Lattorfian–Rupelian boundary in Zone NP23, below the LAD of *Pseudohastigerina* (= Rupel 3) and estimates an age of 30 Ma for the base of the Rupelian based on a K/Ar (glauconite) date of 29.8 ± 0.5 Ma on basal Rupel Clay beds near Kassel, and suggests that the early Oligocene (= pre-Rupelian) spans the time between 37–30 Ma. This is unlikely, however, since, as we have seen above, the LAD of *Pseudohastigerina*, which occurs within the lower part of the Rupelian, and is associated with the NP22–NP23 boundary, lies somewhat below the mid-point of Chron C12R. Chron C12N and C13N correlatives in the White River group at Flagstaff Rim, Wyoming, are bracketed by high temperature K/Ar dates of 32.4 and 34.6 Ma, respectively, with a date of 33.5 Ma about midway in the reversed interval between the two anomaly correlatives (Prothero *et al*. 1982, 1983). An age of 30 Ma is closer to Chron C10N (see discussion on magnetostratigraphy below) with which we would correlate the Rupelian–Chattian boundary.

7. A (high temperature) K-Ar date of 28.7 ± 0.7 Ma at the Whitneyan–Arikareean (land mammal 'age') boundary in an interval of normal polarity tentatively correlated with Chron C9N (Prothero *et al*. 1982, 1983) is in good agreement with magnetic chronology age estimates made here (Fig. 6), and previously (LaBrecque *et al*. 1977), and serves as a calibration point for a level within the Chattian Stage (= LAD *G. opima* s.s. = NP24–NP25 boundary = later part Chron C9N).

8. A large number of K-Ar (glauconite) dates from NW Germany with an age range of approximately 25 Ma (Eochattian) to approximately 23 Ma (Vierlandian = Aquitanian) (Kreuzer *et al*. 1980) and of 26.2 ± 0.5 Ma on the early Eochattian *Asterigerina guerichi* beds (Gramann *et al*. 1980) has led to the following suggestions:

(a) Oligocene–Miocene boundary = 23 Ma (Kreuzer *et al*. 1980) to 24 Ma (Ritzkowski 1982).

(b) Eochattian–Neochattian boundary = 23.6 ± 0.2 Ma (Kreuzer *et al*. 1980).

(c) Rupelian–Chattian boundary = 26 Ma (Ritzkowski 1982).

It is clear that magnetochronologic estimates made here and in the time scale of LaBrecque *et al*. (1977) are in close agreement with the estimate on the Oligocene–Miocene boundary, but are in wide disagreement with that made for the base of the Chattian. The Rupelian–Chattian has been shown above to be approximately equivalent to the LAD of *Chiloguembelina* and/or the NP23–NP24 boundary which are closely linked with Chron C10N, with an estimated magnetochronologic age of approximately 29.5–30 Ma. This estimate should be compared with the value of 26.2 Ma on the *A. guerichi* beds of the lower part of the Chattian.

9. The Oligocene–Miocene boundary is biostratigraphically linked with the LAD of *Reticulofenestra bisecta* and is stratigraphically equivalent to the FAD of *Globorotalia kugleri*. These events are linked with lower Chron C6CN and have an estimated magnetochronologic age of 23.7 Ma in close agreement with prevailing radiometric dates of *c*.23 Ma for the Chattian–Vierlandian boundary in NW Germany and similar dates elsewhere.

In North America, magnetostratigraphic studies of ter-

restrial Oligocene sequences presently are available from Wyoming, Nebraska, North and South Dakota (Prothero *et al.* 1982, 1983) and western Texas (Testarmata & Gose 1979, 1980). Studies from both of these areas include information on mammalian biostratigraphy, magnetostratigraphy, and high temperature K-Ar radioisotopic chronology.

The work of Prothero *et al.* (1982, 1983) samples sediments of the White River Group that extend from Chadronian to Arikareean in age and that preserve magnetic polarity intervals correlative with Chrons C13 to C9. Prothero (1982) and Prothero *et al.* (1982, 1983) indicate that the Chadronian begins prior to Chron C15N (although this is based on correlation with the sections from West Texas, see below, as their Chadronian magnetic polarity sequence extends only to somewhere within Chron C13N) and ends about midway within the time of Chron C11R (this is the Chadronian–Orellan boundary). The Orellan ends about midway within the time of Chron C10R (Orellan–Whitneyan boundary), and the Whitneyan ends at the beginning of Chron C9N (Whitneyan–Arikareean boundary). Prothero (1982) and Prothero *et al.* (1982, 1983) use detailed mammalian bio-stratigraphy to correlate the three overlapping portions of their composite Chadronian to Arikareean sequence. Although unambiguous correlation of any *one* of the three portions to the standard magnetic polarity time-scale, based on polarity pattern alone, would be difficult, the lengthy composite sequence can be definitely correlated to the Chron C13 to C9 segment of the time-scale. Further recent work on the White River Group permits identification of Chron C15N at the base of this sequence (Prothero, pers. comm.). This provides more direct support for the beginning of the Chadronian prior to Chron C15N.

Five stratigraphic horizons located directly within the magnetic polarity sequence of Prothero (1982) and Prothero *et al.* (1982, 1983) have been dated using high temperature K-Ar and fission-track techniques. Within the Chadronian Flagstaff Rim section (polarity events correlative with Chrons C13N to C12N) four horizons have produced high temperature K-Ar dates on biotites and sanidines ranging from 32.4 to 36.6 Ma (Evernden *et al.* 1964; Emry 1973; Prothero 1982; Prothero *et al.* 1982, 1983). We use the high temperature, K-Ar dates from magnetostratigraphic horizons approximately correlative with the tops of Chrons C12N and C13N, within this section, as two of the calibration points for our magnetochronology (see earlier discussions). Obradovich *et al.* (1973) reported two high temperature, K-Ar dates on biotites of 27.7 ± 0.7 Ma and 28.7 ± 0.7 Ma and a fission-track date on zircons of 28.5 ± 3.1 Ma from the Carter Canyon Ash Bed in the Gering Formation, SW Nebraska (see also Emry *et al.*, in press). The Carter Canyon Ash Bed stratigraphically overlies the normal polarity interval (correlated with Chron C9N) at the top of the Chadronian to Arikareean (polarity sequence correlated with Chron C12 to C9) Pine Ridge section of Prothero *et al.* (1982, 1983). As this ash lies within a stratigraphic interval that has not yet been sampled palaeomagnetically, it provides a date for an interval of time that is *within*, or *younger than*, Chron C9N.

Although the Arikareean has traditionally been considered early Miocene in age (see Emry *et al.*, in press; correlation chart of Wood *et al.* 1941) it is clear from the isotopic and palaeomagnetic data from strata of early Arikareean age that much of the Arikareean instead is late Oligocene in age (Emry *et al.*, in press; Prothero *et al.* 1982, 1983; R. H. Tedford, pers. comm.). In particular, the Oligocene–Miocene

boundary in our geochronology falls within Chron C6CN, with an age estimate of 23.7 Ma, while the base of the Arikareean lies near the base of Chron C9N and is older than 28.0–28.5 Ma.

A precise determination of the location of the Duchesnean–Chadronian boundary (= base of the Chadronian) presently is not available. Prothero *et al.* (1982, 1983) place the boundary somewhere older than Chron C15N, based on (1) biostratigraphic correlation between the Chadronian Flagstaff Rim, Wyoming and Vieja Group, Texas sections, (2) recognition that the base of the Vieja Group section is older than the Flagstaff Rim section as indicated by the presence of older, probably Duchesnean (or latest Uintan) faunas in the Vieja Group section, and (3) reinterpretation of the magnetic polarity sequence of Testarmata & Gose (1979) from the Vieja Group. As the Eocene–Oligocene boundary lies within Chron C13R, Prothero *et al.* (1982, 1983) conclude that at least the basal part of the Chadronian is late Eocene in age. However, interpretation of the Vieja Group magnetic polarity sequence is equivocal (Prothero *et al.* 1982, p. 651; see discussion below), and it is unclear precisely where the Duchesnean–Chadronian boundary lies.

Strata of the Vieja Group contain excellent mammalian faunas of Uintan or Duchesnean to Chadronian age (Wilson *et al.* 1968; Wilson 1978, 1980; Emry *et al.*, in press). Associated with these faunas are numerous high temperature, radioisotopic dates from four bracketing horizons (McDowell 1979; Testarmata & Gose 1979, 1980). The Gill Breccia at the base of the sequence is dated at 41.0 ± 2.0 Ma. This is overlain by strata containing the early Duchesnean (= Eocene portion of the Duchesnean of Wilson *et al.* 1968; included within the Uintan by Wilson 1978) Candelaria local fauna, which is then overlain by the Buckshot Ignimbrite with four dates of 39.6 ± 1.2, 36.1 ± 2.3, 37.1, and 37.3 Ma. Overlying the Buckshot Ignimbrite are the late Duchesnean (= Oligocene portion of the Duchesnean of Wilson *et al.* 1968; included within the Chadronian by Wilson 1978) Porvenir and Little Egypt local faunas, which are then overlain by the Bracks Rhyolite dated at 37.4 ± 1.2 and 37.7 Ma. The Bracks Rhyolite is overlain by strata containing the Chadronian Airstrip and Ash Spring local faunas, and the top of the sequence is capped by the Mitchell Mesa Ignimbrite which has been dated at 32.3 ± 0.7 Ma (average of eighteen individual dates).

Testarmata & Gose (1979, 1980) palaeomagnetically sampled the Vieja Group sequence (approximately 400 metres of section) from just above the Buckshot Ignimbrite to the Mitchell Mesa Ignimbrite, which spans the late Duchesnean to Chadronian portion of this sequence. Their results show a very complex pattern of numerous, generally short polarity events and thick stratigraphic intervals at the base and top of the sequence that are of 'undetermined polarity'. The explanation for the discovery of so many (at least 29) polarity events in such a short interval of time, and short stratigraphic section, is unclear. Testarmata & Gose (1979, 1980) recognize two intervals of predominantly normal polarity strata that they tentatively correlate with Chrons C12N and C13N. Prothero *et al.* (1982, 1983) reinterpret these 'normal polarity' intervals as correlatives of Chrons C13N and C15N based on radioisotopic dates and their correlation of the Chadronian Airstrip and Ash Spring local faunas (which lie within the upper 'normal polarity' interval) with faunas from the Flagstaff Rim, Wyoming section that lie within strata of normal polarity correlated with Chron C13N. We believe that

the confusing magnetic polarity data of Testarmata & Gose do not preclude correlation of these 'normal polarity' events with Chrons C13N and C15N or Chrons C15N and C16N. In any case, it is difficult to correlate unambiguously the magnetic polarity sequence of Testarmata & Gose (1979, 1980) wtih the standard geomagnetic polarity time-scale.

The isotopic dates bracketing the Vieja Group faunas provide a relatively precise age estimate of approximately 37.5 Ma for the Duchesnean−Chadronian boundary. In our geochronology (see Fig. 6) this boundary would fall within, or just below, Chron C15N (as was inferred by Prothero *et al.* 1982, 1983 based on other lines of reasoning). Two other dates consistent with this age estimate of the Duchesnean−Chadronian boundary have been published by McDowell *et al.* (1973). They reported a K-Ar date of 37.2 ± 0.7 Ma on biotite from the top of the early Chadronian Ahearn Member (the lowest of the three members, in the type section) of the Chadron Formation, and a K-Ar date of 40.3 ± 0.8 Ma on biotite at the contact between the Duchesnean Halfway (= Dry Gulch Creek) and Lapoint members in the type section of the Duchesne River Formation. Both of these determinations provide important dates bracketing the Duchesnean−Chadronian boundary from sections that have produced the principal reference faunas for the Duchesnean and Chadronian land mammal ages. However, because strata containing Duchesnean to Chadronian faunas have not yet produced a reliable magneto-stratigraphic correlation of the boundary to the magnetic polarity time-scale, we indicate the uncertain position of this mammal age boundary in Fig. 6 by a diagonal line. We shall not discuss here the present controversy among mammalian biostratigraphers as to the composition, extent, or validity of the Duchesnean.

Recent magnetobiostratigraphic studies (Prothero & Rensberger, in press) on the John Day Formation, east central Oregon, suggest that the Oligocene−Miocene boundary (within Chron C6CN; see companion paper by Berggren *et al.*, this volume) occurs near the top of the *Entoptychus-Gregorymys* Concurrent-range Zone (= latest Arikareean) in North American terrestrial sequences.

A summary of our placement of the boundaries of the Oligocene North American Land Mammal Ages relative to the magnetic polarity time-scale is shown in Fig. 6. Our correlations are based on the data and arguments summarized above, and it is important to note that the Chadronian extends from the late Eocene to the early Oligocene, and the Arikareean extends from the late Oligocene into at least the early Miocene.

The Oligocene−Miocene boundary is discussed at greater length in the companion paper dealing with the Neogene time-scale in this volume.

Conclusions

The basis for a geomagnetic reversal chronology for the late Cretaceous and Cenozoic is the polarity sequence obtained from analysis of marine magnetic anomalies, such as suggested by LKC77. Three linear segments of the LKC77 reversal sequence are inferred on the basis of preferred high temperature age calibration tie-points and the assumption of minimum accelerations in sea-floor spreading history. An initial segment is defined by the origin (0 Ma), anomaly 2A (3.40 Ma), and the top of anomaly 5 (8.87 Ma), yielding an

estimated age of T = 10.42 Ma for the base of anomaly 5. Available radiometric age estimates for magnetozones in land sections correlated to the younger portions of anomalies 12, 13 and 21 (32.4, 34.6 and 49.5 Ma, respectively) are used to extend the chronology by a linear best fit anchored to the base of anomaly 5, yielding an estimated age of 56.14 Ma for the base of anomaly 24. Interpolation between this estimated age for anomaly 24 and a radiometric age estimate of 84 Ma for anomaly 34 correlative (near the level of the Campanian−Santonian boundary) completes the reversal chronology to the younger end of the Cretaceous Long Normal Interval. Relative precision of the reversal sequence depends on the spatial resolution of the magnetic anomaly data and the assumption that sea-floor spreading was at a constant rate over tens of million years somewhere in the world ocean. The accuracy of the reversal chronology ultimately depends on the quality and quantity of radiometric age data used for calibration.

Our assessment of published radiometric dates suggests the following age estimates for the major chronostratigraphic boundaries: Oligocene−Miocene: 23.5 Ma; Eocene−Oligocene: 37 Ma; Paleocene−Eocene: 56.5 Ma; Cretaceous−Tertiary: 66 Ma. The palaeontologically correlated magnetochronologic age estimates for these epoch boundaries are as follows: Oligocene−Miocene (mid-Chron C6CN): 23.7 Ma; Eocene−Oligocene (midway in Chron C13R): 36.6 Ma; Paleocene−Eocene (early part of Chron C24R): 57.8 Ma; Cretaceous−Tertiary (later part of Chron C29R): 66.4 Ma.

Our revised Paleogene magnetobiochronology is consistent with much of the palaeontologically controlled radiometric data base. A notable exception is the Eocene where our age estimates on bio- and chronostratigraphic boundaries differ by about 3−4 Ma at the lower and upper limits and by as much as 6−7 Ma at the lower−middle Eocene boundary from (predominantly glauconite) estimates made by some workers.

The fact that Paleogene stage stratotypes are unconformity bounded and related to eustatic sea-level changes makes precise biostratigraphic recognition of the boundaries difficult. A comparison of the (bio) stratigraphic record across some of these unconformity bounded boundaries suggests that, as a first estimate, the eustatic sea-level cycle (regression-transgression) was on the order of 1−3 m.y. If the concept that 'base defines stage' is rigorously maintained it may prove more efficacious to redefine the base of the Cenozoic stages *within* the normal marine cycles allowing easier biostratigraphic recognition and correlation. This would have the effect of making the boundaries younger than currently determined by most stratigraphers, including the boundary positions shown here (Figs 2−5). Alternatively new stratotype sections should be sought in continuous deep water (bathyal) marine sequences.

Features of interest in this revised Paleogene time-scale include the following:
1. The Cretaceous−Tertiary boundary is biostratigraphically linked in marine sequences with a level just below Chron C29N. In terrestrial sequences this boundary has been linked with a level within Chron C28. However, the interpretation of the data is somewhat ambiguous and we await further studies to clarify whether the two boundaries are, in fact, of different ages, or as we suspect, actually coeval.
2. The type Danian is biostratigraphically linked with Chron 28 and the younger half of Chron 29 at least and may extend into the older part of Chron 27 interval. There is a substantial

stratigraphic gap between the top of the Danian s.s. (within Chron C27R) or top of the Danian s.l. (= Montian s.s.) (= Chron C26–Chron C27 boundary) and the base of the Thanetian (= Chron C26N), an interval of approximately 3 and 2 m.y., respectively. Thus the Thanetian would appear to be inappropriate as a time-stratigraphic unit for the entire post-Danian, pre-Ypresian Paleocene. Recent biostratigraphic studies suggest that the Selandian is a more appropriate unit for this stratigraphic interval. Alternatively, the Selandian stage could be subdivided into a lower (as yet unnamed) substage and an upper (Thanetian) substage.

3. The Thanetian Stage is palaeomagnetically linked with at least a part of Chron C26N and the reversed polarity interval above; the main part is biostratigraphically linked with Zone NP8 and its uppermost part is probably correlative with Zone NP8 as well. This agrees well with deep sea correlations which place the Zone NP7–NP8 boundary just above Chron C26N and the Zone NP8–NP9 boundary in Chron C25N.

4. The 'Sparnacian' facies is within the *Apectodinium hyperacanthum* (dinoflagellate) Zone and of terminal Paleocene age, equivalent, at least in part, to Zone NP9 and Chron C25N. The term Sparnacian is inappropriate as a standard time-stratigraphic unit.

5. The Paleocene–Eocene boundary is biostratigraphically associated with the Zone NP9–NP10 boundary and the *Apectodinium hyperacanthum–W. astra* (dinoflagellate) zonal boundary and lies within the early part of Chron C24R. Reliable high temperature dates are apparently not available associated with this stratigraphic interval. However, one set of (revised) age estimates on the Kap Brewster basalts of East Greenland (56.5 Ma) which appear to straddle the Paleocene–Eocene boundary in terms of dinoflagellate biostratigraphy is reasonably consistent with our magnetochronologic estimate for the boundary of 57.8 Ma.

6. The lower (early) Eocene has undergone substantial revisions in this study. Biostratigraphic studies show that the Ypresian–Lutetian boundary is biostratigraphically linked with a level at or slightly above the NP13–14 boundary which is associated with the base of Chron C22N, whereas the FAD of *Hantkenina*, nominate taxon of Zone P10, and which has commonly been used by planktonic foraminiferal biostratigraphers to denote the base of the Lutetian, is associated with the uppermost part of Chron C22N. The temporal difference between these two biostratigraphic levels is on the order of 1 m.y. The eustatic sea-level fall (and corresponding unconformity which is seen between the Ypresian and Lutetian stages and at correlative levels in various sections) occurs within Zone NP13 and P9 and the regressive-transgressive cycle associated with this event is probably, to a first approximation, on the order of 1 m.y. or less. Revised age estimates for the early Eocene are: 52.0–57.8 Ma (compare with previous estimates of 49–53.5 Ma; Hardenbol & Berggren 1978). The age estimates on the early–middle Eocene boundary are consistent with the recent assignment of radiometrically dated levels (c.49 Ma) near the Bridgerian–Uintan 'land mammal age' boundary to the time corresponding to Chron C20R.

7. The precise correlation of the middle–late Eocene boundary with the geomagnetic polarity stratigraphic scale remains somewhat equivocal. Common biostratigraphic criteria include the FAD of *Porticulasphaera semiinvoluta* (top Chron C18N), LAD of the *Morozovella-Acarinina* group (mid-Chron C17N), FAD of *Chiasmolithus oamaruensis* and/ or LAD of *Chiasmolithus grandis* (= later part of Chron

C18N or later part of Chron C17N). We have chosen to place the early–middle Eocene boundary in the later part of Chron C17N with an estimated age of 40.0 Ma.

8. The Eocene–Oligocene boundary is biostratigraphically linked (LAD of *Globorotalia cerroazulensis-cocoaensis* group, LAD of *Hantkenina*, slightly above the LAD of rosette-shaped discoasters, *D. saipanensis*, *D. barbadiensis*) with a level approximately midway between Chrons C13N and C15N, with an estimated age of 36.6 Ma. This age estimate is consistent with several (predominantly glauconitic) dates of c.37 Ma biostratigraphically associated with the boundary in the Gulf Coast and NW Europe and with recent radiometric calibrations of early Oligocene magnetic polarity intervals (see below).

9. The recent integration of high temperature K-Ar dates and magnetic polarity stratigraphy on latest Eocene-early Oligocene 'land mammal ages' in North America has placed new constraints on age estimates of the Eocene–Oligocene boundary. The younger limits of Chrons C12N and C13N have K-Ar dates 32.4 Ma and 34.6 Ma, respectively. The basal part of a reversed interval that may lie between Chrons C13N and C15N, or alternatively, C15N and C16N, has been dated at 37.4 Ma and 37.7 Ma. These dates suggest that the age of the Eocene–Oligocene boundary lies somewhere in the interval of 36–37 Ma.

10. The Oligocene is best served by a two-fold time-stratigraphic subdivision: Rupelian (Lower), Chattian (Upper). The boundary between these two stages is biostratigraphically linked with the LAD of *Chiloguembelina* and the NP23–24 boundary, which are associated with Chron C10N and has an estimated age of 30 Ma. Previous correlations which linked the Rupelian–Chattian boundary (Chron C10N) with the LAD of *Pseudohastigerina* (midway within Chron C12R, c.34 Ma) are seen to be incorrect.

11. Numerous biostratrigraphic criteria have been suggested to determine the position of the Oligocene–Miocene boundary. We have chosen the FAD of *Globorotalia kugleri* and the LAD of *Reticulofenestra bisecta* (associated with mid-Chron C6CN) as definitive criteria. The resulting magnetochronologic age estimate (23.7 Ma) is in close agreement with recent assessments of published radiometric dates which suggest an age of 23–24 Ma for the Oligocene–Miocene boundary. The genus *Globigerinoides* appears sporadically as early as Chron C7N (c.26 Ma) but attains numerical prominence in deep sea faunas only in the latest Oligocene (in the reversed interval earlier than Chron C6CN = Chron 23). It thus retains its usefulness as a guide to the approximate position of the Oligocene–Miocene boundary.

12. Boundary magnetobiochronologic age estimates and duration of informal divisions (in parenthesis) of the Paleogene are as follows: early Paleocene, 66.4 Ma–62.3 Ma (4.1 m.y.); late Paleocene, 62.3 Ma–57.8 Ma (4.5 m.y.); early Eocene, 57.8 Ma–52.0 Ma (5.8 m.y.); middle Eocene, 52.0 Ma–40.0 Ma (12.0 m.y.); late Eocene, 40.0 Ma–36.6 Ma (3.4 m.y.); early Oligocene, 36.6 Ma–30.0 Ma (6.6 m.y.); late Oligocene, 30.0 Ma–23.7 Ma (6.3 m.y.).

has had the uneviable task of seeing this paper through various metamorphoses to its final, published form.

The preparation of a paper of this nature, attempting, as it does, to synthesize a large amount of data from a wide variety of sources – much of it as yet unpublished – requires the cooperation, to say nothing of the patience and indulgence, of many colleagues. We would like to express our sincerest gratitude to all those who have unselfishly aided us with published and unpublished data in the preparation of this paper. If we have appeared persistent or inquisitive at times it is only because we have tried to make this paper as comprehensive and well documented as possible so that it may serve as a standard in future studies of rates of geologic processes.

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Appendix I Age of Kap Brewster basalt flow (John Obradovich)

In dealing with the K-Ar results reported by Beckinsale *et al.* (1970) for the samples from the chilled margins of the Kap Brewster basalt flow, Fitch *et al.* (1978) assert that a 'best fit' regression line age of 54.5 ± 1.0 Ma results from the previously unrecognized presence of initial argon (i.e. the regression line is of type r_2, Fig. 1 in Fitch *et al.* (1978), with an intercept of 310 ± 12 on the $^{40}Ar/^{36}Ar$ axis).' This statement needs to be examined in some detail. First Beckinsale *et al.* provide four conventionally determined ages. However, only three samples are involved and one sample (7147) had a duplicate argon analysis without an accompanying potassium analysis. The age was determined assuming that the K content was the same as for the first determination. The fact that the two argon analyses for sample 7147 differed by 8.6% should have raised some concern at that time and at the time when Fitch *et al.* subjected this data to a regression analysis. The work of Dalrymple & Hirooka (1965) demonstrated that basalts, even on samples as small as a hand specimen, can be extremely heterogeneous. Potassium and radiogenic argon contents varied by 3.11% and 6.76% respectively in their example. Dalrymple & Lanphere (1969) stressed the importance of using immediately adjacent pieces for argon and potassium measurements. The regression results reported by Fitch *et al.* are not questioned. What is questioned, however, is whether or not Fitch *et al.* are entitled to such a treatment of the data. One could equally well assume that the $^{40}Ar_{rad}/^{40}K$ ratio is invariant in this sample of basalt. That is, if the $^{40}Ar_{rad}$ content shows an increase the K content would also increase correspondingly. Such a sample would have the same $^{40}Ar_{total}/^{36}Ar$ ratio of 591.6 but a $^{40}K/^{36}Ar$ ratio of 90580 instead of 83370. This data point, would simply shift to the right on the $^{40}Ar_{total}/^{36}Ar$ vs $^{40}K/^{36}Ar$ isochron plot. Regressing this data would result in an intercept as low as 295 depending on the assigned uncertainties indicating that there is no initial argon in this instance that deviates from a $^{40}Ar/^{36}Ar$ ratio of 296. The purpose of this treatment is to show that this one data point without an accompanying K analysis has such a significant bearing on the intercept that it should not be considered a valid analysis unless potassium is determined for this specific fragment of basalt.

Fitch *et al.* also cite 2 σ (sigma) values of ± 12 for their uncertainty of the intercept. Given the limited number of samples a more realistic treatment would be based on Student's t approximation for n-2 degrees of freedom in the case of a regression analysis. For four samples t is equal to 4.303 and the uncertainty at the 95% confidence level would be 25.8. With this uncertainty the figure of 310 certainly encompasses the value of air argon (296) and there would be no reason to assume any other value in calculating an age.

When Fitch *et al.* made the statement, 'The recomputation of their quoted average conventional K-Ar age (55.4 ± 3.1, 56.1 ± 1.6, 57.8 ± 2.2, 60.1 ± 2.8 Ma) to give a "best fit" regression line age of 54.5 ± 1.0 Ma' they were comparing ages computed using two different sets of decay constants. As the decay constants used by Beckinsale *et al.* are equivalent to those now universally adopted (Steiger & Jäger 1977) the age Fitch *et al.* should have indicated when comparing their results to those of Beckinsale *et al.* is 55.8 ± 1.0 Ma. Nonetheless we consider this result as incorrect for the reasons cited.

Ultimately we must ask what is the most rational treatment of the data of Beckinsale *et al.* Given the variability in age due to the analysis on 7147 the most preferred age for the Kap Brewster flow would be based on the three conventional ages weighted according to the inverse of their variance. This results in a mean age with a weighted standard error of the mean of 56.5 ± 0.6 Ma.

Appendix II

In this paper we have developed a Cenozoic geochronologic scale in which numerous first order correlations between calcareous plankton datum events and magnetic polarity stratigraphy serve as a magnetobio-stratigraphic framework. The derived (magneto) chronology is anchored to several high temperature K/Ar dates which are, in turn, associated with identifiable parts (magnetic anomalies) of the standard magnetic polarity stratigraphy. We have noted above (Fig. 2) the large discrepancy, particularly during the Eocene, between our derived magnetochronology and the radiochronology based on low temperature K/Ar (glauconite) dates of Odin *et al.* (see References at end of this paper).

An exhaustive discussion of the possible reasons for this discrepancy is beyond the scope of this paper. We shall content ourselves here, however, with a discussion of the problems associated with early–middle Eocene geochronology and, more specifically, with the approximately 7 m.y. difference (greater than 45 Ma vs. 52 Ma here) between Odin and ourselves in the age estimate of the early–middle Eocene boundary. We believe the problems are in part due to the lack of precise biostratigraphic positioning of dated samples, but more seriously a basic problem in the dating of glauconitic material itself.

Paleogene K/Ar numerical dates, based predominantly on glauconite samples, have been compiled by Odin (ed.) (1982) from NW European basins (see also Odin *et al.* 1978: 487, 488) and outside of NW Europe (see also Odin, 1982: 624) as a framework for calculating a Paleogene radiometric chronology (see also Odin & Curry 1981; Odin 1982). A number of dates are listed from the Lutetian (which are essentially younger than 45 Ma) and Cuisian (which are older than 46 Ma) leading to the conclusion that the boundary between these two ages can be placed 'fairly precisely at slightly more than 45 Ma; we propose a figure of 45 ± $^1_{0.5}$ Ma' (Curry & Odin 1982: 625; see also Odin 1982: 6). However, the biostratigraphic position of some of these dated levels as well as the magnetobiostratigraphic correlations presented in this paper reveal that there are fundamental problems with these conclusions.

In Table 2 we list the various K/Ar (glauconite) dates cited by Odin *et al.* (1978) and in Odin (ed) (1982) from lower and middle Eocene levels of NW European basins, their biostratigraphic placement (where possible) and present comments on more recent magnetobiostratigraphic correlations based on studies by Townsend (1982) and Aubry (1983).

The following observations may serve to elucidate the problems involved:
1. K/Ar (glauconite) dates on stratigraphic levels in the Bracklesham Beds of SE England which have been identified with Chron C21N (with an estimated duration of about 1.5 m.y.) include: 43.6 ± 1.8 Ma; 43.8 ± 1.0 Ma; 44.2 ± 1.3 Ma; 44.4 ± 2.3 Ma; 46.1 ± 2.1 Ma; 46.4 ± 1.5 Ma (Odin *et al.* 1978, Tables 2, 3; Table 2, this paper). The minimum and maximum values of these dates range from 41.8 Ma to 48.8

TABLE 2   K/Ar (glauconite) dates from early-middle Eocene levels in NW Europe (from Odin *et al.* 1978: Odin (ed.) 1982, Vol. 2, Tables 2 and 3.)

SAMPLE NO.	STRATIGRAPHIC UNIT	(Ma)	BIOSTRATIGRAPHIC DATE Odin (1982);	Aubry (1983)	AGE REMARKS
1. G 96	Fisher Bed IV	46.1 ± 2.1	basal NP13	NP12/13	Top anomaly 23 correlative at Whitecliff and Bracklesham Bays and in DSDP cores.
2. G435	Fisher Bed VI	44.4 ± 2.3	upper NP13	NP14 (mid)	Base anomaly 21 correlative at Whitecliff Bay (Townsend 1982).
3. G145	Same	43.8 ± 1.0	Same	Same	"
4. G437	Fisher Bed IX	43.6 ± 1.8	NP14 (mid)	NP15	Top anomaly 21 correlative at Whitecliff Bay (Townsend 1982).
5. G396	Fisher Bed XIV	40.7 ± 1.4	lower NP15	NP15 (?upper; by correlation)	Below (older than) anomaly 20 correlative in Hunting bridge Formation at Lee-on-Solent by correlation (Townsend 1982).
6. G144	Fisher Bed 2	46.4 ± 1.5	lower NP13	NP14 (mid-upper by correlation)	Lower part of anomaly 21 correlative at Bracklesham Bay.
7. G234	Fisher Bed 6	44.2 ± 1.3	upper NP13	upper NP14	Fisher VII (Whitecliff Bay) and Fisher 6 (Bracklesham Bay) correlated to each other and placed in upper NP13 by Odin *et al.* (1978). But Fisher 6 (B.B.) = Fisher VII (W.B.) and both are in upper NP14 and in anomaly 21 correlative at both localities and in DSDP cores.
8. G150	Fisher Bed 19	40.2 ± 2.3	lower NP15	top NP15	Correlative with Fisher Bed 19 at Bracklesham Bay and Fisher Bed XIV at Whitecliff Bay; below anomaly 20 correlative by correlation and Zone NP15 by correlation (Townsend 1982; Aubry 1983).
9. G480	Cuisian	47.3 ± 1.4	NP12/13	NP12 (lower to mid)	
10. G176A	Niveau d'Aizy, lower part Sables de Cuise	47.8 ± 3.1	NP12/13	NP12	
11. G 49	Calcaire grossier	44.4 ± 2.3	mid-NP14	upper NP14	
12. G513	"	42.9 ± 1.2	same	same	
13. G583A	basal Lutetian, Zone I	46.2 ± 1.6	—	Upper NP14	Just above erosional contact with Cuisian
14. G527A	basal Lutetian, (glauconie grossier) Zone I	43.7 ± 2.1	—	Upper NP14	—
15. 440	Argiles de Varengeville	53.0 ± 2.4	NP11	NP11	Correlative with anomaly 24 time in DSDP cores and SE England (upper part London Clay and Bagshot Sands).
16. G945	Sables d'Aeltre	45.0 ± 1.5	NP13	NP14	Calculated mean age of 46.3 ± 1.0 Ma of G945 and 941 considered representative of numerical age of Lutetian/ Ypresian boundary in NW Europe (in Odin, (ed.) 1982: 682). However, the Sables d'Aaltre (= uppermost part of Panisel Formation) lie above the correlative hiatus which marks the boundary between the Ieper (Sables de Mons-en-Pévèle) and Panisel formations and the Wittering-Earnley formations in SE England (see also Islam 1982– 1983). The dated levels are of earliest Lutetian age and stratigraphically equivalent to the basal Lutetian of the Paris Basin. Probably correlative with NP14 (by correlation).
17. G941		47.7 ± 1.6	NP13	NP14	
18. G128	Bruxelles Sands	45.0 ± 2.2	basal NP14	upper NP14	—
19. G104	Wemmel Sands	41.0 ± 1.8	NP15	NP15	—

Ma, a range of 7 m.y. in other words.

2. A distinct hiatus spanning about 3 m.y. separates the biostratigraphically youngest recognizable horizons of the Cuisian (NP12) in the Paris Basin and the Ypresian Fisher Bed IV (NP12/13, Chron C23N) in SE England from the basal Lutetian (upper NP14; above Chron C22N by correlation) in the Paris Basin and Fisher Bed V, below Chron C21N, in SE England (Aubry 1983), respectively.

TABLE 3 Relationship of Paleogene planktonic foraminiferal datum levels to observed magnetic polarity stratigraphy. Age estimates (Ma) in Tables 3 and 4 are derived from revised geomagnetic polarity time-scale presented in Table 1. These data have provided the basic magnetobiochronologic framework for estimating the chronology of standard time-stratigraphic units and stage stratotypes.

PALEOCENE

Datum	Chron	MAGNETIC POLARITY Event and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Morozovella velascoensis*	C24	between anom. 24 and 25 (closer to 25)	57.8	1,3,4	
2. LAD *Planorotalites pseudomenardii*	C25	anom. 25	58.8	1,3	a lone occurrence of *P. pseudomenardii* noted in anom. 25 in Hole 524 (ref. 2).
3. FAD *Planorotalites pseudomenardii*	C26	upper part of C26R (just below anom. 26)	61.0	3,5−8	located in reversed interval just below interval of no palaeomagnetic data (where anomaly 26 should be situated) in Hole 516F (ref. 1).
4. FAD *Morozovella velascoensis*	C26	upper part of C26R	61.7	3	
5. FAD *Morozovella albeari*	C26	upper part of C26R	61.7	3	
6. LAD *Subbotina pseudobulloides*	C26	upper part of C26R	61.7	2,3	
7. FAD *Morozovella pusilla*	C26	mid-part C26R	62.0	5−8	The FAD of this taxon was recorded near the base of C26R (just above anomaly 27 correlative) in Hole 524 (ref. 2) and above an incomplete normal polarity event in Hole 527 interpreted as anom. 27 (ref. 3, 10) or 28 (ref. 11); see further discussion in text.
8. FAD *Morozovella conicotruncata*	C26	mid-part C26R	62.0	2,3	
9. FAD *Morozovella angulata*	C27	lower part C27R	62.3	2,5−8	located at top of interval of no palaeomagnetic data between anom. 27 (above) and 28 (below) in Hole 516F (ref. 1). Recorded in normal polarity interval interpreted as anomaly correlative 27 (ref. 3, 10) or 28 (ref. 11) in Hole 527.
10. FAD *Morozovella uncinata*	C27	top anom. 27	63.0	5−8	located between anom. 27 and 28 in Site 524 (ref. 2) and in anom. 28 in Hole 516F (ref. 1).
11. LAD *Globoconusa daubjergensis*	C27	mid-part of C27R	64.0	2	
12. FAD *Planorotalites compressus*	C28	anom. 28	64.5	3	
13. FAD *Subbotina praecursoria (trinidadensis)*	C28	anom. 28	64.5	2,3,5−8	
14. FAD *Subbotina pseudobulloides*	C29	base anomaly 29	66.1	1−3, 5−8	
15. FAD *Globoconusa daubjergensis*	C29	upper part C29R	66.35	2	
16. FAD *Eoglobigerina (incl. eugubina)*	C29	"	66.35	1−3, 5−8	
17. LAD *Globotruncana*	C29	upper part C29R	66.4	1,3,4−7	precise datum level correlations with magnetic polarity stratigraphy seems to be possible in the Venetian Alps also (ref. 9).

Refs.:
1. Pujol (1983)
2. Poore *et al.* (1983)
3. Boersma (1984)
4. Lowrie *et al.* (1982)
5. Luterbacher & Primoli Silva (1964)
6. Premoli-Silva *et al.* (1974)
7. Premoli-Silva (1977)
8. Premoli-Silva *et al.* (1977)
9. Channell & Medizza (1981)
10. Shackleton *et al.* (1984)
11. Chave (1984)

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Datum	Chron	MAGNETIC POLARITY Event and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Globorotalia cocoaensis*	C13	mid-way between anom. 13 and 15	36.6	2	located at top anom. 15 in Hole 516F, Rio Grande Rise (ref. 1)
2. LAD *Globorotalia cerroazulensis*	C13	mid-way between anom. 13 and 15	36.6	2,3	located at anom. 16 in Hole 516F (ref. 1)
3. LAD *Hantkenina*	C13	mid-way between anom. 13 and 15	36.6	2	located at top anom. 15 in Mediterranean (ref. 3)
4. LAD *Globigerapsis* sp.	C13	between anom. 13 & 15, but below LAD's of cocoaensis-cerroazulensis	37.0	2	
5. LAD *Porticulasphaera semiinvoluta*	C15	anom. 15	37.6	1,3	located in anom. 16 at Site 523 (South Atlantic; ref. 2) and Bottaccione section (ref. 4).
6. LAD *Acarinina* and *Truncorotaloides*	C17	mid-anom. 17	40.6	2	top anom. 18 in Mediterranean (ref. 3, 4).
7. LAD *Morozovella spinulosa*	C17	base anomaly 17	41.1	1	
8. FAD *Porticulasphaera semiinvoluta*	C18	top anom. 18	41.3	3,4	
9. LAD *Subbotina frontosa*	C18	mid-anom. 18	42.0	2	
10. LAD *Globigerapsis beckmanni*	C18	basal anom. 18	42.6	3	
11. FAD *Globigerapsis beckmanni*	C18	just below anom. 18	43.0	3	
12. LAD *Acarinina bullbrooki*	C18	c. 1/4 way down between anom. 18 and 19	43.0	1	
13. FAD *Globorotalia pomeroli*	C19	just above anom. 20	44.7	3	
14. FAD *Globigerapsis index*	C21	upper part of anom. 20	45.0	1	
15. FAD *Morozovella lehneri*	C20	lower part of anom. 20	46.0	3	
16. LAD *Morozovella aragonensis*	C20	lower part of anom. 20	46.0	3,4	
17. FAD *Globorotalia possagnoensis*	C20	lower part of reversed interval between anom. 20 and 21	48.4	2	
18. FAD *Hantkenina*	C22	upper part of anom. 22	52.0	3	
19. FAD *Planorotalites palmerae*	C22	early part of C22R	53.4	4	
20. FAD *Morozovella aragonensis*	C24	anom. 24	55.2	3,4	
21. FAD *Morozovella formosa*	C24	base of anom. 24	56.1	3	

Refs.: 1. Pujol (1983)
2. Poore *et al.* (1982, 1983)
3. Lowrie *et al.* (1982)
4. Napoleone *et al.* (1983)

Note: Eocene planktonic foraminiferal zones tentatively correlated to palaeomagnetic polarity stratigraphy in the Gubbio section of Italy (ref. 4) but precise datum level correlations were generally not made.

3. Odin & Curry (1981: 1004) observe that the Lutetian–Cuisian boundary age estimate of 45–46 Ma is supported by similar high temperature dates on Bridgerian-Uintan rocks of North America (in the range of 43.8–46.6 Ma; see Curry & Odin 1982, Fig. 5).

However, we have shown in this paper that:
1. The Uintan land mammal age is mid- to late middle Eocene in age and post Chron C22N.
2. The Bridgerian land mammal age brackets Chron C21N.
3. High temperature dates of about 48–50 Ma are associat-ed with the section(s) spanning the Uintan– Bridgerian boundary (which lies within the time represented by Chron C20R).
4. The K/Ar (glauconite) dates on the basal Lutetian (43–45 Ma) at a biostratigraphic level correlative with Zone NP14 and within Chron C21R are seen to be *younger* than the high temperature dates (in the range of 48–50 Ma) on terrestrial beds equivalent to, and slightly younger than, Chron C21N.
5. The assumption by Odin & Curry (1981) and Curry &

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Datum	Chron	MAGNETIC POLARITY Event and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. FAD *Globorotalia kugleri*	23	anom. 6C	23.7	1,2	recorded also between anomalies 6C and 7 at Site 522, South Atlantic (ref. 3) and Mediterranean region (ref. 4). This may be, in part, a taxonomic problem (e.g. *kugleri-pseudokugleri* & *mendacis*
2. LAD *Globorotalia mendacis*	23	anom 6C	23.7	1	
3. FAD *Globigerinoides primordius* (common)	23	just below 6C (in normal interval)	24.5	1,2	
4. FAD *Globigerinoides primordius* (rare)	C7	anomaly 7	25.8	1,2,5	recorded between 6C–7 in Mediterranean (ref. 4)
5. LAD *Globorotalia opima*	C9	upper part anom. 9	28.2	3,4,5	between anom. 8 and 9 in Hole 516F, Rio Grande Rise (ref. 2); specimens become smaller above anom. 9 and are not typical of *G. opima* s.s. in Hole 558 (ref. 5)
6. LAD *Chiloguembelina*	C10	mid-anom. 10	30.0	2,3,5	
7. FAD *Globigerina angulisuturalis*	C11	top anomaly 11	31.6	2,5	recorded midway between anom. 10 and 11 in Mediterranean (ref. 4)
8. LAD *Globigerina angiporoides*	C11	near base anom. 11	32.0	2	located at top of anom. 11 (ref. 3)
9. FAD *Globorotalia opima*	C12	mid-anomaly 12	32.7	5	recorded from anom. 11 and as cf. from anom. 12 and slightly lower (ref 3). Also found in anom. 12 in Contessa Quarry revision (ref. 5)
10. LAD *Globigerina ampliapertura*	C12	anom. 12 (lower part)	32.8	3,4	recorded between anom. 11 and 12 in Hole 516F, Rio Grande Rise, South Atlantic (ref. 2) and Site 558, North Atlantic (ref. 5)
11. LAD *Pseudohastigerina*	C12	between anom. 12 & 13	34.0	2,3	an isolated occurrence in anom. 12 is interpreted as reworking (ref. 3) observed also between anom. 12 and 13 at North Atlantic DSDP Site 563 (ref. 5)

Refs.: 1. Berggren *et al.* (1983)
2. Pujol (1983)
3. Poore *et al.* 1982, 1983
4. Lowrie *et al.* (1982)
5. Miller *et al.* (in press): DSDP Site 558 (North Atlantic)

Odin (1982) that basal Lutetian and Cuisian levels are equivalent to Bridgerian–Uintan levels, and that dates on these levels are supportive of an age estimate of 45–46 Ma for the middle–early Eocene boundary is unfounded. The base of the Lutetian has been shown to be associated with Chron C22N, the Bridgerian–Uintan boundary with Chron C20R (a difference of about 3 to 4 m.y.).

Magnetobiostratigraphic correlations of the Paleogene formations of NW European basins (Aubry 1983) have shown that the boundaries between chronostratigraphic units correspond to eustatically controlled unconformities and that hiatuses of moderate to significant duration may be expected to occur in the more marginally located sequences. In the case of the Lutetian–Ypresian (= middle/early Eocene)

boundary this hiatus corresponds to a duration of time which brackets Chron C22N, probably spans the interval of Zone NP13 and lower half of Zone NP14, and represents about 3 m.y. (Aubry 1983).

The overlap in radiometric dates across the boundaries of chronostratigraphic boundaries separated by a hiatus of about 3 m.y., as well as the extensive range of dates (about 7 m.y.) on stratigraphic levels correlative with a single magnetic anomaly (with a duration of about 1.5 m.y.) serves to illustrate the difficulty in using radiochronologic methods in resolving problems requiring precise calibration. Palaeomagnetic stratigraphy (and its derived chronology) can resolve these problems with a distinctly higher degree of resolution, if not accuracy.

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TABLE 4   Relationship of Paleogene calcareous nannoplankton datum levels to observed magnetic polarity stratigraphy.

PALEOCENE

Datum	Chron	MAGNETIC POLARITY Event and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Fasciculithus*	C24	in reversed interval (approximately midway) between anom. 24 & 25	57.4	3,4,6	
2. FAD *Ericsonia robusta*	C24	lower part of reversed interval between anom. 24 and 25	58.6	4	
3. FAD *Discoaster multiradiatus*	C25	base anom 25	59.2	1,4	FAD *D. multiradiatus* recorded in upper anom. 25 (refs. 2, 5, 6).
4. FAD *Discoaster nobilis*	C25	in reversed interval just below anom. 25	59.4	5,6	
5. FAD *Heliolithus riedeli*	C25	in reversed interval above anom. 26	60.0	4	FAD *H. riedeli* recorded in reversed interval below anom. 25 and just above interval of no polarity data (probably corresponding to anom 26) in Hole 516F (ref. 1).
6. FAD *Discoaster mohleri*	C26	mid-anom. 26	60.4	4,5	FAD *D. mohleri* recorded in reversed interval below anom. 25 (anom. 26 not identified between 25 and 27) in Hole 524 (ref. 2); recorded in interval of no polarity data within interval of reversed polarity between anomalies 25 and 27 in Hole 516F (ref. 1); recorded in upper third of reversed interval between anom. 26 and 27 in Hole 577A (ref. 6).
7. LAD *Chiasmolithus danicus*	C26	in reversed interval below anom 25 (between 25 & 27)	61.0	2	LAD *C. danicus* recorded in Zone NP6 in Hole 524 (ref. 2), which in Hole 527 (South Atlantic) occurs in later half of reversed interval between anom. 27 & 26 and early part of anom. 26 (ref. 4). Anom. 26 not identified in Hole 524 (ref. 2).
8. FAD *Heliolithus kleinpelli*	C26	slightly above mid-point of C26R	61.6	4,6	FAD *H. kleinpelli* recorded in reversed interval between anom. 25 & 27 in South Atlantic Holes 516F (ref. 1) and 524 (ref. 2). Recorded in early part of anom. 26 in Bottaccione and Contessa Highway sections (ref. 5) and Hole 527 (ref. 7).
9. LAD *Cruciplacolithus tenuis*	C26	in reversed interval between anom. 25 & 27	61.8	2	LAD *C. tenuis* recorded in Zone NP5 in Hole 524 (ref. 2) which occurs in mid-part of reversed interval between anom. 26 & 27 in Hole 527 (ref. 4). Anom. 26 not identified in Hole 524 (ref. 2).
10. FAD *Fasciculithus tympaniformis*	C26	mid-part of C26R	62.0	4,5,6,7	FAD *F. tympaniformis* recorded in lower part of reversed interval between anom. 25 & 27 (anom. 26 not identified in Holes 516F (ref. 1) and 524 (ref. 2)).
11. FAD *Ellipsolithus macellus*	C27	mid-part of C27R	63.8	5	FAD *E. macellus* recorded just above anom. 27 in Hole 516F (ref. 1), 524 (ref. 2), 527 and 528 (ref. 4), and 577 (ref. 6) but this is a solution susceptible taxon.
12. FAD *Chiasmolithus danicus*	C28	early part anom. 28	64.8	2,4,6,7	Recorded in later part of anom. 29 in DSDP Site 516F, Rio Grande Rise (South Atlantic) (ref. 1) and Bottaccione section (ref. 5); in mid-part of combined anom. 28–29 in Contessa Highway section (ref. 5).
13. FAD *Cruciplacolithus tenuis*	C29	mid-anom. 29	65.9	1,4,6	FAD *C. danicus* (= NP3) and *C. tenuis* (= NP2) coincide in Hole 524 and therefore FAD of *C. edwardsi* and *C. pelagicus* (s.ampl.) are used to denote base NP2 (ref. 2).
14. FAD *Cruciplacolithus primus*	C29	base anom. 29	66.1	7	
15. FAD *Placozygus sigmoides*	C29	later part of reversed interval (about 2/3 way up) between anom. 29 & 30	66.4	2	

16.	*Biantholithus sparsus*	C29	"	66.4	7
17.	LAD *Micula murus*	C29	"	66.4	2,4,8
18.	LAD *Lithraphidites quadratus*	C29	"	66.4	2,8

Refs.: 1. Cepek (written communication 1982)
2. Poore *et al.* (1983)
3. Lowrie *et al.* (1982)
4. Shackleton *et al.* (1984)

5. Monechi & Thierstein (*in press*)
6. Monechi *et al.* (*in press*)
7. Manivit & Feinberg (1984)
8. Manivit (1984)

EOCENE

Datum	Chron	MAGNETIC POLARITY Event and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Discoaster barbadiensis*	C13	mid-way between anom. 13 & 15	36.7	2,3,4	LAD *D. barbadiensis* recorded in interval of no polarity data above anomaly 15 in Hole 516F (ref. 1).
2. LAD *Discoaster saipanensis*	C13	mid-way between anom. 13 & 15	36.7	2,3,4	
3. FAD *Isthmolithus recurvus*	C15	base anomaly 15	37.8	2,3	FAD *I. recurvus* recorded at top of anom. 16 in Contessa Highway and Bottaccione sections (ref. 5).
4. FAD *Chiasmolithus oamaruensis*	C17	late anomaly 17	39.8	2	
5. LAD *Chiasmolithus grandis*	C17	later part of anomaly 17	40.0	1,2	LAD *C. grandis* reported in later part of anom. 18 in Mediterranean (ref. 3 and 5).
6. LAD *Chiasmolithus solitus*	C18	lower part of anom. 18	42.3	2	
7. LAD *Nannotetrina fulgens*	C20	mid-anom. 20	45.4	2	
8. FAD *Reticulofenestra umbilica*	C20	basal part of anom. 20.	46.0	5	Recorded between anom. 20 & 21 in Contessa Road Section (ref. 3) and midway between anom. 19 and 20 in Contessa Quarry section (ref. 3) and DSDP Site 523 (ref. 2). Aubry (pers. comm., 1983) indicates that the FAD of this taxon is latitudinally dependent.
9. LAD *Chiasmolithus gigas*	C20	in reversed interval about 1/3 way down between anom. 20 & 21	47.0	2	Range of *C. gigas* found to occur within reversed interval between anom. correlative 20 & 21 in Hole 527 (ref. 4).
10. FAD *Nannotetrina fulgens*	C21	lower part anom. 21	49.8	1,4,5	FAD of *Nannotetrina* sp. recorded in lower part of anom. 21 in Contessa Road, (Gubbio; ref. 3) and *Nannotetrina fulgens* in anom. 20 in Contessa Quarry section (Gubbio, ref. 3). Preservational problems precluded taxonomic identification to the species level in the Contessa Road section and the appearance of *N. fulgens* in the Contessa Quarry section is not considered a FAD because the base of the section is within the normal polarity event interpreted as anom. 20 correlative (K. Perch-Nielsen, pers. comm. 1983).
11. FAD *Discoaster sublodoensis*	C22	early part of anom. 22	52.6	5	FAD *D. sublodoensis* recorded just above anom. 22 correlative in Hole 527 (ref. 4), in early part of anom. 22 in Contessa Highway section (ref. 5) but in reversed interval just above anom. 23 in Bottaccione section (ref. 5).
12. LAD *Tribrachiatus orthostylus*	C22	just above anom. 23	53.7	4,5,6	LAD *T. orthostylus* recorded in anom. 23 in Contessa Road section (ref. 3); recorded just above anom. 23 in Site 527 (ref. 4), Contessa Highway section (ref. 5); and Hole 577 (ref. 6). Youngest occurrence in England observed near top of Fisher Bed IV (= *Nummulites planulatus* Beds) of Bracklesham Group,

					Whitecliff Bay (ref. 7). A normal polarity event (polarity interval 'd' = Wittering magnetozone; Townsend 1982), interpreted as anomaly 23 correlative (ref. 7) straddles Fisher Bed IV. Thus the youngest occurrence of *T. orthostylus* at Whitecliff Bay occurs near the top of, or just above, anomaly 23 correlative consistent with refs. 4–6 cited here.
13. FAD *Discoaster lodoensis*	C24	early part C24N-1	55.3	1,3,5	Recorded just above C24N-2 in Hole 527 (ref. 4) and Hole 577 (ref. 6) which are essentially the same level as that cited here (refs. 1,3,5).
14. FAD *Tribrachiatus orthostylus*	C24	in reverse interval just below base anom. 24	56.6	1,3,4,5	Datum events 14 and 17 recorded sequentially in the South Atlantic (ref. 4) and Mediterranean (refs. 3, 5). FAD *T. orthostylus* recorded in C24N-2 subchron in Hole 577 (ref. 6).
15. LAD *Tribrachiatus contortus*	C24	in reversed interval just below anom. 24	56.3	4	
16. FAD *Campylosphaera dela*	C24	in reversed interval just below anom. 24	56.4	4	
17. FAD *Discoaster diastypus*	C24	reversed interval below anom. 24	56.5	4,5	
18. FAD *Discoaster contortus*	C24	reversed interval below anom. 24	56.8	4	
19. LAD *Fasciculithus*	C24	in reversed interval (approximately midway) between anom. 24 & 25	57.6	3,4	

Refs.:　1.　Cepek (written communication 1982);
　　　　2.　Poore *et al.* (1982, 1983);
　　　　3.　Lowrie *et al.* (1982);
　　　　4.　Shackleton *et al.* (1984);
　　　　5.　Monechi & Thierstein (*in press*);
　　　　6.　Monechi *et al.* (*in press*).
　　　　7.　Aubry (1983).

Appendix III Magnetobiochronology of late Cretaceous stage boundaries (Maestrichtian – Campanian and Campanian – Santonian).

Assessment of the magnetochronology of these boundaries, while somewhat beyond the scope of this paper, is included here to complete the discussion of our revised late Cretaceous to Recent time-scale (see also companion paper on the Neogene, Jenkins *et al.* (this volume)). The biostratigraphic framework of late Cretaceous stages has been discussed by Berggren (1964) and more recently by Thierstein (1976), Van Hinte (1976) and Sissingh (1977, 1978).

Based on K-Ar dates on bentonites associated with various baculitid and ammonite zones, Obradovich & Cobban (1975) suggested ages which recalculate to 71 Ma, about 72 to 73 Ma, and about 74 to 75 Ma for the Campanian–Maestrichtian boundary (ages are corrected ages according to tables in Dalrymple 1979). The age estimate of about 71 Ma is based on dates in the western interior of Canada on stratigraphic levels equivalent in the United States to the *Baculites grandis* Zone. The Campanian–Maestrichtian boundary in the western interior of Canada was provisionally drawn by Jeletzky (1968) at a level equivalent in the United States to the boundary between the *Baculites eliasi* and *B. baculus* zones, the next two zones below the *B. grandis* zone (see Obradovich & Cobban 1975: 47). However, this boundary is incorrectly correlated to the stratotype Maestrichtian and is clearly too young in terms of chronology. Jeletzky (1951) has

shown that the base of the Maestrichtian Stage coincides with the base of the *Belemnella lanceolata* and *Acanthoscaphites tridens* Zone which is correlative, in turn, with the top of the *Globotruncana calcarata* Zone (Van Hinte 1976). The base of the Maestrichtian is correlative also with the initial appearance of *Rugotruncana subcircumnodifera* (Berggren 1962). Pessagno (1967, 1969) recognized this boundary in the Gulf Coast using the same criterion. But the boundary determined in this way corresponds approximately to the boundary between the *Didymoceras nebrascense* and *D. stevensoni* zones in the United States, some eight zones below the boundary as correlated by Jeletzky (1968) (Obradovich & Cobban 1975: 48). Thus the correlation made by Jeletzky (1968) from the Western Interior to the stratotype Maestrichtian is clearly too young. That made by Pessagno (1967, 1969) is more nearly correct and the age estimate of approximately 74 to 75 Ma is based on radiometric dates made on the *D. nebrascense* and the *Exiteloceros jenneyi* Zones (two zones above). Olsson (1964) determined the Campanian–Maestrichtian boundary in New Jersey at a level correlated with the *Baculites scotti* Zone, a zone below the *D. nebrascense* Zone. The radiometric dates on these biostratigraphic levels can serve as the basis for geochronologic estimates of the age of the Campanian–Maestrichtian boundary, not those made at the level of the *B. grandis* Zone.

The Campanian–Santonian boundary has been dated (K-Ar date on bentonite in the *Desmoscaphites bassleri* Zone from the Western Interior of the United States) at 84.5 Ma

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Datum	Chron	MAGNETIC POLARITY Event and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Zygrhablithus bijugatus*	23	about 1/3 way down between anom. 6C & 7	24.6	1,4	
2. LAD *Sphenolithus ciperoensis*	23	just above anom. 7	25.2	1,3,4,5	LAD recorded just below anom. 6C at Site 511 in South Atlantic (ref. 2) but last common occurrence shown at anom. 7.
3. LAD *Sphenolithus distentus*	C9	top anom. 9	28.2	2,5	LAD *S. distentus* recorded between anom. 9 & 10 in Mediterranean (ref. 3) and within anom. 10 in Hole 558 (ref. 4).
4. FAD *Sphenolithus ciperoensis*	C10	early anom. 10	30.2	2	FAD *S. ciperoensis* recorded in late anom. 9 in Mediterranean (ref. 3), at base anom. 11 in Hole 558 (ref. 4), and in interval of no polarity data between anom. correlative 10 & 11 in Hole 528 (ref. 5).
5. FAD *Sphenolithus distentus*	C12	midway in the reversed interval between anom. 12 & 13	34.2	2,3	The anomalous reversed order of FAD *S. ciperoensis* and LAD *S. distentus* in the Mediterranean was noted in ref. 3. Recorded just above anom. 13 (*ca.* 35.0 Ma) in Hole 558 (ref. 4).
6. LAD *Reticulofenestra umbilica*	C12	between anom. 12 & 13 (*ca.* 1/3 distance above anom. 13)	34.6	2,4,5	
7. LAD *Reticulofenestra hillae*	C12	"	34.7	2,4	
8. LAd *Isthmolithus recurvus*	C12	a short distance above anom. 13	34.9	3,4	LAD *I. recurvus* recorded in upper part anom. 13 in Site 522 in the South Atlantic (ref. 2).
9. LAD *Ericsonia formosa*	C12	just above anom. 13	35.1	2,3,4,5	LAD *E. formosa* recorded near base anom. 13 in Hole 516F (ref. 1).

Refs.:   1. Berggren *et al.* (1983)
2. Poore *et al.* (1982. 1983)
3. Lowrie *et al.* (1982)
4. Miller *et al.* (*in press*)
5. Shackleton *et al.* (1984)

which led Obradovich & Cobban (1975: 47) to suggest an age (recalculated) of about 84 Ma for the Campanian−Santonian boundary.

Direct correlation between the biostratigraphic and magnetostratigraphic record was treated initially by Alvarez *et al.* (1977) for the Cenomanian to Maestrichtian interval at Gubbio, Italy. They show the Maestrichtian−Campanian boundary (*G. calcarata/G. tricarinata* zonal boundary) in the upper part of the Gubbio normal zone B+, which is correlated to the youngest part of Chron C33N, and place the Campanian−Santonian boundary (*Globotruncana carinata/G. elevata* zonal boundary) at a level just below the top of the Gubbio Long Normal Zone (Chron C34N). Supporting evidence for these correlations has been obtained by Channell & Medizza (1981) from the Caroselle section in the Venetian Alps and by Berggren *et al.* (1983) from DSDP Site 516F in the South Atlantic.

In the derivation of a revised geomagnetic reversal time-scale, we have assumed an age (corrected) of 84.0 Ma for both the Campanian−Santonian boundary and the top of Chron C34N according to the information outlined above.

Thus, the magnetochronologic and biochronologic age estimates for the Campanian−Santonian boundary are made identical and no meaningful comparison of such age estimates can be made. If a change were to be required either in the biochronologic age estimates for the Campanian−Santonian boundary or in the correlation of this boundary to the top of Chron C34N, then according to our methodology, a corresponding change would need to be made in the magneto-chronology of the earliest of the three distinct segments of the geomagnetic polarity time-scale.

It is, however, possible to make a more meaningful assessment of age estimates for the Maestrichtian−Campanian boundary since this level was not used to calibrate the geomagnetic polarity time-scale. According to the revised geomagnetic polarity time-scale, placement of this boundary near to the top of Chron C33N gives a magnetochronological estimate of 74.5 Ma (Fig. 1). This estimate compares very favourably with the biochronological estimate for the Maestrichtian−Campanian boundary of 74−75 Ma (corrected age) according to the preferred correlations of Pessagno (1967, 1969).