The Geological Society Memoir Number 10

The Chronology of the Geological Record

EDITED BY N.J.SNELLING

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Preface

The measurement of geological time and the construction of a geologic time-scale composed of standard stratigraphic divisions based on rock sequences and calibrated in years has long attracted the attention of geologists and has done much to provoke international co-operation. Thus the Committee for the Measurement of Geological Time, set up in December 1923 by the National Research Council of the USA almost immediately attracted the co-operation of the pioneers concerned with the dating of rocks by radioactive decay and served as a world-wide forum through to the 1950s. The first steps to establish a chronostratigraphic scale were taken much earlier at the International Geological Congress held in Bologna (Italy) in 1881 and both of these activities are now co-ordinated through the various Sub-commissions of the Commission on Stratigraphy of the IUGS.

Progress in nuclear physics and the development of new tools for isotope research saw the effective birth of *isotope geology* in 1950 (Rankama 1954) and the following decade saw a major data explosion in this subject. Much of this early work was essentially geochronometric though not necessarily directed towards the establishment of a geologic time-scale. However, the possibility of improving the time-scale, at that time the virtual brain child of one man — Arthur Holmes — resulted in the holding of an interdisciplinary symposium by the Geological Society of London, and the subsequent publication in 1964 of *The Phanerozoic Time-Scale* followed by a supplement in 1971. Important aspects of these influencial publications were the inclusions of over 300 *items* or 'abstracts of published radiometric and stratigraphic data with comments' which now constitute the foundation data bank for virtually all time-scale publications.

A notable problem apparent in these pioneering works is the considerable element of uncertainty introduced by differing opinions concerning the numerical values of the decay constants, particularly for potassium-40 and rubidium-87, the two parent isotopes with the widest application in practical geochronometry. The use of the differing decay constants for the same analytical data could result in a discrepancy of about 30 Ma for Palaeozoic rocks. Fortunately the vigorous persual of this problem by a few researchers, backed by pressure from the Subcommission on Geochronology which circulated questionaires for surveys, resulted in the presentation at the International Geological Congress in Sydney, Australia, in 1976 of a 'Convention on the Use of Decay Constants in Geochronology and Cosmochronology' (Steiger & Jäger 1977) which has since been universally adopted. The effective resolution of this problem has prompted re-evaluations of the 1964 time-scale, notably by Harland, Cox, Llewellyn, Pickton, Smith, and Walters (1982) and by Odin (1982). The former work relies heavily on a data base generated by Armstrong (1978) which re-evaluated *The Phanerozoic Time-Scale* data bank eliminating nearly half of the pre-Cenozoic data but replacing them by as much and more new data. The latter work also reassesses the previous data and in addition incorporates a new data bank of 251 items with very detailed comment on, and evaluation of, their radiometric and stratigraphic significance.

The contributors to this volume have among other things attempted a further reassessment of the aforementioned data bases. Hopefully this iterative process will in time produce an accurate time-scale, but at the moment it seems more realistic to view the suggested summary/compromise time-scale as ephemeral. A particularly noticeable aspect of this volume is the combined use of ocean floor spreading and reversals of the Earth's magnetic field as secondary time keepers. It is already clear that this seconday clock will allow a very fine resolution of Cenozoic time, but it is equally clear that the clock has yet to be accurately calibrated. Consideration in this volume is also given to the problems of the Precambrian time-scale and it would seem that following the deliberations of the Subcommission on Precambrian Stratigraphy and the findings of IGCP Projects 99 and 118 (Geochronological Correlation of Precambrian Sediments and Volcanics in Stable Zones, and Upper Precambrian Correlations respectively) the geological community is on the threshold of a significant advance, albeit one that may be proceded by valuable controversy.

Although there is still room for improvement in the accuracy and resolution of the time-scale it is nevertheless sufficiently accurate to allow very useful estimates of the rates at which many geological processes proceed to be made, some of the these processes are the subject of papers in the second part of this volume.

It is inevitable that the task of drawing up a geological time-scale will remain a collaborative and co-operative effort and it is encouraging that the hope expressed in *The Phanerozoic Time-Scale* 'that this co-operative venture will continue' continues to be realized. The efforts of many of the authors to produce up to the minute reviews of their subject is particularly appreciated by the sponsors of this symposium.

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Geochronology and the geological record

N.J. Snelling

Some history

By about the middle of the nineteeth century natural scientists had come to the conclusion that geological time was to be measured in hundreds of millions of years. Darwin in the first edition of *The Origin of Species* published in 1859 estimated that 300 million years had elapsed since the end of the Mesozoic Era; Lyell concluded in 1867 that about 240 million years had elaspsed since the beginning of the Ordovician Period. Despite the temptation to point out the obvious discrepancy in these particular estimates and to suggest that they were no more than educated guesses it must be remembered that they reflected the most careful marshalling of a vast number of verifiable observations. Indeed they can now be seen to be remarkably realistic and in marked contrast to the estimates based on more theoretical considerations such as ocean salinity, sediment thickness, the cooling of the Earth or the Sun's energy source.

Towards the end of the nineteenth century the geologists' views regarding the age of the Earth came under particularly strong attack from W. Thomson (later Lord Kelvin) whose arguments were based on the energy source of the Sun and the cooling history of the Earth, and who concluded that the Earth was probably younger than about 100 million years. That geologists found Thomson's estimates thoroughly unpalatable is very clear from the literature of the time but his calculations were apparently irrefutable. However, the impasse was soon to be resolved following the discovery of the phenomenon of radioactivity by Becquerel in 1896 and the realization during the following decade that here was an unforeseen heat source which completely negated the basis of Thomson's calculations. More important, however, were the various positive aspects of this discovery, within the same decade Rutherford realized that the accumulation of the products of radioactive decay would provide a means of dating minerals containing radioactive elements and by 1905 he had calculated the first age, based on the accumulation of helium in a uraniferous fergusonite. By 1907 similar determinations by Boltwood indicated that the Palaeozoic Era occurred some 400 to 500 million years ago and that Precambrian rocks might be as old as 2000 million years. In 1911 Arthur Holmes, with both the enthusiasm of youth and the foresight of genius, was to write, '...... data will be collected from which it will be possible to graduate the geological column with an ever-increasingly accurate time-scale'.

Not all geologists, however, were as optimistic as Holmes and indeed it clearly took some time for the significance of Rutherford's idea to 'catch on' among geologists. Giekie, writing in the now famous 1911 eleventh edition of the Encyclopaedia Britannica appeared to be unaware of Rutherford's work but it must be remembered that this was Edwardian Britain and Rutherford was still in his thirties, a young New Zealander working in a Canadian University. Gradually, however, the possibility of dating minerals became apparent to the geological establishment, though one suspects that many geologists found this exciting new development scientifically far too high powered to be assimilated. During the first half of the present century 'geochronology' was the domain of a small group of enthusiasts of which Arthur Holmes was to become the doyen, in stature if not in age, and who reported to the Committee on the Measurement of Geological Time, established by the National Research Council of the USA.

It is just as well that the early geochronologists were scientists of high ability. The analytical work pushed their experimental skills to the limit and interpretation was confused by helium leakage and by uncertainties as to how much of the lead present in an analysed mineral was due to the decay of uranium, how much was due to the decay of thorium and how much was original. In addition, the decay constants were not that well known and indeed a reasonable resolution of this particular problem did not come about until the 25th International Geological Congress (IGC) in 1976 (see Steiger & Jager 1977). The pioneers also had to cope with some scepticism from their colleagues as to whether or not the decay constant (λ), a measure of the proportion of atoms disintegrating in unit time, was indeed a constant under all geological conditions or had remained a constant throughout geological time. The familiar pleochroic halos were to bring about the resolution of this problem. They occur in various minerals surrounding minute uranium and thorium bearing mineral inclusions, and are due to the structural breakdown of the host mineral brought about by the emission of energetic alpha particles. The distance travelled by the alpha particles bears a definite relationship to the decay constant of the disintegrating parent nuclide (the Geiger-Nuttall rule). In detail the halos consist of various rings with discrete and constant radii, each related to a particular alpha-emitting parent. The constancy of the radii of these rings, irrespective of the age of the host mineral, confirmed that the decay constants of alpha emitting isotopes must be invariant with time (see Rankama 1954 for a complete discussion of this important subject).

The modern era of geochronology came into being as the result of the technological breakthrough of Alfred Nier who in the 1930s developed the modern sector mass spectrometer. This instrument made possible the distinction between the radiogenic and non-radiogenic isotopes of the same element, and later, with the production of artifically enriched isotope tracers (spikes) provided a means of accurate and precise analysis of both parent and daughter elements for all the decay schemes of geochronometric interest. The potential of this instrument has been widely and rapidly exploited and together with other technological developments the analytical aspects of geochronology have become routine and to a considerable extent automated.

Time-scales

Attempts to give numerical values for the ages of the 'Traditional Stratigraphic Scale' were first attempted by Holmes in 1911 and continue to present a challenge to the earth scientist. Some of the recent Phanerozoic time-scales are given in Fig. 1 together with one of the early scales by Holmes & Lawson (1927) for comparison. Conventional decay constants have been widely adopted following the 25th

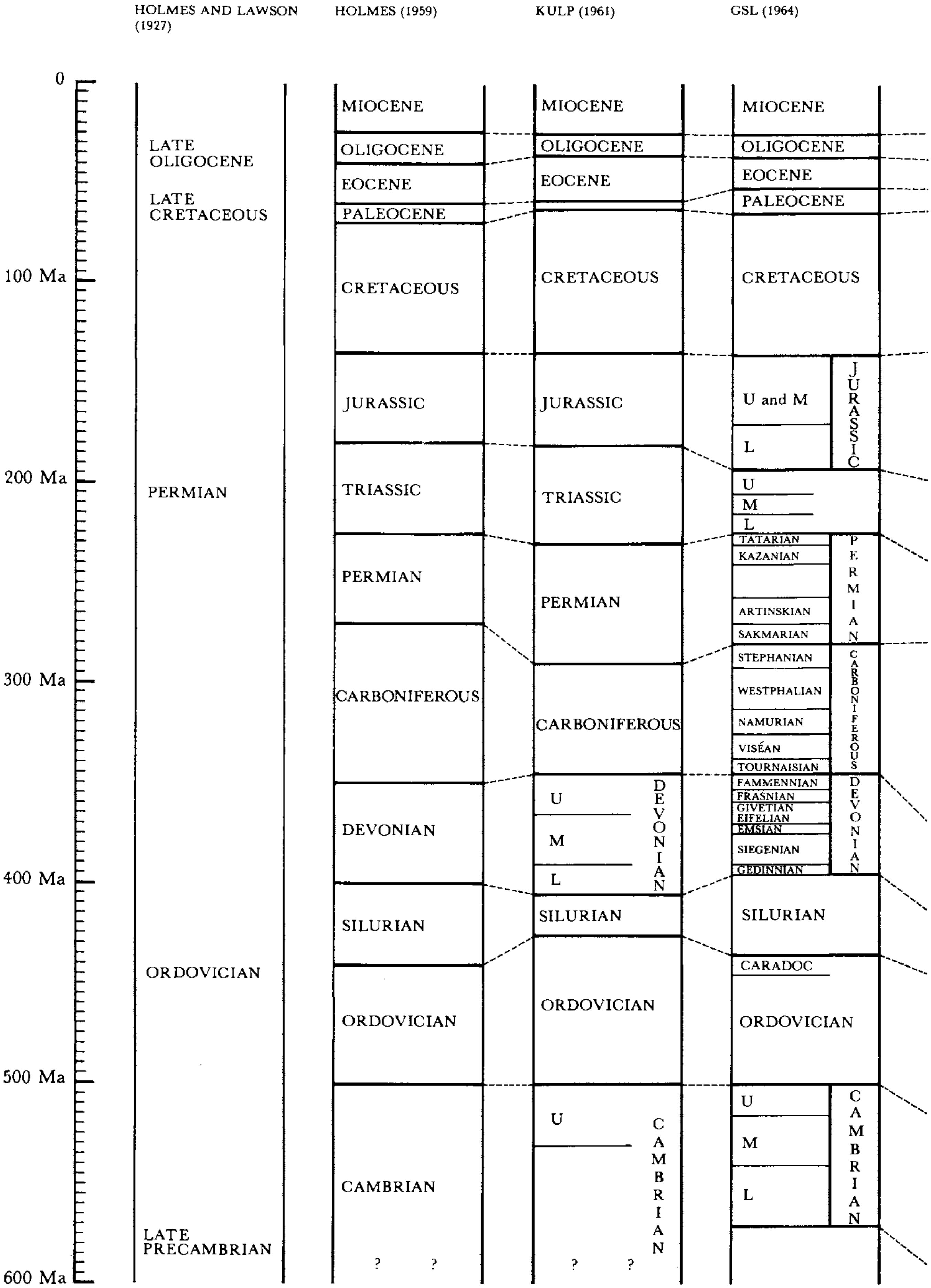
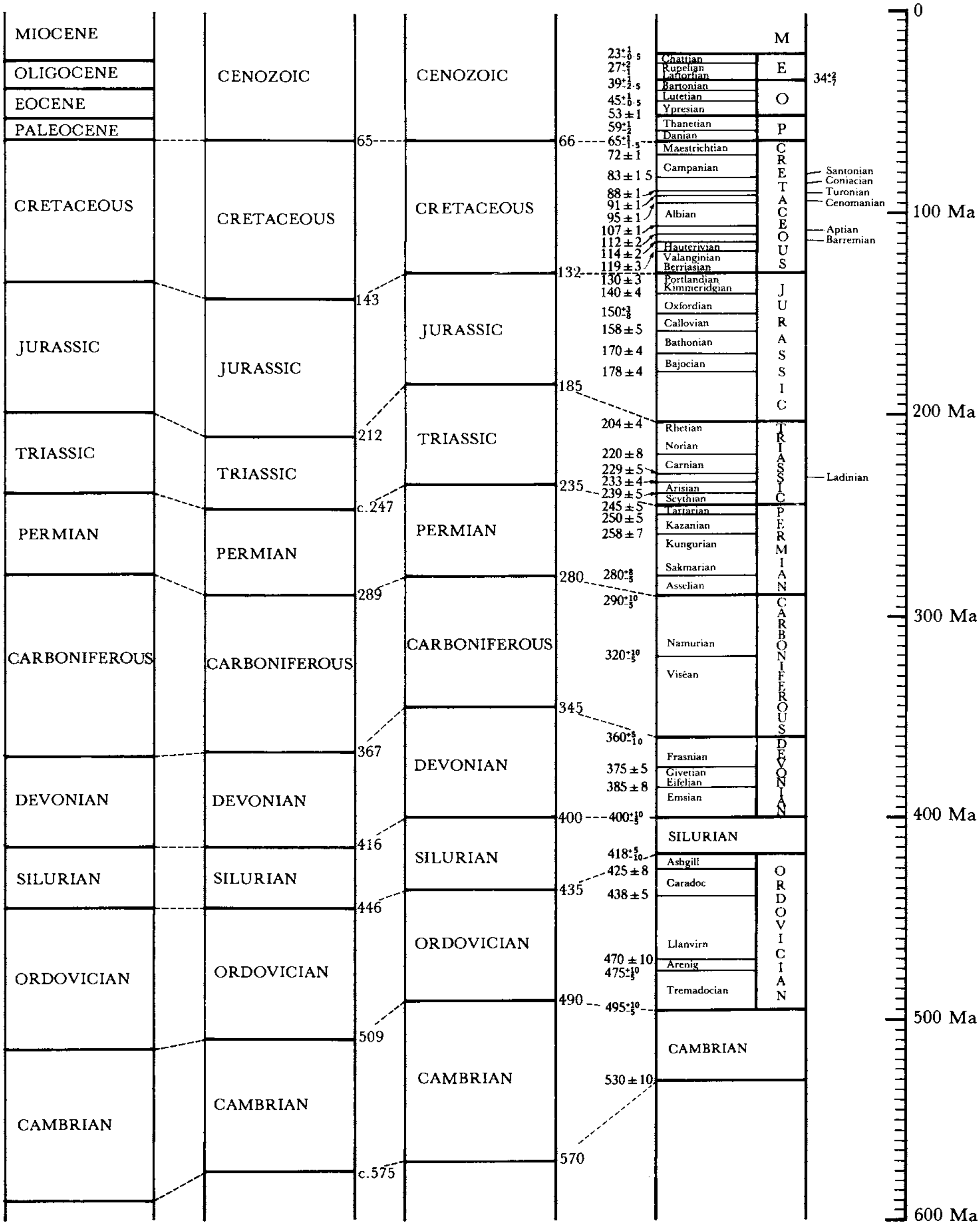


FIG. 1. Phanerozoic time-scales.

LAMBERT (1971) ARMSTRONG (1974) AFANAS'YEV AND ZYKOV (1975) ODIN *et al.* (1982)



IGC and the last three scales in Fig. 1 are based on these constants. Despite this the age estimates of some of the system boundaries still show significant differences. The discrepancies reflect both different or uncertain assessments of the stratigraphic position of certain key dated rock bodies, and different interpretations of the meaning and reliability of certain ages. The scale by Afanas' yev & Zykov (1975) tends to assign younger dates to the system boundaries than are found in the other recent scales. Armstrong (1978) suggested that this might be due to the heavy reliance of Russian workers on glauconite and whole rock K:Ar ages both of which can be too young due to argon loss.

A particular problem is presented when attempting to date the base of the Cambrian which simply reflects the difficulties of defining this boundary in stratigraphic and palaeontological terms. It now seems likely that the boundary will be placed near the first incoming of Tommotian-type (s.l.) fossil assemblages with archaeocyathids and small fossils (Cowie 1981). By convention this may also be adopted as the base of the Palaeozoic although as Cloud (1976) and other workers have pointed out, underlying strata occur in some places equivalent in time span to about 150 Ma, and contain metazoa which could rightly also be considered as Palaeozoic. Such strata have been assigned to the Ediacarian system (Termier & Termier 1960; Cloud & Glaessner 1982) by some workers and to the Vendian (Sokolov 1968, 1972) by others.

Unlike the Phanerozoic, Precambrian time is entirely without formal subdivision and indeed there is no concensus of opinion as to how it should be divided. Most recent attempts have adopted essentially local successions of orogenies (or major tectono-metamorphic-plutonic events), or equally local successions of strato-types. These are both extensions of the practical approach adopted by geologists in the field and because they reflect practice in the field they have much to recommend them. They can be applied over remarkably wide areas in all the continents. Both types of phenomena, however, are diachronous and the scales break down when attempts are made to extend them to a world wide scale. Such attempts result in compromises and even negative definitions of boundaries 'chosen so as to interrupt as few as possible major sequences of sedimentation, igneous emplacement, or orogeny' (Sims 1979). The writer has little enthusiasm for the application of such boundaries on a world wide scale. In the Phanerozic, fossils are used for the purely practical purpose of correlation and geologists did this effectively long before age determinations became commonplace. In the Precambrian, age determinations are now commonly available and provide an adequate basis for correlation. It is surely sufficient to be able to say that an event or succession in one continent is more or less contemporaneous with an event or succession in another; to say that such events or successions are 'Archaean' or 'Lower Proterozoic' is significant only if some independent, contemporaneous, and world wide phenomona are also associated with the particular terminology.

Despite these personal views it seems that the terms Archaean and Proterozoic are now so firmly established in the geological vocabulary as to be virtually unassailable and the Subcommission of Precambrian Stratigraphy has recommended acceptance of a time boundary at 2500 Ma (James 1978). This age seems to have been adopted essentially as a convention though it seems likely that most earth scientists would accept, at least in broad general terms, the view that 'the Archaean was terminated by a diachronous wave of craton-forming sialic plutonism and probably major volcanism and outgassing between 2.5 and 3 G.y. ago' (Cloud 1976). The lower boundary to the Archaean can probably be set at about 3800 Ma, essentially the limit of the geological record on Earth.

The various subdivisions of the Proterozoic that have been proposed mainly reflect local intra-continental sequences of orogenic events or stratotypes. However, an interesting attempt at a rational subdivision has been advanced by Cloud (1976, 1978) and is coupled with a strongly reasoned case for replacing the Proterozoic Eon by two eons, viz the Proterophytic and the Palaeophytic, the names of which denoted levels of floral evolution, with a boundary between the two at about 2000 Ma. The lower boundary of the Proterophytic would now be taken at 2500 Ma although phenomena appropriate to the Proterophytic occur well into the Archaean. The upper boundary of the Palaeophytic could correspond to the base of the Cambrian or perhaps more appropriately the base of the Ediacarian. If the base of the Cambrian is synchronized with the base of the Palaeozoic traditionalists might like to retain the name Proterozoic for the hundred or so million years before the Cambrian when unambiguous metazoa first appeared.

Cloud's scheme has the attraction of dividing an extremely long period of time into two more manageable periods divided by a world-wide, independent, and more-or-less synchronous phenomenon clearly reflected in the lithological characteristics of the stratigraphic record. This phenomenon is the extension of biogenic oxygen from the hydrosphere to the atmosphere. During the Proterophytic, weathering in a more-or-less oxygen free atmosphere resulted in ferrous iron being transported in solution into the oceanic reservoir. In suitable environments, probably epicontinental seas, much of this iron was precipitated in the ferric state by oxygen generated by oxygen-releasing photosynthesizing organisms giving rise to the banded iron-formations (a grossly simplified view of a very complex problem; see Maynard, 1983). By about 2000 Ma ago the evolution of oxygen-mediating enzymes released the primative photosynthesizing organisms from the quasi-symbiotic relationship with ferrous iron, allowing a rapid (tachytelic) evolutionary expansion of primitive algae. Ferrous iron ceased to be necessary for the support of such life forms, and as the rapidly spreading and evolving life forms swept the oceans free of ferrous iron, oxygen was released into the atmosphere. This dramatically reduced the input of ferrous iron to the oceans as iron released in the weathering cycle was oxidized and fixed in insoluble ferric forms, which found lithological expression in the first true red-beds (see also Cloud 1976, 1978; Schopf 1978).

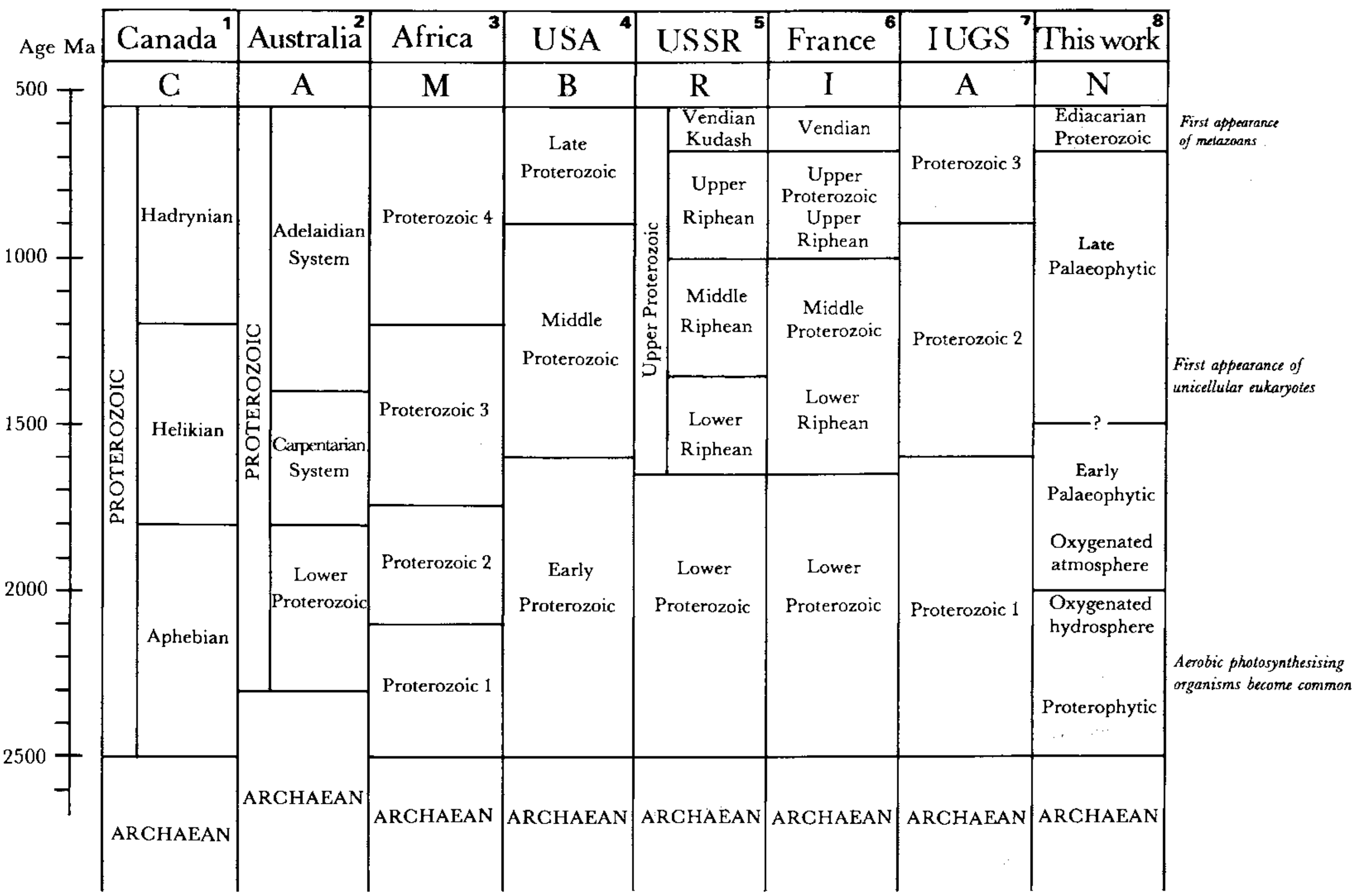
The Palaeophytic still remains an inordinately long period of time and further subdivision is probably desirable. Undoubtedly the most significant event during this period, certainly the most important event as far as we human beings are concerned, was the evolution of the eukaryotes some time between 1000 and 1500 Ma ago (Schopf 1978). Unfortunately this dramatic evolutionary step is very difficult to detect and is of no practical use to the geologist working in the Precambrian. Nevertheless so important an event surely deserves to be enshrined in the time-scale and I suggest it be used to divide the Palaeophytic into Early and Late divisions at a point in time *yet to be decided* but somewhere about 1500 Ma.

Geologists concerned with the practicalities of field work will undoubtedly continue to divide the Proterophytic and Palaeophytic on the basis of orogenic cycles and or stato-types. Well established biostratigraphic schemes, particularly for the Upper Palaeophytic, already exist and hold out the promise of inter-continental correlation (Bertrand-Sarfati 1981; Bonhomme & Bertrand-Sarfati 1982). The established biostratigraphy reflects the widespread occurrence of relatively undisturbed and unmetamorphosed Palaeophytic sediments and volcanics which hold out the hope that eventually palaeomagnetic subdivisions will also become possible.

Some of the more recent Precambrian time-scales are summarized in Fig. 2.

Rates of geological processes

Once methods of determining geological ages and expressing them in conventional units became possible the rates at which geological processes occurred could be assessed without resort to undue speculation. However, earlier discussions of the rate of accumulation of salt in the oceans, and the rates of sedimentation are now seen to be relatively naive. A far more detailed and essentially dynamic view of process rates can now be taken. Sedimentary rocks, the oceans, and indeed whole segments of the crust must be looked upon as geochemical reservoirs with variable survival prospects and we can further give detailed consideration to the residence times of the individual elements in these reservoirs. In a closed-



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7 Sims 1979.
8 Cloud 1976; Schopf 1978.

FIG. 2. Some Precambrian time-scales.

system-world subject to a dramatic increase in the level of environmental pollution such studies may be vital to our survival, and we would perhaps be wise to reverse Llyle's famous dictum and look upon the past as the key to the present.

Our ability to date rocks with relative ease has for some time now encouraged attempts to define apparent polar wandering paths during the Precambrian. Such studies in Phanerozoic rocks established the reality of Continental Drift and ushered in the concepts of ocean floor spreading and plate tectonics. Did such processes occur during the Precambrian? Did they operate at a faster rate simply because the Earth was generating more heat than in recent times and if so did this effect the overall tectonic behaviour of the crust? Establishing the Earth's Precambrian palaeo-magnetic history by correlating ages and pole positions for Precambrian rocks holds the exciting possibility of elucidating the tectonics of the Precambrian, and answering some at least of these fundamental questions.

Although a wide variety of rocks and minerals can be dated the significance of the calculated ages still requires careful assessment. The first minerals to be dated were commonly from pegmatites and it was a reasonable assumption that the calculated age was also the time of crystallization of the analysed mineral. Initially this assumption was also made when it became possible to date the common rock forming minerals and discordant results were generally attributed to 'later events and disturbances'. Krummenacher (1961) appears to have been one of the first to question this assumption. In an investigation of the K:Ar ages of micas from Himalayan metamorphites he pointed out that the K:Ar age of potassic mineral corresponded to the time when the mineral commenced to retain completely its radiogenic argon, and that if this was to date a metamorphism, then one must suppose that the host rock, after metamorphism, was displaced to a zone in the earth's crust where temperature and pressure were such that radiogenic argon could be completely retained. Krummenacher noted that micas from catazonal metamorphites in the Himalayas gave significantly younger ages than those from mesozonal and epizonal rocks although all rocks had been metamorphosed at essentially the same time. He suggested that the catazonal rocks had remained longer in the deeper and hotter parts of the crust at temperatures such that radiogenic argon was expelled from the host mineral as it was generated. From this, followed the interpretation that certain mineral ages might reflect the cooling history rather than the crystallization of the mineral concerned, and furthermore if it was assumed that the temperature at which argon was retained was always the same for a given mineral, it would be possible to calculate from discordant age patterns such as he had observed the rate of uplift and erosion; in a short footnote Krummenacher (op. cit.) reported such a calculation.

The approach pioneered by Krummenacher was more fully exploited by Jäger whose work was published in the following year (Jäger 1962). Jäger attempted to elucidate a complete orogenic history and viewed calculated ages within a dynamic framework of heating, deformation, uplift, erosion and cooling. The type of attack developed by Jäger has become the *modus operandi* of the geochronologist, to which has subsequently been added consideration of the relationship between the cooling history of rocks and their magnetic properties, and the radiogenic isotope geochemistry of rocks and minerals. The latter consideration sheding light on the history of the rocks before the last major event or orogeny by using the radiogenic isotopes as time dependent tracers.

Gerling (1942) was one of the first to use radiogenic isotopes as time dependent tracers and by taking the isotopic composition of leads from galenas of assumed ages arrived at an estimate of the age of the earth. Without knowing of Gerling's work, and quite independently, Holmes (1946) and Houtermans (1947) made similar calculations and all came to the conclusion that the earth was over 3000 Ma old (Holmes 1956). The particular problem of the age of the earth was finally resolved by Paterson (1956) and will not be further discussed here. Holmes, however, could see wider applications of the observed variations in the isotopic composition of the elements containing radiogenic isotopes and in 1956 posed the following question which he clearly considered to be resolvable by reference to such elements, 'Has continental material, or more particularly sialic material gradually accumulated during the earth's history, or did it reach the surface at an early stage and then begin to circulate through successive cycles of denudation, deposition, metamorphism, granitization, etc?' We are still attempting to find the answer to this question and with the advance of technology can now look at the isotopic variations in strontium and neodymium as well as lead. However, despite the advances resulting from our rapidly evolving technology it is humbling to recall that many of the topics briefly reviewed in this introduction were mooted by Holmes nearly fifty years ago, and, in one of the most foresighted papers in the geological literature (Holmes 1932), the paths we should follow to resolve the various problems were clearly mapped out.

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Precambrian geochronology and the geological record

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SUMMARY: Rb-Sr, Sm-Nd and Pb/Pb whole-rock isochron methods and the U-Pb zircon method have been widely applied to dating Precambrian igneous and meta-igneous rocks. Application of these methods in regional geochronological studies provides reliable constraints on the age and temporal evolution of rock units, even when some parent-daughter systems have suffered disturbance in metamorphic and/or metasomatic events which significantly post-date primary rock formation. Initial isotopic compositions of Sr, Nd and Pb aid geological interpretation of isochron age data, constrain crustal residence time of the protoliths and can be used to assess the relative contributions of mantle and crust in magma genesis.

Precambrian sediments can be indirectly dated by reference to age data on interbedded volcanics, or by age-bracketing between dated basement rocks and dated cross-cutting intrusives. Significant progress has also been made in the direct dating of sedimentary rocks. Published isotopic age determinations on early Precambrian sediments containing biogenic remains are briefly reviewed. Provided that stratigraphical and geochronological correlations are correct, it appears that stromatolitic limestones were being deposited by *c.*3400 – 3300 Ma ago, and that true microfossils occur in rocks dated at *c.*3200 – 3000 Ma. Evidence relating to possible biogenic markers in sediments reliably dated at *c.*3800 – 3700 Ma is not yet definitive.

Isotopic methods of age determination can now be applied to many common rocks and minerals in order to date geological events ranging from primary rock formation to the closure of different minerals to radiogenic isotope diffusion, which may relate to the time of uplift and cooling of a metamorphic terrain. Application of different decay schemes in regional geochronological studies has provided reliable constraints on the ages, geological history and petrogenesis of major rock units, even in cases where parent-daughter systems have suffered some disturbance during metamorphic and/or metasomatic events significantly post-dating primary igneous crystallization. Initial isotopic compositions of Sr, Pb and Nd have proved essential to the geological interpretation of isochron age data, and they also serve to constrain crustal residence time for the protoliths of a given rock unit. Such processes as melting, migmatization, metamorphism and metasomatism cannot totally eradicate the radiogenic isotope record in ancient, crustal rock units, notwithstanding the contrary view expressed by a few workers (see below).

The most frequently dated rocks in the Precambrian (*c.*3700 – *c.*570 Ma) are in the first place of igneous origin, and comprise basement gneisses (orthogneisses), major and minor plutons of diverse chemical compositions, and volcanic rocks within supracrustal associations, including greenstone belts. Problems of interpretation may arise in determining exactly what type of geological event is being dated, particularly in the case of major orthogneiss units. Here the question frequently arises whether measured whole-rock isochron dates (Rb-Sr, Sm-Nd, Pb/Pb) refer to the time of magmatism or of regional metamorphism, where these are resolvable. When the principal rock units from a high-grade terrain yield broadly concordant whole-rock isochron dates from different decay schemes, together with mantle-type initial Sr, Nd and Pb isotopic compositions, the time interval between (1) primary extraction of magmatic precursors of orthogneisses from a mantle-type source region, on the one hand, and (2) intracrustal magmatic, geochemical and metamorphic differentiation to produce stabilized continental crust and late granitic plutonism, on the other hand, may take place within a comparatively short time span not usually exceeding 200–300 Ma, and sometimes much less (Moorbath 1975a, 1975b, 1977; Wells 1981). Individual cratons yield convincing age and isotopic evidence for the episodic occurrence of such 'crustal accretion-differentiation superevents' (CADS). Thus, De Laeter *et al.* (1981) write, 'Probably the most striking geochronological characteristic of either of the Western Australian Archaean blocks is the vastness of the area of the Yilgarn Block over which the isotope evidence points to a penecontemporaneous generation of sial over a relatively short interval' With the accumulation of more geochronological evidence from different continents, it is becoming evident that CADS are not contemporaneous on a global scale. For example, the two earliest CADS in different sectors of the North Atlantic craton at *c.*3700 – 3500 and *c.*3000 – 2600 Ma are not matched by what could be the earliest CADS on the Indian sub-continent at *c.*3350 – 3000 Ma (Beckinsale *et al.* 1980; Moorbath & Taylor 1981). Nevertheless, most workers agree that the late Archaean (*c.*3000 – 2500 Ma) witnessed the most rapid and voluminous production of mantle-derived, juvenile continental crust during earth history. Estimates of the relative mass of the present-day continental crust in existence by *c.*2500 Ma ago vary between *c.*50 and 85% (Veizer 1976; Jacobsen & Wasserburg 1979; O'Nions *et al.* 1979a; De Paolo 1980; Dewey & Windley 1981, McLennan & Taylor 1982).

Whilst debate continues on the genetic relationships, if any, between Precambrian and Phanerozoic tectonic environments, it is clear that major tectonic activity is at present largely confined to constructive and destructive plate margins, leaving large areas of continental crust tectonically undisturbed, although the loci of tectonic activity may well change in future to reactivate currently stable shield rocks. In Phanerozoic terrains, it is relatively straightforward to resolve detailed geological and geochronological relationships in the vicinity of destructive plate margins, where juvenile calc-alkaline crust has been, and is still being, added to older continental crust (e.g. Circum-Pacific belt). Such juvenile crust is generated by partial melting processes within upper mantle and/or subducted oceanic crust, and by subsequent magmatic fractionation processes. Even allowing for the effects of mantle heterogeneity, involvement of sea-water-altered oceanic crust, crustal contamination etc., the radiogenic isotopic relationships in magmatic rocks at sites of continental growth tend to be far less complex than in collisional orogenic belts where magma genesis by crustal melting may greatly predominate over mantle-derived magmatism and where rock units of widely different ages have suffered a complex metamorphic history. In Precambrian shield areas, sites of exceptional geochemical and isotopic complexity occur in areas where two or more CADS are superposed. Such areas contrast sharply with the simpler age and isotope relationships observed over the greater part of many ancient cratons formed during a single, major CADS. New interpretative methods developed by several workers to deal with areas of cratonic overlap have provided a means of characterizing the separate events, and of determining the extent of crustal interaction between the rocks formed during each event, particularly as regards magma genesis.

It is, in principle, possible to determine the timing and magmatic evolution in a single CADS from the commencement of mantle-derived magmatism which produced the precursors of calc-alkaline tonalitic-to-granodioritic orthogneisses, through to the terminal magmatic episode, frequently characterized by normal granites whose isotopic characteristics (both radiogenic and stable) suggest derivation by crustal anatexis following prolonged intracrustal differentiation. In line with the thermal and structural models for crustal accretion and thickening of Wells (1979, 1981), the progressive change from mantle-derived to crust-derived magmatism during a CADS is entirely to be expected.

In the following somewhat selective review of recent (until 1982) work on the geochronology of Precambrian rocks, particular attention is given to the question of concordance and discordance of measured ages, as well as to the geological interpretation of ages and initial isotopic compositions. Later on, recent (until 1982) work on the relatively neglected topic of isotopic dating of Precambrian sedimentary and metasedimentary rocks is reviewed, a subject of particular relevance to stratigraphers and palaeontologists, as well as organic geochemists and 'molecular palaeontologists' concerned with the origin and nature of the earliest life forms. Some early Precambrian sediments are more richly endowed with biogenic markers, and primitive fossils, than was supposed until very recently.

No descriptions are given here of isotopic age methods, or associated methodology. Rb-Sr, U-Pb, Pb/Pb, K-Ar methods are adequately described by Faure (1977), whilst the Sm-Nd method is treated by O'Nions *et al.* (1979b) and De Paolo (1981a), and the recently developed Lu-Hf method by Patchett & Tatsumoto (1980). All ages quoted in this paper have been calculated, or recalculated, with the following decay constants:
$\lambda^{238}U = 1.55125 \times 10^{-10}a^{-1}$, $\lambda^{235}U = 9.8485 \times 10^{-10}a^{-1}$, $\lambda^{87}Rb = 1.42 \times 10^{-11}a^{-1}$, $\lambda^{147}Sm = 6.54 \times 10^{-12}a^{-1}$, $\lambda^{176}Lu = 1.96 \times 10^{-11}a^{-1}$.
Wherever possible, errors on quoted ages are given at the 2σ level.

Precision of Precambrian age measurements

For the majority of published Rb-Sr and Pb/Pb whole-rock isochrons on Precambrian orthogneiss units, the scatter of data points about the computed best-fit regression line exceeds the analytical error of the individual measurements. The quoted errors on the age are therefore customarily enhanced to take into account this 'geological scatter', which may reflect: (1) inhomogeneity of initial Sr or Pb isotopic compositions; (2) localized, partial Sr or Pb isotopic rehomogenization at one or more later times; (3) open-system behaviour resulting in loss or gain of Rb, Sr, U or Pb at the scale of the analysed whole-rock samples.

The statistical analysis and interpretation of imperfectly-fitted Rb-Sr whole-rock isochrons (or 'errorchrons') from polymetamorphic terrains has been treated by Cameron *et al.* (1981) in the light of the various geological possibilities listed above. Allowing for the geological and analytical uncertainties, many Rb-Sr and Pb/Pb whole-rock isochron ages are quoted with errors of ± 1.5 to 5%, or ± 45 to ± 150 Ma for 3000 Ma-old rocks. Thus the overall error on an Archaean age result may cover a time-span which is comparable with the duration of an entire CADS. Even so, the age and isotope data obtained from an errorchron may be perfectly adequate for constraining the timing of a CADS, and the crustal residence time of the protoliths (Moorbath 1975a, 1975b).

Well-fitted, or near-perfect, Rb-Sr and Pb/Pb whole-rock isochrons are commonly obtained in Precambrian (and Phanerozoic) terrains on post-tectonic granitic intrusives which have suffered little or no subsequent tectonism with consequent isotopic disturbance. In such cases, the error on the age is frequently minimized by a large spread of Rb/Sr and U/Pb ratios in individual intrusives. Two examples, recently measured in the authors' laboratory, are (1) the Qôrqut granite of West Greenland, with a Rb-Sr age of 2530 ± 30 Ma (Moorbath *et al.* 1981) and (2) the Chitradurga granite of southern India, with a Pb/Pb age of 2605 ± 18 Ma (Oxford unpublished data). Other examples of this type are frequently reported in the literature.

Concerning the choice of appropriate dating methods in Precambrian terrains, we note the following three points:

1. Very few Sm-Nd whole-rock isochron ages have so far been reported for orthogneisses. Only by selecting orthogneiss samples covering a very broad range of bulk compositions is it possible to obtain an adequate spread of Sm/Nd ratios to define a precise isochron age result. Thus rock types of questionable consanguinity might be wrongly associated on the same isochron diagram. The precision of Sm-Nd whole-rock isochron ages on orthogneiss units appears to be comparable with that obtained from the other whole-rock isochron dating methods. Some recent interpretations of Sm-Nd whole-rock and mineral data suggest that open-system behaviour during subsequent metamorphism and/or slow cooling could be more common than was originally surmised (e.g. De Paolo *et al.* 1982; Humphries & Cliff 1982). The Sm-Nd whole-rock method, however, seems particularly well suited to mafic volcanic suites from Precambrian greenstone belt assemblages, as will be described later. The spread of Sm/Nd ratios in such rocks is adequate for precise age determinations, whilst the rare-earth elements (REE) have been shown to be less susceptible to post-depositional chemical mobility than Rb, Sr, U and Pb (Hamilton *et al.* 1977; O'Nions *et al.* 1979b);
2. The two decay schemes $^{235}U \rightarrow {}^{207}Pb$ and $^{238}U \rightarrow {}^{206}Pb$ are usually combined in the form of plots of $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ (for treatment of Pb isotope systematics, see Faure 1977). The reason for this is not only to obviate the necessity for measurement of U and Pb concentrations, but

also because there is good evidence for geologically recent loss of U in many surface outcrops of common silicate rocks, presumably through the medium of percolating ground waters. *Recent* U-loss will seriously affect U-Pb systematics, as is evident from conventional isochron plots of $^{238}U/^{204}Pb$ versus $^{206}Pb/^{204}Pb$, and $^{235}U/^{204}Pb$ versus $^{207}Pb/^{204}Pb$, but will have no detectable influence on Pb/Pb systematics (e.g. Rosholt *et al.* 1973; Moorbath *et al.* 1975);

3. The highly promising Lu-Hf method has begun to be applied to ancient rocks with very promising results (Patchett & Tatsumoto 1980; Patchett *et al.* 1981; Pettingill & Patchett 1981).

For recent advances in high-precision dating of Precambrian rocks, we must look to the U-Pb dating of accessory zircons. In favourable cases, Archaean rocks can be dated to a precision of one or two million years. This is particularly applicable to the dating of volcanic rocks in low-grade metasupracrustal assemblages. Thus Nunes & Thurston (1980) reported zircon U-Pb data from three metavolcanic units in the Uchi-Confederation Lake greenstone belt of north-western Ontario which gave ages of 2958.6 ± 1.7 Ma, ~ 2794 Ma, and 2738 $\pm^{5}_{2}$ Ma. An intrusive monzonite pluton within the greenstone belt also gave a very precise age of 2729.6 ± 1.3 Ma. The original paper should be referred to for details of the geological interpretation. More recently, Krogh & Turek (1982) have reported a zircon U-Pb date of 2713.2 $\pm^{2.6}_{1.9}$ Ma from a volcanic unit in the Gamitagama greenstone belt of the south Superior Province of Ontario, with two post-orogenic intrusions in the same belt dated separately at 2667.8 $\pm^{3}_{2}$ Ma and 2668.3 $\pm^{3.3}_{2.2}$ Ma. Similar precise zircon U-Pb ages have been reported from the Wabigoon Subprovince of north-west Ontario by Davis *et al.* (1982). Yet another example is provided by the Sudbury Nickel Irruptive of Ontario, which yields two independent zircon U-Pb ages of 1849.6 $\pm^{3.4}_{3.0}$ Ma and 1849.4 $\pm^{1.9}_{1.8}$ Ma, contrasting sharply with previous whole-rock Rb-Sr determinations which gave ages between 1700 and 2000 Ma, with individual errors of about ± 100 Ma (Krogh *et al.* 1982). Slightly less precise, but still quite impressive, is a zircon U-Pb age of 3769 $\pm^{11}_{8}$ Ma reported by Michard-Vitrac *et al.* (1977) on a metarhyolite in the oldest known volcano-sedimentary sequence on earth, namely the Isua supracrustal succession of West Greenland.

Whether or not zircons extracted from deep-seated plutonic rocks, including basement gneisses, could yield equally precise U-Pb ages remains to be seen. It is also questionable just how useful such precise ages would be, since it is not yet known exactly what stage or event in an extended magmatic and/or metamorphic evolutionary history a measured zircon age would actually represent.

Concordant and discordant dates

The majority of published Precambrian whole-rock ages have been determined by the Rb-Sr method. However, in recent years, the Pb/Pb method has been more widely used, enabling a comparison to be made between the two methods. In many cases, there is agreement within the error of the measured isochron or errorchron ages. In addition, comparatively precise zircon U-Pb ages may also be available for the same rocks. Several age-concordant rock units are summarized in the first part of Table 1. Two of the oldest known rock units on earth, namely the Isua supracrustals and the succeeding Amîtsoq gneisses of West Greenland, each yield satisfying agreement of whole-rock and zircon ages by the different age methods (Tables 2 and 3), although Rb-Sr and K-Ar measurements on separated minerals from both rock units exhibit complex age patterns resulting from late Archaean (*c.*2900 – 2600 Ma) and mid-Proterozoic (*c.*1800 – 1600 Ma) thermal events which caused open-system behaviour of micas and amphiboles to diffusion of radiogenic isotopes (Pankhurst *et al.* 1973; Baadsgaard *et al.* 1976). This late open-system behaviour is also reflected in the scatter of whole-rock data points about best-fit regression lines significantly exceeding analytical error, leading to errorchrons rather than isochrons (Cameron *et al.* 1981)

The close agreement between the respective ages for the Isua supracrustals and Amîtsoq gneisses represents the earliest known example of the characteristically tight grouping of measured ages from orthogneiss terrains and associated volcano-sedimentary associations of greenstone belt type, such as is more frequently reported from *late* Archaean shield areas on several continents (Condie 1981). Such gneisses and associated stratigraphically older or younger supracrustals thus form part of the same major CADS, extending over a total period of about 100–300 Ma, although in a few cases supracrustal rocks are known to overlie *much* older basement gneisses formed in a previous CADS, as for example in parts of Zimbabwe (Wilson *et al.* 1978). The geological nature of the basement to a particular greenstone belt assemblage may be controversial, but there is little doubt that greenstone belts may be deposited on crust of either continental or non-continental affinity (Lowe 1982). Initial Sr, Nd and Pb isotopic compositions of the oldest West Greenland rocks, and of many later associations of gneisses and supracrustals in different shield areas, limit crustal residence to only about 100–200 Ma prior to measured isochron ages. In contrast, where significant volumes of ancient continental crust became involved in much later juvenile crust formations, as for example in the God-

TABLE 1. Some concordant and discordant whole-rock Rb-Sr and Pb/Pb isochron ages

Rock type and locality	Rb-Sr Ma	Pb/Pb Ma	Reference
	Concordant		
Orthogneiss, Fiskenaesset, W. Greenland	2840 ± 70	2820 ± 70	Taylor *et al.* (1980)
Qôrqut granite, W. Greenland	2530 ± 30	2580 ± 80	Moorbath & Taylor (1981)
Orthogneiss, Gwenoro, Zimbabwe	2705 ± 80	2675 ± 40	Hawkesworth *et al.* (1975); Oxford unpubl. data.
	Discordant		
Orthogneiss, Sermilik, W. Greenland	2750 ± 40	3000 ± 90	Taylor *et al.* (1980)
Orthogneiss, Fadugu, Sierra Leone	2750 ± 60	2960 ± 50	Beckinsale *et al.* (1980)
Orthogneiss, Rhodesdale, Zimbabwe	2745 ± 100	2980 ± 130	Moorbath *et al.* (1977b); Oxford unpubl. data.

TABLE 2. Summary of ages for Isua supracrustals, West Greenland.

Rock-type	Method	Age, Ma	Reference
Banded iron-formation	WR, Pb/Pb	3710 ± 60	Moorbath *et al.* (1973)
Metarhyolite	WR, Pb/Pb	3670 ± 100	Oxford unpubl. data
Metapelite	WR, Pb/Pb	3700 ± 100	Oxford unpubl. data
Galena	Pb/Pb model age	~ 3740	Appel *et al.* (1978)
Quartzite	Zircon, U-Pb	3670	Baadsgaard (1976)
Metarhyolite	Zircon, U-Pb	3770; 3769^{+11}_{-8}	Baadsgaard (1976); Michard-Vitrac *et al.* (1978)
Metarhyolite conglomerate & matrix, metapelite	WR, Rb-Sr	3650 ± 60	Moorbath *et al.* (1977a)
Metarhyolite	WR, Rb-Sr	3690 ± 50	Oxford unpubl. data
Metarhyolite, conglomerate, garbenschiefer	WR, Sm-Nd	3770 ± 40	Hamilton *et al.* (1978)

TABLE 3. Summary of ages for Amîtsoq gneiss, West Greenland

Metamorphic Facies	Method	Age, Ma	Reference
Ampbibolite	WR, Rb-Sr	3640 ± 50 (mean)	Moorbath *et al.* (1972; 1975; 1977a)
Granulite	WR, Rb-Sr	3560 ± 140	Griffin *et al.* (1980)
Amphibolite	WR, Pb/Pb	3700 ± 50 (mean)	Moorbath *et al.* (1975); Baadsgaard *et al.* (1976); Oxford unpubl. data
Granulite	WR, Pb/Pb	3625 ± 130	Griffin *et al.* (1980)
Amphibolite	Zircon, U-Pb	3600 ± 50	Baardsgaard (1973); Baardsgaard *et al.* (1976).
Amphibolite	WR & Zircon, Lu-Hf	3550 ± 220	Pettingill & Patchett (1981).

thaabsfjord region of West Greenland (Taylor *et al.* 1980), the initial isotopic ratios unambiguously record the presence and participation of the older crust.

We now consider cases where there is a significant age discrepancy between different decay schemes when applied to a given rock unit, even to the identical suite of samples. In the majority of cases so far reported, the whole-rock Rb-Sr isochron or errorchron age is significantly lower than the whole-rock Pb/Pb isochron or errorchron age by as much as several hundreds of millions of years. Several examples are quoted in the second part of Table 1. Where zircon U-Pb analyses are also available, the relatively precise ages tend to agree more closely with the Pb/Pb age, suggesting that this is closer to the age of primary rock formation. Possible explanations for the age discordance are that (1) whole-rock Rb-Sr systematics are more easily reset by later metamorphic and/or metasomatic events, or, (2) during slow cooling in the course of a single CADS, large-scale homogeneity of Sr isotopes can be maintained by diffusion in a rock unit over a period represented by the age discordance, or, closely related to this, (3) diffusive loss of radiogenic Sr from whole-rock samples can proceed to lower temperatures than diffusive loss of radiogenic Pb, or, (4) combinations of (1) to (3).

We favour possibility (3) for those cases where the age discordance is not more than about 200–300 Ma, and we invoke the well-established concept of different minerals in a gneiss complex reaching their characteristic blocking temperatures for radiogenic isotope diffusion at different times during very slow cooling. Calc-alkaline orthogneisses typically contain plagioclase and biotite, plus variable amounts of potash feldspar. Of these minerals, biotite usually has the highest Rb/Sr ratio, and may contain a significant proportion of the sites in which radiogenic ^{87}Sr is generated within a rock. Furthermore, biotite probably has a lower blocking temperature for Sr diffusion than plagioclase or potash feldspar, although there is not much relevant data on this particular problem (Dodson 1979). This hypothesis is based solely on observed, discordant Rb-Sr mineral age patterns in feldspars and micas in thermally overprinted granitoid rocks and pegmatites (e.g. Giletti *et al.* 1961; Francis *et al.* 1971). During slow cooling of a gneiss complex, feldspars may thus become closed systems to Sr diffusion long before biotite, and during this interval feldspars cannot act as acceptors for Sr which is diffusing out of biotites. This Sr may therefore be lost from hand-specimen sized domains, and may even leave the rock unit altogether, effectively assisted by the transporting action of intergranular fluids. Throughout this cooling history, the feldspars and the U-rich accessory minerals, which are the respective principal repositories of common and radiogenic Pb, presumably remain closed to diffusion of Pb. As a result, whole-rock Pb/Pb isochron/errorchron ages may significantly exceed whole-rock Rb-Sr isochron/errorchron ages, but may still fall within the overall age range of a single, major CADS. On this hypothesis, *concordance* of whole-rock Pb/Pb and Rb-Sr ages for a rock unit might signify a cooling interval not exceeding the errors of the age measurements.

The situation outlined above probably differs fundamentally from those well-documented cases where complex metamorphic and/or metasomatic events, recognizable from field observations, and significantly post-dating the primary rock formation age, have disturbed whole-rock Rb-Sr systematics to an extent which sometimes approaches total resetting with associated Sr-isotope homogenization, but more often yields errorchrons with intermediate 'age' values of no particular geological significance, or simply a meaningless scatter of data points which fall somewhere between limiting reference lines for the ages of primary rock formation and subsequent non-isochemical reconstitution. Precambrian examples of this type have been reported from southern Norway by Field & Råheim (1979, 1980, 1981), from eastern Australia by Black *et al.* (1979), and from western Australia by Nieuwland & Compston (1981). In the last case it is clear that post-emplacement metamorphism at *c.*2500 Ma produced variable loss of ^{87}Sr from whole-rock samples of 3200 Ma-old orthogneisses.

To demonstrate the perversity of nature, however, we note that in other cases regional metamorphism and associated deformation have *not* reset or significantly disturbed whole-rock Rb-Sr systematics (e.g. Jacobsen & Heier 1978; Welin & Kähr 1980). The work of Skjernaa & Pedersen (1982) on several Proterozoic granitoid complexes of south-eastern Norway clearly demonstrates that rock units with different whole-rock Rb-Sr ages have been subjected to common deformational and metamorphic (up to amphibolite facies) episodes. In addition, rocks from one continuous complex gave the same whole-rock Rb-Sr ages both north and south of a major metamorphic and deformational front. Skjernaa & Pedersen (1982) echo the view of Moorbath (1975b) when they state that 'well-defined whole-rock Rb-Sr isochrons from metaplutonic rocks should not, without the greatest care, be interpreted as defining metamorphic ages (resetting ages)' (see also Field & Råheim, op.cit.).

As pointed out earlier, the time interval between initial generation of sialic material from the mantle and eventual stabilization of continental crust, during which important geochemical changes occur (e.g. marked depletion of the heat-producing elements K, Rb, U, Th and other large-ion lithophile trace elements in granulite facies rocks), can often be broadly constrained by Rb-Sr and U-Pb (and Pb/Pb) techniques. Recently, Hamilton *et al.* (1979a) have applied the Sm-Nd method to the Lewisian gneiss complex of north-west Scotland and propose that the application of all three techniques may produce more satisfactory resolution of the timing and duration of a given crust-forming event. They report a whole-rock Sm-Nd isochron age of 2920 ± 50 Ma for a suite of eleven granulite-and amphibolite-facies gneiss samples from two groups of localities respectively in the Scourian and Laxfordian sectors of the Lewisian complex. This age is some 250 Ma (actually 240 ± 90 Ma) older than the mean of a selection of whole-rock Rb-Sr, Pb/Pb and zircon U-Pb ages from other parts of the Lewisian complex, including some areas which suffered quite severe Laxfordian metamorphism and migmatization in mid-Proterozoic (*c.*1900 – 1700 Ma) times. Hamilton *et al.* (1979a) conclude from their Sm-Nd isotope data that the igneous calc-alkaline precursors separated from previously undifferentiated mantle at 2920 ± 50 Ma ago. Since U and Pb, as well as Rb and Sr, are respectively much more easily fractionated by crustal processes than Sm and Nd, Hamilton *et al.* (1979a) interpret published Rb-Sr, U-Pb and Pb/Pb ages in the range *c.*2850 – 2600 Ma as representing the times of final differentiation and stabilization of the Lewisian complex as granulite- and amphibolite-facies crust. Whilst the overall interpretation of Hamilton *et al.* (1979a) may be perfectly valid — in which case it is of the greatest signifcance — it could be criticized on the grounds that Sm-Nd analyses were not carried out on the same samples or individual rock units as those dated by the other decay schemes. Without the evidence of discrepancy between Sm-Nd, Rb-Sr and Pb/Pb dates on the *same* samples it is not possible to resolve whether the differentiation/stabilization time of *c.*250 Ma proposed by Hamilton *et al.* (1979a) actually applies to the Lewisian Complex as a whole, or whether extraction of the igneous precursors of the Lewisian orthogneisses from a mantle-type source region extended over a total period of *c.*250 Ma, with differentiation/stabilization occurring geochronologically 'instantaneously' in each sector. Such fundamental problems can only be solved by the combined application of all relevant decay schemes to several rock suites from a given basement complex.

An interesting example of apparent major age discordance between whole-rock Rb-Sr, Sm-Nd and Pb/Pb isochron ages is provided by ancient gneisses occurring as remnants within the Singhbhum granite batholith of eastern India. The published Rb-Sr age (Sarkar *et al.* 1979) is approximately 3100 Ma, whilst the present authors have obtained an Rb-Sr age of 3280 ± 260 Ma, with an initial $^{87}Sr/^{86}Sr$ ratio of 0.701 ± 0.001, and a Pb/Pb age of 3317 ± 25 Ma, on samples of gneisses kindly supplied by S. N. Sarkar and A. K. Saha. The high error on the Rb-Sr age results from the restricted spread of Rb/Sr ratios in the suite of samples. Basu *et al.* (1981) have reported a whole-rock Sm-Nd isochron age of 3775 ± 89 Ma for these and closely related samples, with an initial $^{143}Nd/^{144}Nd$ ratio of 0.50798 ± 0.00007, which is significantly higher (ε_{Nd} = +3.3 ± 0.9 at 3775 Ma) than the value for a mantle source with a chondritic evolution curve of $^{143}Nd/^{144}Nd$ (based on several mantle-derived rock units reported by O'Nions *et al.* 1979b). Basu *et al.* (1981) interpret the measured Sm-Nd date as the true age of the eastern India gneisses, and conclude from the initial $^{143}Nd/^{144}Nd$ ratio that parts of the earth's mantle were already differentiated with respect to the chondritic Sm/Nd ratio at about 3800 Ma ago.

However, there are serious problems with the interpretation by Basu *et al.* (1981) of their Sm-Nd array as a true isochron. The regression line is based on two widely separated clusters of points (the upper cluster only comprising two points). Calculated Sm-Nd model ages for the low and high cluster of points, using the Sm-Nd model parameters of Jacobsen & Wasserburg (1980), group closely around 3400 and 3100 Ma respectively. Thus the two clusters probably do not belong to a homogeneous, cogenetic sample suite, and the regression line which joins the two clusters is more likely a spurious tie-line which yields a fictitious age and initial $^{143}Nd/^{144}Nd$ ratio. The measured whole-rock Pb/Pb date of 3317 ± 25 Ma is probably close to the true age of these gneisses. The Pb/Pb age agrees well with a whole-rock Rb-Sr age of 3358 ± 66 Ma (initial $^{87}Sr/^{86}Sr$ = 0.7000 ± 0.0004) for a suite of samples from the Peninsular Gneisses of the South India craton (Beckinsale *et al.* 1980). This may represent the age of a major, and perhaps the earliest, CADS to form the nucleus of the Indian subcontinent.

Problems in the interpretation of Pb/Pb linear arrays

Many major Precambrian orthogneiss terrains yield whole-rock Pb/Pb isochrons or errorchrons dating some stage (probably an early stage) of a CADS. Frequently, Pb/Pb isotope systematics indicate derivation of the magmatic precursors from source regions with a very restricted overall range of $^{238}U/^{204}Pb$ ratios, approximating to single-stage Pb isotopic evolution of the mantle from the time of formation of the earth to the measured isochron age (Moorbath & Taylor 1981). Calculated $^{238}U/^{204}Pb$ (μ_1) values for the source of the parent magmas of many Archaean orthogneisses fall in the approximate range 7.5–8.0 (Moorbath & Taylor 1981), using the Pb isotopic model parameters of Oversby (1974). Orthogneisses with parental μ_1 values within this range are regarded as having been derived from magmatic precursors originating in the upper mantle.

The Pb/Pb system is extremely sensitive to a variety of crustal processes (see Moorbath & Taylor 1981) which makes it a particularly useful tracer for some aspects of continental evolution. For example, granites derived by partial melting of ancient, U-depleted, sialic crust may yield whole-rock Pb/Pb isochrons with anomalously low, apparent μ_1-values, showing that a conventional single-stage Pb isotopic evolutionary model is inappropriate. Thus the Qôrqut granite complex of West Greenland yields a whole-rock Pb/Pb isochron age of 2580 ± 80 Ma, in close agreement with a whole-rock Rb-Sr isochron age of 2530 ± 30 Ma (initial $^{87}Sr/^{86}Sr$ = 0.7081 ± 0.0008), but with an apparent μ_1-value of 6.2. These data have been interpreted as indicating derivation of Qôrqut granite magma by partial melting of a mixture of approximately equal proportions of *c*.3650 Ma-old Amitsoq gneisses and *c*.2900 Ma-old Nûk gneisses (Moorbath *et al.* 1981). This is strongly supported by Sm-Nd model ages in the range 3300–3100 Ma for the Qôrqut granite (Oxford unpublished data).

The early Archaean Amîtsoq gneisses and most late Archaean othogneisses in West Greenland have relatively simple Pb/Pb isotope systematics, indicating derivation of their respective magmatic precursors from mantle type source regions not long before the measured whole-rock Pb/Pb (and Rb-Sr) isochron ages. However, in the area of spatial overlap between the two terrains in the Godthaabsfjord region, the Pb/Pb isotope systematics in the younger Nûk gneisses are complex, because the *c*.2900 Ma-old Nûk magmas became variably contaminated with much older, very unradiogenic, crustal Amîtsoq-gneiss-type Pb. However, the resulting mixing relationships enable the proportions of Amîtsoq-type Pb and primary Nûk-type Pb within any given Nûk gneiss sample to be determined (Taylor *et al.* 1980). This type of Pb isotope evidence for crust-magma interaction provides a powerful tool for the detection, subsurface 'mapping' as well as geochronological and geochemical characterization of deep, ancient continental crust.

A classic example of a linear mixing relationship between Pb isotopes from rock units of very different ages is provided by Lower Tertiary (*c*.55–60 Ma) volcanic and intrusive basic and acid rocks of the Isle of Skye, north-west Scotland. This area is known to be underlain by the late Archaean Lewisian Gneiss complex, which outcrops on the nearby Scottish mainland. Pb isotope analyses for Skye igneous rocks form a linear array on a $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram with an errorchron age of 2920 ± 70 Ma (Dickin 1981), actually interpreted as a mixing line between *c*.2900 Ma-old crustal Pb and *c*.60 Ma-old mantle-derived Pb (Moorbath & Welke 1969; Dickin 1981). Interpretation of the Pb/Pb linear array as a conventional Pb/Pb isochron would in this case lead to the conclusion that the Skye igneous rocks were *c*.2900 Ma old, which is totally contravened by all geological and conventional geochronological evidence! The fact that the Lower Tertiary igneous rocks may contain anywhere between 0 and *c*.90% of Lewisian-type Pb has given rise to many fundamental speculations about the mechanisms of crust-magma interaction, and its effect on the geochemistry and petrogenesis of igneous rocks in different tectonic environments (Dickin 1981; Thompson *et al.* 1982).

A further situation in which whole-rock Pb/Pb isotopic age data must be treated with great caution is now discussed. Spurious ages may be obtained when high-grade metamorphism affects a rock unit long after its primary formation, i.e. when crustal accretion and *final* metamorphic/geochemical differentiation and stabilization do *not* form part of the same CADS. Taylor (1975) published a whole-rock Pb/Pb errorchron date of 3410 ± 70 Ma for the migmatitic Vikan gneisses of Langøy, Vesterålen, North Norway and concluded that this was the age of a primary rock-forming event. Further data from the same and other decay schemes (Griffin *et al.* 1978; Jacobsen & Wasserburg 1978; Taylor,

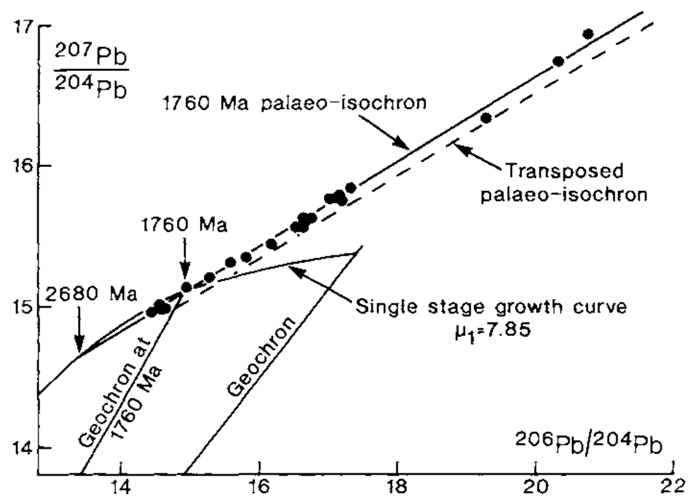


FIG. 1. Pb isotopic evolution in the Vikan migmatitic gneisses, North Norway. The plotted data points have been corrected to 1760 Ma ago and thus define the 1760 Ma palaeo-isochron, the slope of which corresponds to an age of 2680 Ma. Assuming two stages for Pb isotopic evolution up to 1760 Ma, a μ_1 value of 7.85 is calculated for the first stage. The broken line represents the actual array of present-day isotopic compositions of the Vikan gneisses (Taylor 1975). The slope of this line is indistinguishable from the slope of the 1760 Ma palaeo-isochron. The broken line is a 'transposed palaeo-isochron', resulting from small, fairly uniform increments in $^{206}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ ratios in all samples since the establishment of very low and fairly uniform $^{238}U/^{204}Pb$ ratios by severe U depletion during high-grade metamorphism at 1760 Ma ago (Griffin *et al.* 1978; Jacobsen & Wasserburg 1978; Taylor, unpublished data).

unpublished data) led to a major re-interpretation of the Pb isotopic data. The high μ_1-value of 8.9 implied for the source of the Vikan gneisses on the basis of a single-stage model (compared with typical values of 7.5–8.0), as well as the lack of any very unradiogenic Pb isotopic compositions, were very unusual features compared with other Precambrian high-grade gneiss terrains with available Pb/Pb data. The present-day range of Pb isotopic compositions could not have resulted from a single stage of Pb isotopic evolution from a uniform initial Pb isotopic composition with the present small range of low $^{238}U/^{204}Pb$ ratios. In fact, the slope age of 3410 ± 70 Ma reported by Taylor (1975) for the Vikan gneisses is now regarded as spurious. It results from formation of the gneiss protoliths from an upper mantle source region at *c*.2680 Ma ago, followed by granulite facies metamorphism at *c*.1760 Ma ago which caused severe U-depletion and consequent disturbance of U-Pb whole-rock systems. In other words, relatively normal crustal U/Pb ratios between *c*.2680 and *c*.1760 Ma ago were reduced to such an extent at *c*.1760 Ma ago that radiogenic Pb isotopic evolution almost — but not quite — ceased within the granulite gneiss. This is illustrated in Fig. 1. The present-day Pb-Pb array for the Vikan gneisses is not a true secondary isochron originating via U/Pb fractionation from a single-stage source region, but a 'transposed palaeoisochron' (TPI), in which the present-day Pb isotope data fall on a line slightly transposed from the isochron for Pb isotopic compositions of 2680 Ma-old rocks which developed undisturbed until 1760 Ma ago. Interpretation of the line joining points 2680 Ma and 1760 Ma on the single-stage growth curve gives an anomalously old age if a conventional two-stage evolution model (i.e. single-stage mantle derivation followed by simple, one-stage crustal evolution) is assumed. The transposition of the 'palaeoisochron' developed between 2680 and 1760 Ma results from radiogenic development of Pb in a uniformly low U/Pb environment from 1760 Ma to the present.

Cases of the Vikan-type may be quite common. The Hebron gneisses of Labrador have yielded a whole-rock Rb-Sr isochron age of *c*.3500 Ma (Barton 1975), similar to the Uivak gneisses of Saglek Bay, Labrador (Hurst *et al.* 1975). Zircon U-Pb and whole-rock Rb-Sr data for this region indicate metamorphism and crustal reworking at 2800–2500 Ma (Baadsgaard *et al.* 1978; Collerson *et al.* 1982). Whole-rock Pb/Pb data on Hebron gneiss (P. N. Taylor, unpublished data) yield a spurious apparent 'age' of *c*.4450 Ma, almost equal to the accepted age of the earth, and an apparent μ_1 value of 32 (Fig. 2). Clearly, this could be subject to gross misinterpretation. In the absence of U and Pb concentration data, this case cannot be positively identified as a TPI, but the most plausible interpretation is that the Hebron Pb/Pb linear array results from late Archaean (*c*.2500 Ma) U-depletion of early Archaean (*c*.3600 Ma) protoliths.

Rb-Sr isotope systematics may be similarly affected, so that in any sector of continental crust which becomes severely depleted in Rb relative to Sr during a much later granulite-facies metamorphism, growth of radiogenic ^{87}Sr virtually ceases. This can lead to linear arrays on Rb-Sr isochron plots which yield anomalously old apparent ages. The Rb-Sr errorchron of 2240 ± 150 Ma (initial $^{87}Sr/^{86}Sr$ = 0.7126 ± 0.0011) reported by Taylor (1975) for the Vikan gneisses is now interpreted as a product of Rb-depletion during granulite-facies metamorphism at *c*.1760 Ma ago (see above).

Thus the Vikan gneisses, and similar cases, provide clear examples of pro-grade metamorphism and geochemical activation involving U and Pb (and other elemental) depletion of much older continental crust and, as such, are isotopically as clearly characterized as the contrasted case of crustal accretion with penecontemporaneous geochemical differentiation during a single CADS. This is further demonstrated by Sm-Nd, Rb-Sr and U-Th-Pb isotopic studies of granulite-facies rocks from Enderby Land, Antarctica (De Paolo *et al.* 1982). Granulite-facies metamorphism *c*.2500 Ma ago produced mobility of Rb, U, Sm and Nd, together with severe depletion in Rb and U, in crustal rocks which yield a wide range of Sm-Nd, U-Pb and Rb-Sr model ages suggesting primary crust formation at *c*.3500 Ma. In this case the Sm-Nd system preserves the most convincing evidence for the age of the primary crust, although the authors are careful to point

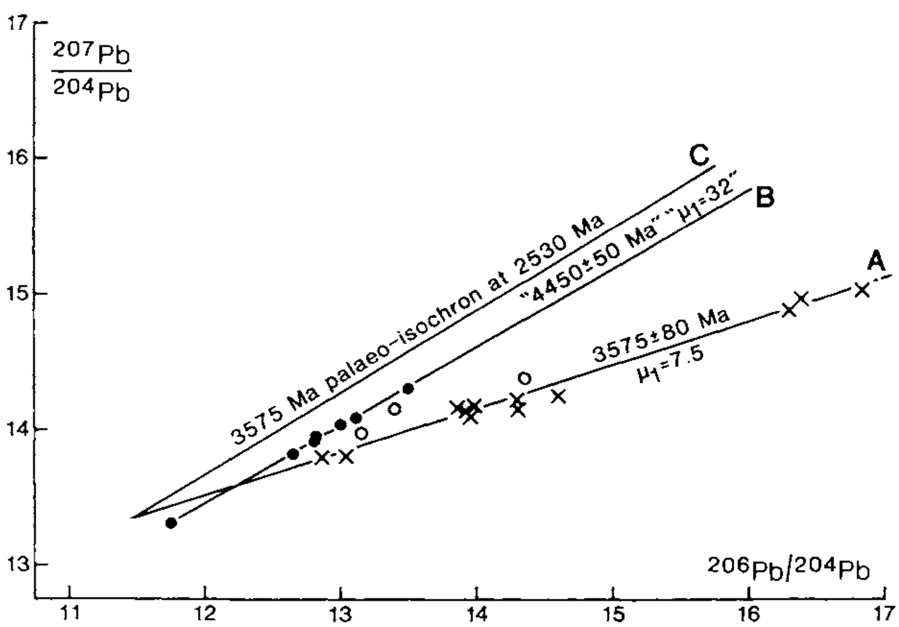


FIG. 2. Plot of $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ for whole-rock samples of the Hebron gneiss, Labrador. (Samples were kindly provided by K. D. Collerson).

Seven out of ten samples define a well-fitted but spurious 'isochron' (filled circles, line B) which yields an apparent age of 4450 ± 50 Ma and an apparent μ_1 value of 32. The remaining three samples (open circles) do not fall on this line. The 3575 ± 80 Ma Pb/Pb isochron for selected Uivak gneisses from Labrador (data from Baadsgaard *et al.* (1979), and Oxford unpublished data) is shown at the present day (crosses, line A) and as developed up to 2530 Ma ago (palaeo-isochron, line C) which is the U-Th-Pb age of sphene and zircon from a late granite and other rocks in the area. The Hebron gneiss linear array (line B) is sub-parallel to the calculated palaeo-isochron for Uivak gneisses (line C), indicating that a late Archaean U depletion (at *c*.2500 Ma) from early Archaean protoliths of broadly similar age to the Uivak gneisses is a plausible explanation of the Hebron gneiss Pb isotopic data.

out the unexpectedly high degree of mobility of Sm and Nd during the later granulite-facies metamorphism.

Converse processes, in which ancient, high-grade sialic crust, depleted in Rb, U and other incompatible elements, undergoes much later retrograde metamorphism accompanied by metasomatic introduction of magmatic or non-magmatic fluids containing incompatible elements, will yield easily predictable isotopic effects. An example of this type has been documented by Collerson *et al.* (1982) in gneisses from northern Labrador, where Rb-Sr and Pb/Pb age and isotope studies show that 'crustal reworking due to interaction of old crust (*c.*3500 Ma) with aqueous fluids was accompanied by introduction of granitic melts (*c.*2800 Ma) derived by melting of ancient deep crustal rocks or contamination of juvenile granitic melts by ancient deep crustal material.' The observed effects of crustal reworking on isotopic systematics of the Labrador gneisses are different from those predicted by Collerson & Fryer (1978) in an unconventional model of massive mantle-crust metasomatism and crustal transformation, as discussed below.

Eradication of radiogenic isotope evidence of the crustal residence history of a rock unit — fact or fantasy?

Powerful isotopic criteria now exist for the recognition of juvenile, mantle-derived sialic crust, on the one hand, and of reworked sialic crust, on the other hand. The basic principle is that partial melting and reworking of old continental crust can give rise to a wide variety of initial Sr, Nd and Pb isotope compositions in the resulting magmas and rocks, depending upon the crustal level at which melting or reworking occurs. Granulite-facies lower crust, characteristically with low Rb/Sr, U/Pb and Sm/Nd ratios, with time produces low (mantle-type) initial $^{87}Sr/^{86}Sr$, but unradiogenic Pb and unradiogenic $^{143}Nd/^{144}Nd$ compared to a contemporaneous mantle-type source region; intermediate-level, amphibolite-facies crust, characteristically with higher Rb/Sr, low U/Pb and low Sm/Nd, would produce high initial $^{87}Sr/^{86}Sr$, unradiogenic Pb and unradiogenic initial $^{143}Nd/^{144}Nd$ compared to a contemporaneous mantle-type source region; high-level crust, characteristically with high Rb/Sr, high U/Pb and low Sm/Nd would produce radiogenic initial $^{87}Sr/^{86}Sr$, unradiogenic initial $^{143}Nd/^{144}Nd$, and Pb either similar to, or more radiogenic than, mantle Pb. These are, of course, only three broadly defined regions in what really amounts to a continuous spectrum of geochemical variations within the crust, in which each of the parent/daughter ratios has its own characteristic gradient within a given crustal sector. The most important point to note is that all three combinations noted above are different to the combined radiogenic isotope characteristics of a contemporaneous mantle-type source region (i.e. low Rb/Sr, high U/Pb, high Sm/Nd, with corresponding isotope ratios), However, some combination of crustal and mantle isotopic characteristics can also be produced in mantle-derived magmas undergoing crustal interaction and contamination, although diverse petrological, chemical and other isotopic criteria may be applied to the recognition of this phenomenon.

In contrast to conventional geochronological and isotopic interpretations, a few workers have proposed that some apparently well-dated rock units in early Precambrian cratonic areas have crustal residence ages greatly in excess of their measured isochron ages (e.g. Bridgwater & Collerson 1976; Collerson & Fryer 1978; Hart *et al.* 1981; Welke & Nicolaysen 1981; Bridgwater *et al.* 1981a). Initial Sr, Nd and Pb isotopic compositions, as well as μ_1-values, may provide support for such a proposal if the values of these parameters differ significantly between the rock unit and contemporaneous mantle. However, some claims of an extended crustal residence time for certain rock units have been made without this type of supporting evidence, and these claims are highly controversial.

Stated simply, the controversy hinges on whether or not the isotope geochemistry of several parent-daughter systems can be *completely* reset, and thus give the appearance of a rock unit newly derived from a mantle source long after the original time of rock formation and crustal accretion.

The quoted authors (see above) envisage that such resetting results from some combination of the following processes: (1) essentially complete expulsion of daughter elements out of a rock either by high-grade metamorphism, or by metasomatic 'flushing', (2) swamping of grown-in radiogenic isotope characteristics by influx of daughter elements from a juvenile (e.g. mantle) source, (3) massive influx of parent elements into the rock unit, (4) large-scale isotopic rehomogenization of daughter elements, (5) large-scale redistribution of parent elements within the rock unit.

Whilst individual parent-daughter systems might be disturbed by one or more of these processes, it seems most improbable that the Rb-Sr, Sm-Nd and U-Th-Pb systems would all behave coherently. Consequently, there is little likelihood of obtaining concordant age results from different decay schemes in rock units affected by such 'resetting' processes.

In view of the potentially far-reaching nature of the controversy, two particular cases from South Africa and West Greenland will be discussed.

The Vredefort Dome, South Africa

Hart *et al.* (1981) have recently reported voluminous Rb-Sr and U-Th-Pb whole-rock isotopic data on four principal rock units of the Vredefort Dome. They conclude that crustal evolution of the Vredefort Dome commenced *c.*3800 Ma ago, although the whole-rock isochron dates which they report give no direct support for this conclusion.

The amphibolite-facies Outer Granite Gneiss (OGG) yields near-concordant Rb-Sr, Th-Pb and Pb/Pb whole-rock isochron ages of 3000 ± 30 Ma, 3060 ± 50 Ma and 3080 ± 20 Ma respectively (note that 1σ errors are used here in the discussion of the Vredefort data, following the practice of Hart *et al.* 1981). The initial $^{87}Sr/^{86}Sr$ ratio of 0.7019 ± 0.0002 for OGG is within the range of values typical of late Archaean upper mantle. With the present average $^{87}Rb/^{86}Sr$ ratio of ~ 1.63 for OGG, no significant crustal residence prior to *c.*3050 Ma is permitted. The μ_1-value of 7.74 for the source of OGG is also typical of late Archaean mantle sources.

The *only* possible indication of an earlier crustal history for OGG is provided by the unusually low initial $^{208}Pb/^{204}Pb$ ratio of 31.78 ± 0.14 derived from the Th-Pb isochron. However, the excellent agreement between three independent dating methods, as well as the implausibility of large-scale, regional metamorphic rehomogenization of Sr and Pb isotopes (Moorbath 1975b) are compelling arguments for rejecting the proposal of Hart *et al.* (1981) for an extended crustal pre-history of some 750 Ma for OGG.

The Inlandsee Leucogranofels (ILG), comprising granulite-facies rocks in its central part, but transitional rocks from amphibolite- to granulite-facies in its outer parts, has also been investigated by Hart *et al.* (1981). They interpret its whole-rock Rb-Sr isochron date of 2830 ± 30 Ma as the time of metamorphic homogenization of Sr isotopes. The initial $^{87}Sr/^{86}Sr$ ratio of 0.7044 ± 0.0004 and the average Rb/Sr ratio of 0.2 together permit a crustal residence time of up to *c*.550 Ma for the protoliths of ILG prior to 2830 Ma.

Division of the ILG U-Th-Pb data into two groups, namely the ILG and the Steynskraal Metamorphic Zone (SMZ) felsic rocks is considered by Hart *et al.* (1981) to provide evidence of an early Archaean age for at least those four samples which plot together with data for SMZ metabasic enclaves on a 3590 ± 60 Ma whole-rock Th-Pb isochron. However, seven other ILG samples yield a whole-rock Th-Pb isochron date of 3070 ± 110 Ma and an initial $^{208}Pb/^{204}Pb$ ratio of 31.67 ± 0.22, in good agreement with the OGG Th-Pb study (Fig. 3(a)).

Welke & Nicolaysen (1981) have interpreted U-Pb isotope data on selected ILG and SMZ rock samples as also providing evidence for early Archaean rock formation. Using an ingenious but unconventional three-stage model for Pb isotopic evolution, they conclude that the samples analysed commenced their crustal residence at *c*.3860 Ma and then suffered U-depletion at *c*.2760 Ma.

The validity of these interpretations may be questioned on the following grounds:

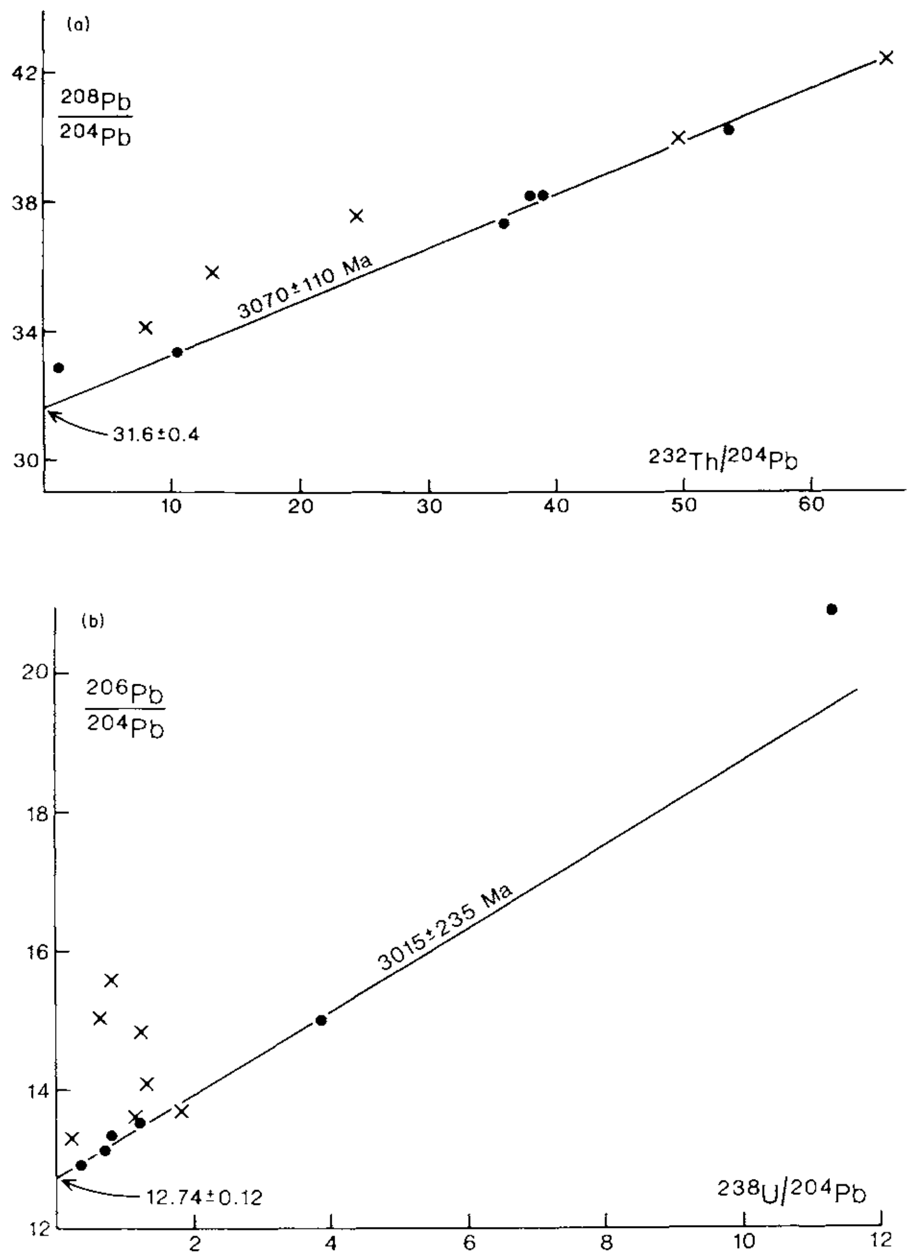
1. Hart *et al.* (1981) state that the SMZ 'includes rocks of sedimentary and igneous origin, which are not necessarily contemporaneous, but geological evidence clearly indicates that they are all older than the Inlandsee Leucogranofels'. The construction of an 'isochron' from SMZ and ILG rocks which are neither contemporaneous nor cogenetic hardly seems justifiable. Such 'isochron' age results must be treated with great caution.

2. Metamorphic homogenization of Sr isotopes is explicitly proposed by Hart *et al.* (1981) in their interpretation of Rb-Sr isochron results on ILG, whilst isotopic homogenization of Sr and Pb on a regional scale is an implicit requirement for explaining other isochron relationships in OGG, ILG and SMZ Rb-Sr and U-Th-Ph systems. However, isotopic studies in many other high-grade metamorphic complexes suggest that metamorphic processes are unable to effect regional rehomogenization of Sr and Pb isotopes (see, for example, the case of the Vikan gneisses, described earlier). Even within the Vredefort crust itself it is clear that SMZ-mafic enclaves did not equilibrate with adjacent ILG host-rock with respect to their Rb-Sr isotopic systems. Another anomaly concerns the near-concordance of OGG isochron results for amphibolite-facies rocks, compared with the marked discordance of ILG and SMZ 'isochron' results on granulite-facies rocks. It would be surprising if amphibolite-facies metamorphism had produced more effective rehomogenization of Sr and Pb isotopes than granulite-facies metamorphism.

3. Hart *et al.* (1981) appeal to major metasomatic influx and/or redistribution of elements affecting the Rb-Sr and U-Th-Pb isotopic systems. Some of the proposed elemental migrations, such as the 'net movement of Sr towards the lower regions of the crust' in the OGG, appear to defy rather well-established geochemical behaviour patterns of elements during metamorphism. In the debate on extended crustal residence, *ad hoc* models of Rb metasomatism have by now become commonplace. This mechanism is invoked where measured initial $^{87}Sr/^{86}Sr$ ratios and present-day Rb/Sr ratios would not otherwise allow a lengthy crustal residence time (e.g. Bridgwater & Collerson 1976; Collerson & Fryer 1978). Hart *et al.* (1981) invoke Rb metasomatism to explain the proposed crustal residence time of OGG between *c*.3860 and 3050 Ma. This hypothesis similarly requires metasomatic addition and/or major redistribution of U and Th, both for OGG and ILG. Observational evidence in support of the postulated metasomatic effects is lacking. No discussion is given regarding the nature of the source or the causes of the proposed metasomatism. The case for metasomatism at Vredefort rests on a circular argument, and this mechanism is invoked solely because it is a necessary requirement for erecting the hypothesis of extended crustal residence time, followed by total eradication of all previous isotopic memory and complete isotopic resetting. That Rb, U and Th metasomatism *may* occur is not in dispute. Indeed, we have ourselves proposed that late Archaean Nûk magmas in West Greenland were contaminated by fluids enriched in Rb, U (probably Th) and Pb derived from rocks undergoing high-grade metamorphism, up to granulite-facies, contemporaneous with the Nûk magmatic event (Taylor *et al.* 1980). However, observed isotopic patterns there in no way match those reported by Hart *et al.* (1981) for Vredefort.

We now propose a markedly different interpretation of the Vredefort isotopic data, which takes into account our criticisms, and which is based on well-established principles of geochemical behaviour of the Rb-Sr and U-Th-Pb isotopic systems.

Hart *et al.* (1981, p. 10667) showed that seven ILG samples define a Th-Pb whole-rock isochron date of 3070 ± 110 Ma and an initial $^{208}Pb/^{204}Pb$ ratio of 31.67 ± 0.22 (Fig. 3(a)). Re-examination of ILG data in the U-Pb and Pb/Pb whole-rock isochron diagrams (Figs. 3(b) and 3(c)) demonstrates that six samples define a poorly-fitted U-Pb isochron giving a date of 3015 ± 230 Ma, whilst seven samples define a well-fitted Pb/Pb isochron with a date of 2980 ± 45 Ma and a μ_1 value of 7.72 for the source of ILG. These ILG isochron results are in mutual agreement, and also agree well with OGG isochron dates. Initial $^{208}Pb/^{204}Pb$ ratios and μ_1 values for OGG and ILG are also in close agreement. In order to achieve these results, a substantial proportion of ILG data points have been excluded from the isochron fitting procedures, namely four from the Th-Pb plot (Fig. 3(a)), seven from the U-Pb plot (Fig. 3(b)) and six from the Pb/Pb plot (Fig. 3(c)). This is justified in the present case in view of the geographical distribution of the samples in relation to the variation of metamorphic grade through the Vredefort Dome. Almost all aberrant data in all three diagrams (Figs. 3(a)–(c)) are samples from the inner, highest-grade (granulite-facies) core of the ILG outcrop, close to the 1950 ± 60 Ma Central Intrusive Granite (CIG). Furthermore, the aberrant data plot above and to the left of the Th-Pb, U-Pb and Pb/Pb isochrons, reflecting retardation of radiogenic Pb evolution by Th and/or U depletion significantly post-dating commencement of isochron development. Aberrant data in the Pb/Pb diagram indicate that in only one case can U- depletion be attributed to very recent U- loss (e.g. by ground water leaching). All other aberrant samples show clear isotopic evidence for earlier U- depletion and have very low $^{238}U/^{204}Pb$ ratios (< 1.5). However, the palaeo-isochron for 2980 Ma-old rocks at 1950 Ma (Fig. 3(c)) demonstrates that none of the aberrant samples need have suffered U-



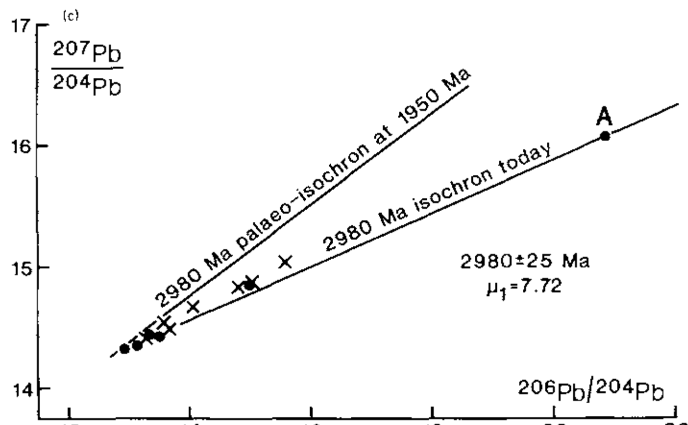


FIG. 3. Whole-rock data plots for the Inlandsee Leucogranofels (ILG) of the Vredefort Dome. Data are from Hart *et al.* (1981). Circles represent samples from the outer, lower-grade part of the ILG outcrop; crosses represent samples in the inner granulite-facies area of ILG, close to the Central Intrusive Granite (CIG). (a) Plot of $^{208}Pb/^{204}Pb$ vs. $^{232}Th/^{204}Pb$. Seven of the samples define an isochron of 3070 ± 110 Ma (1σ error). The four data points falling above the isochron are here interpreted as resulting from Th depletion significantly later than *c*.3000 Ma. (b) Plot of $^{206}Pb/^{204}Pb$ vs. $^{238}U/^{204}Pb$. Six of the samples define an isochron of 3015 ± 235 Ma (1σ error). Seven samples plot above the isochron indicating loss of U significantly later than *c*.3000 Ma. Six of these aberrant samples, all from the inner granulite facies area of the ILG, have $^{238}U/^{204}Pb$ < 1.5. (c) Plot of $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$. Seven of the samples define an isochron of 2980 ± 25 Ma (1σ error), with a μ_1 value of 7.72. Data for six samples plot well above the isochron, reflecting U depletion significantly later than *c*.3000 Ma, but long before the present. The palaeo-isochron for 2980 Ma as developed up to 1950 Ma demonstrates the possibility that partial U depletion of ILG samples could be related to emplacement of the *c*.1950 Ma-old CIG, particularly as no ILG data points plot above the palaeo-isochron. Sample A also shows evidence of U loss in the U-Pb isochron diagram (Fig. 3(b)), but plots on the present Pb/Pb plot, demonstrating that U loss in this case was a recent event.

depletion prior to 1950 Ma. The aberrant data points are therefore interpreted as reflecting U- and Th- depletion during high-grade metamorphism of ILG up to granulite facies, which may be related temporally and spatially to the emplacement of the 1950 ± 60 Ma CIG or to the so-called Vredefort 'catastrophe' that formed the domal structure about 2000 Ma ago and which is associated in space and time with an extended period of high-temperature static metamorphism and a shock event causing high-intensity dynamic metamorphism (Schreyer 1983).

The Rb-Sr isochron date of 2830 ± 30 Ma for ILG might be significantly lower than the Rb-Sr date from OGG and from U-Th-Pb dates on both OGG and ILG for one or more of the following reasons:

1. Loss of radiogenic Sr during a protracted cooling history after rock formation (see earlier discussion).
2. Loss of radiogenic Sr during a later metamorphic event.
3. Loss of Rb during high-grade metamorphism, especially from samples with the lowest Rb/Sr ratios.

Possibility (3) is not favoured. Not all samples with low Rb/Sr ratios are from the inner granulite-facies ILG outcrop in which the severest Rb-depletion might be expected. Indeed, Rb concentrations in ILG do not indicate severe metamorphic depletion of Rb.

Our 'least-extravagance' two-event interpretation of OGG and ILG isotope data differs fundamentally from the model of crustal evolution in the Vredefort Dome proposed by Hart *et al.* (1981), which requires several geochemical events to explain the isotopic characteristics of OGG and ILG. Several of these events are essentially necessities in a model in which early Archaean crust formation is an *a priori* assumption. The contrasted hypotheses of early Archaean (Hart *et al.* 1981) and late Archaean (this paper) crust formation at Vredefort should provide a convenient test case for Sm-Nd model age dating of crust-mantle separation (McCulloch & Wasserburg 1978; Hamilton *et al.* 1979a). However, for the present, we consider that the Vredefort sialic crust was generated from a mantle-type source region at about 3100 Ma ago — i.e. as given by the concordant ages reported by Hart *et al.* (1981) for OGG — and has no pre-history dating back to *c*.3860 Ma. We maintain that the case for eradication of radiogenic isotopic evidence after a long crustal residence time proposed for Vredefort by Hart *et al.* (1981) remains unproven.

Kangimut sangmissoq, Godthaabsfjord region, West Greenland

At this locality, which is situated some 30 km east of the main outcrop of reliably dated 3650 Ma-old amphibolite-facies Amitsoq gneiss (Moorbath *et al.* 1972), there is a small area of retrogressed (amphibolitized) granulite-facies gneisses. By application of field criteria (McGregor 1973) several workers now regarded these rocks as true Amîtsoq gneisses which underwent granulite-facies metamorphism at about 2800 Ma ago and which were subsequently retrogressed to amphibolite-facies (Bridgwater *et al.* 1981a; McGregor pers. comm. 1982). The salient isotopic facts for the Kangimut sangmissoq gneisses (KSG) (Oxford unpublished data) are as follows: (1) A whole-rock Rb-Sr errorchron (six data points) yields a date of 2770 ± 160 Ma, with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7019 ± 0.0005. (2) Five samples yield Sm-Nd model ages close to 2800 Ma. (3) The Pb/Pb isotope pattern of 18 samples is exactly analogous to the pattern observed in those *c*.2800–3000 Ma-old Nûk gneisses of the Godthaabsfjord region whose magmatic precursors were emplaced through and into Amîtsoq gneisses and which became contaminated with isotopically easily characterized, unradiogenic Amîtsoq-type Pb (Taylor *et al.* 1980). Between 0–30% of total Pb in analysed KSG samples is derived from Amîtsoq gneiss. The Pb/Pb isotope pattern in KSG differs fundamentally from that in the Amîtsoq gneisses (Black *et al.* 1971; Baadsgaard *et al.* 1976; Gancarz & Wasserburg 1977).

The observed isotopic data cannot be reconciled with the view that KSG represent reworked Amîtsoq gneisses, as suggested by Bridgwater *et al.* (1981a) and McGregor (pers. comm. 1982). These workers postulate that the *c*.2800 Ma granulite-facies metamorphism totally purged the *c*.3650 Ma-old Amîtsoq gneisses of Rb, Sr, U, Pb, REE (and, presumably, other elements) including the radiogenic isotopes which had built up during a period of some 800 Ma, and that not long afterwards they were replaced during retrogression by the same suite of elements transported in juvenile, mantle-derived fluids with mantle-type isotopic signatures. We do not believe that granulite-facies metamorphism can totally remove all these elements from deep crust, though it is well known that Rb and U (and K) may be severely depleted in medium-to-high pressure granulite-facies rocks. However, granulite-facies gneisses are not normally severely depleted in Pb, Sr and Nd, which carry the isotopic memory of crustal residence time. It is difficult to see how they could be so completely replaced by mantle-derived fluids as to leave no isotopic memory.

More detailed discussion of the controversial KSG will be presented elsewhere. However, it appears that the geological field criteria erected by McGregor (1973) in the main part of the Godthaabsfjord region, in which basic dykes (the Ameralik dykes) are used to separate the Amîtsoq and Nûk gneisses, cannot be extended to the whole of the Godthaabsfjord region. It is most unlikely that Rb-Sr, U-Pb and Sm-Nd isotopic records can all be *totally* reset without trace of a previous crustal residence time of some 800 Ma.

Dating Precambrian supracrustal rocks

Supracrustal rocks may be either dated directly, or age-bracketed between dated rocks such as basement gneiss and cross-cutting intrusives, although the latter approach may not always produce an adequately tight age-bracket. Sediments or metasediments can sometimes be dated directly but, more commonly, they are age-bracketed by stratigraphically intercalated, dated volcanic rocks. A wide variety of techniques is therefore available for constraining the age of deposition of supracrustal rocks, provided that the rocks are either unmetamorphosed or at low metamorphic grade. Many Archaean greenstone belts, as well as Proterozoic supracrustals, have been dated quite precisely (e.g. see earlier discussion of high-precision zircon U/Pb dating). However, complex high-grade gneiss terrains containing supracrustal enclaves are more likely to give a blanket age with no possibility of age resolution. In such areas, Sm-Nd dating of basic meta-igneous enclaves in calc-alkaline orthogneisses may offer one fruitful approach.

Direct dating of sedimentary rocks

The geochronology of Precambrian sediments, in particular of those which contain the earliest known biogenic markers and primitive fossils, is a topic of particular current interest. The direct-dating approach frequently meets with methodological and interpretative difficulties. Chemical sediments containing biogenic remains, e.g. limestone and chert, may not have adequate contents, or ratios, of parent and daughter nuclides — although this merits much further research, as there are now grounds for optimism in applying the Pb/Pb method to the direct dating of some chemical sediments (see later). In clastic rocks the presence of unrecrystallized detrital minerals (e.g. micas, feldspars, zircon, some clay minerals etc.) may yield erroneously old ages. Moreover, fine-grained sediments of any type may be affected by low-grade metamorphism and/or metasomatism resulting in open-system behaviour of parent and/or daughter nuclides with accompanying partial or complete resetting of ages. Fine-grained volcanic rocks frequently suffer from the same limitations.

K-Ar and Rb-Sr dating of unaltered, authigenic glauconite from rocks that have not been deeply buried, deformed or metamorphosed may yield a depositional age. This approach has been used for calibrating the younger parts of the Phanerozoic time-scale. Glauconite is rare in Precambrian rocks, but good K-Ar dates have been reported from independently age-bracketed *c.*1600 Ma–old sediments from the Northern Territory of Australia (Webb *et al.* 1963; McDougall *et al.* 1965), whilst reliable Rb-Sr and K-Ar dates of *c.*1100 Ma have been published for the Belt Series of Montana (Obradovich 1968). Doubts about the reliability of glauconite dating in more complex situations were expressed early on by Hurley *et al.* (1960), whilst hydrothermal experiments on glauconite (Odin *et al.* 1977) demonstrated that low-temperature metamorphism affects the apparent isotopic age of glauconite grains to an extent which depends on their chemical composition and temperature of metamorphism.

In the early 1960s, it was discovered that fine-grained, unmetamorphosed, predominantly detrital sedimentary rocks, such as shales, could yield whole-rock Rb-Sr isochrons (or errorchrons) whose slopes indicated the time elapsed since Sr isotope homogenization occurred within the particular rock units. Good agreement was obtained in some cases for stratigraphically well-defined Palaeozoic sediments between the measured Rb-Sr age and the independently calibrated age for the relevant part of the Phanerozoic time-scale (Compston & Pidgeon 1962; Whitney & Hurley 1964; Bofinger & Compston 1967; Bofinger *et al.* 1970). The method was subsequently applied to unmetamorphosed Precambrian — mostly Proterozoic — sediments, and in some cases the agreement between the measured Rb-Sr age and an independently bracketed age for a given sediment was satisfactory (Chaudhuri & Faure 1967; Obradovich & Peterman 1968; Faure & Kovacb 1969; Moorbatb 1969; Pringle 1973). A whole-rock Rb-Sr isochron age of 1595 ± 24 Ma for shales from the Gunflint Iron Formation of Ontario, Canada, was of particular interest in view of its varied assemblage of well-preserved microfossils in chert horizons (Faure & Kovach 1969).

The favoured explanation for the reasonable agreement between measured and expected age in some fine-grained sediments is that chemical and mineralogical changes during diagenesis of fine-grained argillaceous sediments are sufficient to produce both homogenization of $^{87}Sr/^{86}Sr$ ratios and localized variations in Rb/Sr ratios within a single stratigraphic horizon. These are the fundamental requirements for the Rb-Sr dating of *any* rock. Thus it is possible to date the time of diagenesis which may be penecontemporaneous with deposition.

In many cases, whole-rock Rb-Sr measurements on fine-grained sediments yield problematic age results. This is because Sr isotope homogenization not only results from diagenesis penecontemporaneous with deposition, but also from much later diagenesis, structural deformation and metamorphic recrystallization. Greenschist-facies metamorphism effectively resets whole-rock Rb-Sr dates to the time of metamorphism (Peterman 1966; Powell *et al.* 1969). In contrast, presence of detrital minerals from an older source region and resistant to isotopic equilibration (e.g. muscovite) may result in measured ages that exceed the true depositional age (Chaudhuri & Faure 1967). These methodological and interpretative difficulties have long been realized (e.g. Bofinger & Compston 1967; Dasch 1969; Clauer 1973; Perry & Turekian 1974; Gebauer & Grünenfelder 1974; Chaudhuri 1976; Spanglet *et al.* 1978).

A new chapter in the Rb-Sr dating of sediments has opened up with the work of Clauer (1979; this Volume), who maintains that many published Rb-Sr age studies on sediments and metasediments are misleading because, (1) most studies, with the exception of some on glauconites and separated clay minerals, were carried out on whole rocks, (2) most studies were carried out without detailed mineralogical control, (3) the metamorphic history of the sediments was usually unknown or disregarded.

In Clauer's approach, it is essential to separate an argillaceous sediment into its constituent mineral and size fractions, and then to identify the different clay minerals by X-ray techniques, in order to characterize whether they are detrital, authigenic, diagenetic or metamorphic, and thus to distinguish clearly between these origins. Rb-Sr dating is then carried out on different minerals and size fractions, so that different stages in the history of the sediment may be identified, but remembering that low-grade metamorphism may overprint and erase the age record. These newer techniques have not so far been seriously applied to Archaean sedimentary rocks, most of which show the effects of at least low-grade metamorphism, and are thus unlikely to yield a valid depositional Rb-Sr age unless metamorphism happened to be penecontemporaneous with deposition within the error limits. Recent applications to Proterozoic and Phanerozoic sediments from several continents have been reported by Clauer (1979), and in a special volume dealing with the geochronological correlation of Precambrian sediments and volcanics in stable zones, edited by Bonhomme (1982).

We now consider several case-histories involving attempts to date Archaean sediments and metasediments. The Fig Tree Shale from the Swaziland System of the Barberton region of southern Africa has yielded a good whole-rock Rb-Sr isochron age of 2910 ± 40 Ma (Allsopp *et al.* 1968), although this most probably dates a low-grade metamorphism because cross-cutting granites yield significantly older whole-rock Rb-Sr ages of *c.*2960 –3000 Ma (quoted in Allsopp *et al.* 1968). The Fig Tree Shale date is thus a minimum age of sedimentation and of the fossil biota which they bave long been recognized to contain (Schopf & Baarghorn 1967). The Fig Tree Group as a whole is some 3500 m thick, and immediately overlies the Onverwacht Group, which on conventional estimates is about 15200 m thick (Anhaeusser 1973). It is emphasized that the brief, repeated discussions in the following pages of the Swaziland System are based on this conventional interpretation. However, a re-evaluation of the stratigraphy and thickness of the entire Barberton belt is currently under way, resulting from the recognition of nappe and overthrust tectonics, with major tectonic repetitions (De Wit 1982).

Subsequently, Hurley *et al.* (1972) reported a whole-rock Rb-Sr isochron age of 3275 ± 70 Ma for the fine-grained, quartz-mica-calcite-bearing 'Middle Marker Horizon' of the Onverwacht Group, whose sedimentology has recently been described by Lanier & Lowe (1982). The low, initial $^{87}Sr/^{86}Sr$ ratio of 0.7015 ± 0.0018 precluded significant crustal residence time and suggests that the measured age is close to the true depositional age (Hurley *et al.* 1972). The Middle Marker Horizon is some 7700 m below the base of the Fig Tree Group discussed earlier. Despite this apparent depth of burial, some parts of the Onverwacht Group have remained undeformed and virtually unmetamorphosed. This is quite difficult to understand if current estimates of the thickness of the Onverwacht Group are correct.

The microstructures in some Onverwacht Group sediments are currently regarded as true fossils (Knoll & Baarghorn 1977), particularly those in the *c.*900 m-thick Swartkoppie Formation, *c.*6800 m above the Middle Marker Horizon and immediately underlying the Fig Tree Group (Anhaeusser 1973). Similar examples are found within the 1920 m-thick Kromberg Formation, with its base *c.*4850 m above the Middle Marker Horizon. More problematical ones occur within the 1900 m-thick Theespruit Formation (Muir & Grant 1976) whose top is some 3500 m below the Middle Marker Horizon. From the limited age evidence so far discussed there is little doubt that if the Onverwacht and Fig Tree microstructures are indeed biological (Schopf 1976), then cellular organisms existed by about 3300–3200 Ma ago. Further evidence bearing on the age of deposition of the Onverwacht Group from intercalated volcanics is presented in a subsequent section.

Direct dating of the early Precambrian Isua supracrustals of West Greenland was mentioned earlier and summarized in Table 2. Brief geological descriptions of the area have been published by Allaart (1976), Bridgwater *et al.* (1976), and Appel (1980). In contrast to the South African samples, intense deformation and amphibolite-facies metamorphism (*c.* 550°C and 5 kbar) have obliterated most primary igneous and sedimentary textures (Boak & Dymek 1982). Nonetheless, the original nature of some of the rocks can be clearly discerned and the Isua supracrustals exhibit many similarities to the much larger greenstone belt assemblages of later Archaean times. Age and isotopic data suggest that the interval between their deposition and first metamorphism was short (> 100–150 Ma). No field or isotopic evidence has yet been found concerning the age and nature of the substratum on which the Isua supracrustals were deposited.

In view of the intense metamorphism suffered by the Isua supracrustal rocks and the resulting interpretative problems regarding biogenic markers, it is not yet possible to say whether life as we know it existed in Isua times. Controversial morphological, chemical and isotopic evidence from the Isua supracrustals has given rise to highly conflicting interpretations (e.g. Monster *et al.* 1979; Pflug & Jaeschke-Boyer 1979; Schidlowski *et al.* 1979; Bridgwater *et al.* 1981b; Nagy *et al.* 1981; Roedder 1981; Walters *et al.* 1981). There is nothing in the nature of the palaeo-environmental evidence, as deduced from the geological character of the Isua sediments, to contradict the existence of early life, whilst the occurrence elsewhere of stromatolites and cellular micro-organisms some 300–500 Ma later (see below) adds an indirect measure of plausibility. One may hope that relatively unmetamorphosed and undeformed supracrustal rocks of Isua age will turn up somewhere. Undoubtedly, several early Archaean terrains still remain to be discovered, mapped and dated.

Indirect dating of sedimentary rocks

Dating of stratigraphically intercalated volcanic rocks

Volcanic and metavolcanic rocks frequently have suitable parent/daughter ratios for Rb-Sr and/or U-Th-Pb and/or Sm-Nd dating. The Rb-Sr method has been the most frequently used for dating Precambrian volcanics, although basic volcanics tend to have low and uniform Rb/Sr ratios. Good examples of Rb-Sr dating in Archaean greenstone belts have been reported from Northwest Territories, Canada (Green & Baadsgaard 1971), north-eastern Minnesota, USA (Jahn & Murthy 1975), Manitoba, Canada (Clark & Cheung 1980), Finland (Vidal *et al.* 1980), and from many other areas.

Just as with fine-grained sediments, post-depositional low-grade metamorphism, hydrothermal alteration and metasomatism can yield large data scatter on isochron plots and/or anomalously low ages, particularly in the Rb-Sr and U-Pb systems. An example is provided by a suite of acid volcanic rocks from near the middle of the Onverwacht Group which yields a whole-rock Rb-Sr isochron date of 2560 ± 40 Ma (Allsopp *et al.* 1973). This is quite incompatible with whole-rock Rb-Sr dates of 3275 ± 70 Ma and 2910 ± 40 Ma for the Middle Marker Horizon and Fig Tree Shale discussed earlier, or with whole-rock Rb-Sr dates of *c.*3000 Ma for coarse-grained granites which cut the Swaziland Sequence (Allsopp *et al.* 1973). Another example has been reported from the late Archaean Chibougamou greenstone belt of Quebec, Canada, where acid volcanics yield a whole rock Rb-Sr isochron age of 2290 ± 170 Ma, but are cut by a tonalite pluton with a whole-rock Rb-Sr isochron age of 2520 ± 160 Ma (Jones *et al.* 1974). Acid volcanics can be particularly prone to loss of radiogenic Sr as a result of post-depositional disturbance, because they contain little or no Ca-bearing minerals which can act as acceptors for displaced radiogenic Sr, or if they do contain calcic phases these may be blocked to Sr diffusion at greenschist-facies temperatures. These cautionary remarks also apply to Phanerozoic acid volcanics used for time-scale calibration. In contrast, isotope systematics of *basic* volcanic rocks are frequently disturbed by post-depositional introduction of mobile elements such as Rb and U.

Several techniques overcome these problems. Accessory zircons in acid volcanics can yield meaningful U-Pb ages even when whole-rock Rb-Sr and/or U-Th-Pb systematics have been disturbed or reset. Of particular importance is the recent application of the Sm-Nd method to basic volcanic and metavolcanic rocks in Precambrian supracrustal successions, reviewed by O'Nions *et al.* (1979b) and De Paolo (1981a). Because Sm and Nd are chemically such closely related elements, they may be used in cases where the application of other decay schemes such as Rb-Sr and U-Pb is hampered by post-depositional disturbances, leading to chemical fractionation of parent and daughter element. Indeed, Sm and Nd do not appear to be significantly fractionated by crustal processes such as erosion, transport, diagenesis, hydrothermal alteration or metamorphism (McCulloch & Wasserburg 1978), making this a particularly powerful method for constraining the age of separation of protoliths of sedimentary rock sequences from mantle sources (see O'Nions & Hamilton, this volume).

Indirect dating, using stratigraphically related volcanic horizons, has been applied to the early Archaean stromatolite-bearing Warrawoona Group of the eastern Pilbara region of western Australia (Lowe 1980). The fossils occur in the Strelley Pool chert near the top of a 10 km section of the Warrawoona Group, consisting largely of low-grade meta-volcanics interstratified with chert units less than 30 m thick (Barley *et al.* 1979). The fossiliferous cherts lie 1–2 km above dacitic lavas of the Duffer Formation, from which zircons have given a U-Pb age of 3452 ± 16 Ma (Pidgeon 1978), interpreted as the age of extrusion. Lowe (1980) is careful to point out that the 'possibility of intervening unconformities cannot be dismissed, although the overall igneous and sedimentological continuity of the Group suggests that major hiatuses are absent. Following deposition of the uppermost part of the Warrawoona Group and the overlying Gorge Creek sediments, the rocks were extensively deformed and intruded by granitoid plutons. The main period of deformation and associated plutonism has been dated at *c.*3100 – 2900 Ma (De Laeter & Blockley 1972; Oversby 1976). The possible age range for the stromatolites is, therefore, *c.*3450 – 3100 Ma, but their close association with the older rocks suggests deposition at 3400 Ma'. Further down the succession, the North Star basalt has yielded a whole-rock Sm-Nd isochron age of 3560 ± 32 Ma (Hamilton *et al.* 1981). Also in western Australia, Walter *et al.* (1980) have reported stromatolites from a chert-barite unit in the Warrawoona Group at a locality named North Pole. The unit consists of bedded, shallow-water greenschist facies metasediments 40 m thick that persists for at least 30 km along strike. This unit is regarded as the approximate stratigraphic equivalent of the dated Duffer Formation.

Returning now to the Onverwacht Group of southern Africa, Sinha (1972) studied common Pb and U-Pb systematic of five basaltic lavas and obtained a 'best' age of 3290 ± 45 Ma. The paper neither quotes the decay constants used, nor the stratigraphic position of the dated samples within the *c.*15000 m-thick Onverwacht Group. However, the age agrees well with the Rb-Sr age of 3275 ± 70 Ma for the Middle Marker Horizon (Hurley *et al.* 1972). Jahn & Shih (1974) have reported a Rb-Sr isochron age of 3420 ± 20 Ma (initial $^{87}Sr/^{86}Sr$ = 0.70048 ± 0.00005) for separated minerals from a basaltic komatiite in the 3500 m thick Komati Formation immediately underlying the Middle Marker Horizon. Hamilton *et al.* (1979b) have reported a Sm-Nd age of 3540 ± 30 Ma for a suite of basaltic and peridotitic komatiites from the Komati Formation. Note that, on current interpretations of the thickness of the Onverwacht Group, the top of the Komati Formation is stratigraphically some 7700 m below the top of the Swartkoppie Formation, from which the most convincing micro-organisms in the Onverwacht Group have been reported to date (Knoll & Baarghoorn 1977).

In Zimbabwe, Precambrian stromatolitic limestones have long been recognized (Schopf *et al.* 1971). Some of the best examples are from the Belingwe greenstone belt (Bickle *et al.* 1975) which has been stratigraphically correlated with dated volcanic horizons in neighbouring areas. The Belingwe sediments and volcanics lie unconformably on gneissic basement, and indicate a depositional environment interpreted as changing from beach, through tidal flat, to deeper water conditions. Hawkesworth *et al.* (1975) reported whole-rock Rb-Sr isochrons on volcanics from several greenstone belts, although the analytical data from the Belingwe belt itself was too scattered for meaningful age interpretation. Stratigraphically broadly equivalent greenstone belt volcanics from three adjacent areas yielded whole-rock Rb-Sr errorchron dates of 2480 ± 180, 2470 ± 280 and 2650 ± 140 Ma, with correspondingly scattered initial $^{87}Sr/^{86}Sr$ ratios. The data scatter was attributed to locally variable post-depositional redistribution of Rb and/or Sr. In contrast, Jahn & Condie (1976) obtained a whole-rock Rb-Sr isochron of 2690 ± 70 Ma for eleven selected samples from the Belingwe belt volcanics. Subsequently, Hamilton *et al.* (1977) reported a whole-rock Sm-Nd age of 2640 ± 140 Ma for ten mafic and ultramafic lavas from several greenstone belt successions, including four samples from Belingwe. At the time that the Belingwe stromatolites were reported (Bickle *et al.* 1975), they were amongst the oldest known. Since then much older ones have been reported from western Australia (see above) and Zimbabwe (see below).

Dating by bracketing between basement and intrusives

This is clearly an extension of the previous approach. Within the framework of a single, major CADS, the bracketing limits may be ≤ 50–100 Ma. Where only one bracket can be closed it may only be possible to set a close upper or lower age limit for deposition. The literature abounds in examples of Precambrian supracrustal associations which have been age-bracketed in this way, and in some cases this bracketing method may yield more reliable results than the direct methods. Two cases were already mentioned earlier in which cross-cutting plutons gave older measured dates than the intruded supracrustals. As explained above, this happens because of the generally higher susceptibility of fine-grained sediments and volcanics to subsequent alteration and metamorphism than coarse-grained plutonic or hypabyssal rocks.

In Zimbabwe, Orpen & Wilson (1981) have reported the 'best examples of *c.*3500 Ma biogenic structures (stromatolites) known which add to the growing body of evidence that life was manifest as early as *c.*1000 Ma after the formation of the earth'. These stromatolites occur in limestone of the Fort Victoria greenstone belt, tentatively assigned to the *c.*3500 Ma-old Sebakwian Group (Wilson *et al.* 1978; Wilson 1979; Nisbet *et al.* 1981). The limestone is regarded as older than the nearby Mushandike granite, though the two rock units are not seen in direct contact. The Mushandike granite has yielded a whole-rock Rb-Sr age of 3445 ± 260 Ma (Hickman 1974). The limestone was formerly regarded as part of the *c.*2600 – 2700 Ma-old Bulawayan succession. Near Selukwe, some 90 km north-west of the stromatolite locality, the Sebakwian succession is cut by the Mont d'Or Granite which has yielded a whole-rock Rb-Sr age of 3340 ± 120 Ma (Moorbath *et al.* 1976) and a whole-rock Pb/Pb age of *c.*3370 Ma (Oxford unpublished data). The Sebakwian Group may be underlain by orthogneisses which have yielded whole rock Rb-Sr ages of 3490 ± 400 Ma (Hawkesworth *et al.* 1975) and 3480 ± 120 Ma (Moorbath *et al.* 1977b) respectively some 30 km west-north-west and 75 km west-south-west of the stromatolite locality. The above dates and their associated errors, as well as the uncertainties in geological correlation, show that the age of deposition of the Sebakwian Group is not yet as satisfactorily bracketed as might be desired. The situation is further complicated by the fact that the presumed *c*. 3500 Ma-old limestone of Orpen & Wilson (1981) has recently yielded a Pb/Pb isochron age of *c*.2800 Ma (Oxford unpublished data) which is provisionally interpreted as the age of deposition, rather than as isotopic resetting. Thus the identification of the stromatolitic limestone as Sebakwian is still an open question.

The dating of the younger (Bulawayan) greenstone belt volcanics of Zimbabwe and their stromatolitic horizons, referred to previously, has been supplemented by whole-rock Rb-Sr ages of 2570 ± 30, 2630 ± 140 and 2590 ± 80 Ma for the cross-cutting Chilimanzi granite (Hickman 1978), Sesombi and Somabula tonalites (Hawkesworth *et al.* 1975) which provide a younger age limit for the supracrustal rocks. Although the *c*.2700 – 2600 Ma Belingwe greenstone belt with its stromatolites (Bickle *et al.* 1975) is demonstrably underlain by basement gneisses, it is not yet clear how much of this basement is made up of early (*c*.3500 Ma) or late (*c*.2900 – 2700 Ma) Archaean gneisses.

Another important stromatolite locality occurs in a dolomite unit of the Yellowknife Supergroup in the northern part of the Slave Structural Province, north-western Canada (Henderson 1975). The stromatolites are regarded as having a minimum age of *c*.2500 Ma on the basis of correlation with probably broadly equivalent volcanic rocks of the Yellowknife Group and their intrusive rocks from the southern part of the Slave Province at Yellowknife, some 550 km south of the stromatolite locality (Henderson 1975). Here, the metavolcanics have yielded a whole-rock Rb-Sr age of 2570 ± 150 Ma, whilst zircon U-Pb ages of *c*.2540 – 2510 Ma are reported for three successive suites of intrusive igneous rocks (Green & Baadsgaard 1971).

Finally, convincing stromatolites from the *c*.2500 m-thick Pongola Supergroup of southern Africa are almost certainly *c*.3000 Ma old (Mason & von Brunn 1977). The Pongola Supergroup unconformably overlies granites dated at *c*.3200 – 3100 Ma, whilst a zircon U-Pb measurement from a volcanic quartz-porphyry in the Upper Insuzi group yielded an age of *c*.3050 Ma (Van Niekerk & Burger 1978). A minimum age is provided by a whole-rock Rb-Sr age of 2805 ± 30 Ma on acid and basic rocks of the Usushwana igneous complex, which cuts the Pongola Supergroup (Davies *et al.* 1970).

Summary and conclusions

Geochronological and associated radiogenic isotope measurements on igneous, sedimentary and metamorphic rocks have now developed to the point where they can be used, in principle, with a high degree of confidence to establish the temporal and broad petrogenetic evolution of a given sector of continental crust to distinguish between juvenile, mantle-derived sialic crust and reworked continental crust, and to ascertain the relative proportions of juvenile and reworked material. Only a few Precambrian shield areas have been studied in sufficient detail to establish more than the broad outlines of crustal evolution, whilst the rather common combination of too few isotopic data with inadequate geological control can easily lead to widely conflicting interpretations and models.

One of the most interesting problems for the future is the detailed elucidation of the physical, chemical and geological processes which comprise a crust-accretion-differentiation superevent (CADS) by a truly multi-disciplinary approach, in which geochronology and isotope geochemistry can make a substantial contribution. This is closely linked with such problems as the growth rate of continental crust through geological time, the complementary geochemical evolution (depletion) of the upper mantle, and the quantitative role (if any) of recycling of continental crust-derived material through the mantle. Many global models, heavily dependent upon radiogenic isotopes, have been proposed. Some require the entire mass of continental crust to have been produced very early in the earth's history (*c*.4000 – 3500 Ma ago) and recycled through the mantle at a decreasing rate throughout geological time (e.g. Armstrong 1968; 1981a). Most widely acceptable, on the basis of multidisciplinary evidence, are those models that require an accelerating growth rate of sialic crust between *c*.3700 and 2500 Ma, by which time perhaps as much as *c*.70 – 80% of the present mass of sialic crust was already in existence, followed by a decelerating growth rate to the present (e.g. Jacobsen & Wesserburg 1979; O'Nions *et al.* 1979a; De Paolo 1980; Jacobsen & Wasserburg 1981; O'Nions & Hamilton 1981; Dewey & Windley 1981; McLennan & Taylor 1982). However, the mathematical isotope-modelling in these two contrasted models of crustal evolution may be non-unique in distinguishing between them (Armstrong 1981b; De Paolo 1981b). From this viewpoint, a decreasing rate of crustal recycling through the mantle throughout geological time is equivalent to a decreasing rate of irreversible chemical differentiation of the mantle. Definitive evidence must clearly be sought elsewhere, not least from geology and geochronology. At any rate, Armstrong's (1981a) objection to continental growth on the basis of the constant freeboard model of Wise (1974) has been convincingly answered by Dewey & Windley (1981) and by McLennan & Taylor (1982), who maintain from independent arguments that continental growth and constant freeboard are entirely compatible.

Evidence for age, residence time and extent of Precambrian sialic crust can be obtained from direct dating of exposed rocks or of detrital components (e.g. zircon, feldspar, muscovite), from Pb/Pb isotopic mixing lines between magmatic rocks and contaminating basement (e.g. Taylor *et al.* 1980; Dickin 1981), or from Sm-Nd model ages of sediments (e.g McCulloch & Wasserburg 1978). Extensive coverage is now needed for all shield areas. So far, the oldest known terrestrial rocks, the Isua supracrustals (Table 2), have not yielded any age or isotope evidence for pre-3800 Ma-old sialic crust (Oxford unpublished data). At the same time, great restraint is advisable in the attribution of early Archaean ages (> 3500 Ma) from indirect age and isotope evidence interpreted on the basis of geochemically implausible models. The same caution must be voiced with respect to several recent reports of directly measured ages exceeding *c*.3800 Ma, for which the data base is insufficient and the documentation inadequate.

Another major topic addressed in this paper is the geochronological evidence bearing on early life. Taking account the quoted errors on individual age determinations, as well as uncertainties in geochronological/stratigraphical correlations, it seems probable from the summarized evidence that stromatolite-bearing rocks were being deposited by *c*.3400 – 3300 Ma ago, and that Archaean microfossils showing cell division were in existence by *c*.3200 – 3000 Ma ago. Additional work to confirm and possibly narrow these estimates should be carried out on rocks which are stratigraphically closely associated in space and time with the fossiliferous horizons, or on the latter themselves. There is ample scope for the development of promising new variants of existing dating methods, in particular Pb/Pb dating of banded iron-formation (so far only reported from Isua, see Table 2), as well as continuing research into applying different age methods to limestones and cherts. Detailed collaborative, interdisciplinary research between palaeobiologists, geologists and geochronologists will be essential.

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Subdivision of the Precambrian

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SUMMARY: The Precambrian record is so fragmentary and from such widely contrasting terrains that attempts to erect any formal subdivisions would appear to be premature. The methods that have been used in the Phanerozoic rocks cannot be used for subdividing the Precambrian as they are peculiarly suitable for fossiliferous largely undeformed sedimentary units in which the sedimentary record is relatively full. Isotopic dating methods also become so imprecise over the major portion of Precambrian time that it is only rarely possible to use them to define particular events with the same precision as those dated in the Phanerozoic record, although this is changing with increasing analytical accuracy.

The use of biological criteria for the erection of a chronostratigraphic scale is seen as a future possibility, but the methods of defining the units that are used in the Phanerozoic are thought inappropriate, and the characterization of a particular chronostratigraphic unit (e.g. by the presence of metazoan trace fossils) is seen as more appropriate.

International agreement on the time for particular boundaries in the Precambrian is regarded as an attempt merely to define the terms already in use rather than having any scientific basis. It is possible that future research may reveal global events which are recognized to be unique and these may provide more reliable datum planes to use. If they are also related to changes in the way the earth behaved through time it would be logical to use these as markers for any major subdivisions of Precambrian time. None have so far been generally accepted, so any units must, at present, be purely informal or local.

A large proportion of Precambrian rocks presently exposed are not undeformed sedimentary sequences but are metamorphosed and deformed complexes of sediments and igneous intrusions, now often gneissose. Such complexes are not amenable to normal stratigraphic subdivision, and their lithostratigraphic nomenclature is also not simply related to the nomenclature of sedimentary piles. It is suggested that a lithotectonic hierarchy of terms be used for the description of such complexes with a chronotectonic major unit, the 'Cycle', being used to define the development of such complexes within a specific time in a particular place. Precambrian chronotectonic cycles may then be used to describe the formation of the rocks at a particular time without implying that the cycle is necessarily of world wide development. Such cycles should never be used as the basis for a Precambrian stratigraphic scale.

After many years of suggestions, proposals and questionnaires, Precambrian stratigraphic subdivision has barely arrived at an internationally agreed starting point. Some schemes have been used locally but most national schemes use different methods and different criteria for the subdivision so there is no international agreement as to how the rocks should be divided or what the names of these subdivisions should be.

There have been suggestions (George *et al.* 1967; Hedberg 1974) that the same principles which are used for the subdivision of Phanerozoic rocks should be extended back into the Precambrian eras and although there is much to recommend this as an intellectually ideal stand to take, in practice there is little possibility of such methods being used. There are three main reasons for suggesting that Precambrian stratigraphic units are unlikely ever to be defined in the same way, or to be of the same kind, as Phanerozoic units.

1. The methods used in Phanerozoic rocks are not suitable for largely unfossiliferous rocks (except those used for lithostratigraphic subdivision).
2. The imperfection of the sedimentary record compared with the immensity of Precambrian time makes the exercise of attempting to erect a Precambrian chronostratigraphic column of little practical use even if it were possible.
3. The majority of Precambrian rocks occur in crystalline complexes which have evolved in ways which differ so much from surface sedimentation that little correlation is possible.

Precambrian chronostratigraphy — problems

The concepts which have been developed for the subdivision of Phanerozoic rocks are suitable for subdividing sequences in which a large number of more or less contemporaneous successions can be seen and which it is possible to correlate on a world wide basis. There is almost no part of the Precambrian for which this is yet possible. There are many 'normal' stratigraphic practices and methods of definition (particularly with regard to lithostratigraphic units) which are perfectly applicable to the Precambrian and it is not suggested that normal stratigraphic practice cannot be applied to Precambrian strata. However, the fundamental basis for Phanerozoic chronostratigraphic definition and correlation (i.e. the stratotype sequence) is not applicable to the Precambrian.

Great advances are being made in stromatolite and microfloral zonation within the unmetamorphosed sedimentary sequences of the Precambrian, but it is the diversity of faunas and the use of a whole variety of different zonal schemes that enables subdivision and refinement of the Phanerozoic chronostratigraphic column. This has taken 150 years to build up and as yet only a few of the major boundaries have an agreed international status. The Precambrian biological record is likely to be very sparse in any one area even when zonal schemes have been devised, and it seems highly improbable that the same methods of correlation of a diverse flora covering a wide variety of environments will be found in Precambrian strata. It is most likely that in any one sequence some particular, favourable, geological environment will preserve what fossils there are, and it is only if this environment is repeated elsewhere that correlations can be made. Although stromatolites have been widely used to zone the Upper Proterozoic, there is no doubt that they were rather restricted in their environmental tolerances so that only inter-tidal deposits are likely to contain such material in the first place, although some are now thought to be freshwater (Siedlecka 1982). With stromatolites and phytoplankton, two major types of zonation (i.e. inter-tidal and perhaps deeper water) may eventually lead to correlation between different environments but it seems likely that some years of research effort are needed before the zonal schemes are themselves devised.

Another major difficulty is the very process of correlation itself. The Phanerozoic chronostratigraphic scale was worked out by successive sequences found to lie on top of each other and a time sequence deduced by classic Huttonian methods. This method of widespread correlation is not available for most of the Precambrian. Although sequences of sedimentary groups have been deduced for various cratonic areas for much of the Proterozoic many of these sequences have been arrived at by interpolation over miles of unexposed ground, and they have been verified largely by radiometric dating, not by classical stratigraphic methods. There is no way of correlating between cratonic sequences in, say, South Africa, with those in Russia or China. Even if they contain the same sequence of stromatolites or microflora, there can be no certainty that times of deposition in one continent equate with times of deposition on another. They might equally correlate with the periods of uplift and unconformity.

The building up of a stromatolite zonal scheme would seem to depend ultimately on accurate radiometric dating of the sequences on different cratons and this itself presents difficulties. Most frequently those sequences for which an adequate palaeontological zonation has been established are those for which radiometric data are not available. The real problem with radiometric dating for chronostratigraphic subdivision is the lack of datable material in sedimentary rocks — a problem that is not confined to the Precambrian. The use of Rb/Sr on shales, K/Ar on glauconite and U/Pb on some uraniferous organic sediments, is likely to be subject to uncertainties of interpretation. The use of intrusions which cut strata to give a younger limit to the age can never be precise even though some of the 'best' ages for major divisions in the Phanerozoic are still the ages of intrusions either cutting or unconformably overlain by the strata (i.e. give only upper or lower limits). The sheer length of Precambrian time makes this less and less satisfactory the further back in time one goes. Only when volcanic horizons within a sequence (either lavas or bentonitic ashes) are accurately dated can useful discussion of the radiometric age of strata be undertaken. Volcanic sequences are not uncommon in the Precambrian, but enormous thicknesses of sediment are found in which volcanicity is not evident and these will, by present techniques, remain very difficult to date.

It has been suggested that the magnetic reversal sequence may ultimately be extended into the Precambrian to give a means of correlation. Magnetic reversals are probably less diachronous than most palaeontological changes, but the use of magnetic reversal sequences for correlation in pre-Jurassic rocks can never be absolutely certain as one can never know whether deposition (or eruption) was sufficiently continuous to cover the whole time interval. This is not such a problem with Recent or even Tertiary non-marine sequences but becomes an almost insuperable problem in Precambrian times. The suggestion that a magnetic stratigraphy may be used to extend the chronostratigraphic column back into the Precambrian falls down completely because of the highly fragmentary nature of the Precambrian record and the inaccuracy of radiometric dating. In order to be able to correlate two reversal sequences in the Precamhrian one would have to be certain that a fairly complete sequence is present and the ages of the reversals would have to be dated (initially to erect the scale) at an accuracy greater than is likely to be possible.

As much of the Precambrian record is so poorly fossiliferous it seems unlikely that any except very gross subdivisions will be possible by biostratigraphic correlation. Even the most recent Precambrian faunas — the Ediacaran assemblages — are imperfectly known. Should we adopt the standard method of defining the base of an 'Ediacaran System', say by the appearance of the Ediacara assemblage and the definition of a stratotype (insertion of a 'golden spike') at a particular locality, we are much more likely to hit a spot which is not the chronological base of the start of the Ediacara fauna than in more recent rock sequences. There are fewer rock sequences of this age preserved and at 650–750 Ma the errors in radiometric dating are becoming 10% greater in terms of years than at the base of the Cambrian. As this is a unique occurrence (the start of many phyla) it is surely better to define the boundary as a concept than at a particular point. If the fauna is subsequently found lower in the rock sequence at another point later on, then no controversial decisions will have to be taken (either to move the golden spike or to say that some pre-Ediacaran rocks contain an Ediacara fauna). The correlation of faunas and floras in the Precambrian will, hopefully, eventually be more precise than radiometric age dating, but there is surely no justification now for quoting dates as precisely as 707 Ma (Vidal & Zoubek 1981) for the base of the Vendian.

The imperfections of the Precambrian sedimentary record

Whatever the ultimate consensus regarding correlation of the Phanerozoic rocks, it can only be applied to deposited strata. There has been general recognition that the sedimentary rock record is very fragmentary and that in most depositional environments the rock record represents less than half (possibly a very small fraction) of geological time (Sadler 1981). The immense length of Precambrian time (about six times as long as Phanerozoic) and the general paucity of sedimentary sequences in parts of the Precambrian record means that there is a very much greater likelihood of miscorrelation between Precambrian sedimentary sequences than those of the Phanerozoic. A stratigraphic subdivision of the predominantly unmetamorphosed sedimentary groups of South Africa has recently been made (Kent & Hugo 1978) and it is claimed (p. 374) that is covers the time between 3750 Ma to *c*.570 Ma with only slight hiatuses. This seems highly unlikely. Each of the main groups are represented by considerable thicknesses of sediment; the total thickness probably exceeds 65 km (Anhaeusser 1973), which is an average rate of *c.*20 m/Ma although there is a high proportion of rapidly accumulated volcanic deposits, conglomerate or coarse arenaceous sediments. Of more importance is the fact that these are highly episodic depositional environments. This rate, in any case, is highly inaccurate being based on the total thickness over the total time, whereas more than two-thirds of the thickness was deposited in the first two-fifths of the time, so that the rate for the Proterozoic is only *c.*10 m/Ma. With probably several metres being deposited in a day in volcanic and conglomeratic deposits, 'continuous' takes on a very curious meaning!

This sequence is divided into eight major groups or supergroups which are presumably related to periods of time when there was a somewhat stable sedimentary environment giving a unity to each major lithostratigraphic unit. It seems highly unlikely that only eight major changes took place in 3000 Ma during which there was 'continuous' sedimentation. However, the boundaries of major chronostratigraphic units which subdivide this time span (early Archaean to late Proterozoic) are taken at the major lithological boundaries of these lithostratigraphic divisions (Table 1, from Kent & Hugo 1978) so that the boundaries of chronostratigraphic units will be major unconformities. It is likely that these unconformities represent a considerable proportion of this time.

The early Archaean part of the sequence (the greenstone belt, Onverwacht, Fig Tree and Moodies groups) are well developed in South Africa but similar sequences occur in many other shields at about this time. The sequence from ultramafic and mafic lavas (Onverwacht Group) to argillaceous sediments (Fig Tree Group) and arenaceous sediments (Moodies Group) is one that is repeated in other greenstone belts throughout the world, but it is apparent that it represents the working of some sort of Archaean 'greenstone' geotectonic cycle that may be repeated several times in adjacent areas, so that no correlations can be made on the basis of the rock types occurring in the belts – as was done for the Sebakwian, Bulawayan and Shamvaian 'groups' in Zimbabwe where Sebakwian meant ultramafic volcanic rocks, Bulawayan meant mafic and felsic volcanic rocks and Shamvaian meant sedimentary rocks. At least three successive greenstone belt sequences with representatives of these three 'groups' are now known to be present in that area (Wilson 1973). These three greenstone belt sequences probably cover much the same period of time as the Swaziland and Witwatersrand Supergroups of the South African craton and there are no means (other than radiometric dating) of deciding which of the Zimbabwian greenstone belt sequences correlates with the Swaziland greenstone belt, if indeed any are exact equivalents. It could well be that an individual greenstone belt sequence is a unique geotectonic occurrence controlled by rather localized forces within the mantle beneath the particular greenstone basin. Thus no correlation of greenstone belt sequences from one area to another is yet possible except by radiometric

TABLE 1. Stratigraphic subdivision of South African Precambrian (after Kent & Hugo 1978)

Approx. age, Ma.	Chronostratigraphic Unit: Era	Lithostratigraphic Unit: Group	Lithostratigraphic Unit: Supergroup/ Sequence	Thickness (from Anhaeusser 1973)
	Palaeozoic	Nama/Malmesbury		
	Namibian	Gariep, Nosib	Damara	
1080	Mogolian	Koras Waterberg/Soutpansherg		6½ km 5 km sed 1½ km volc
2070	Vaalian	Rooiberg		
		Oilfantshoek Pretoria/Postmasburg Chuniespoort/Campbell/ Griquatown Wolkberg	Transvaal/ Griqualand West	11 km 9 km sed 2 km volc
2630	Randian	Pniel Platberg Klipriviersberg	Ventersdorp	5 km 1 km sed 4 km volc
		Central Rand West Rand	Witwatersrand	11 km 9 km sed 2 km volc
		Dominion Limpopo (Beit Bridge)		
2800				
2900	Swazian	Pongola		10 km 5 km sed 5 km volc
		Moodies Figtree Onverwacht	Swaziland	21 km 5 km sed 16 km volc
3750				

dating, although it may be possible to use the presence of greenstone belts to denote a major subdivision, like the Archaean. Workers in Australia (Hickman 1981) have shown that normal lithostratigraphic methods can be applied to the correlation of closely adjacent greenstone belts and they suggest that over large parts of the Pilbara block of Western Australia there was in fact only one sequence of depositional events that may be correlated laterally. However, even they would not propose that this is the correlative of the greenstone belts of the Yilgarn block where Sm-Nd dates give ages of 3050 ± 100 Ma, 2980 ± 120 Ma and 2780 ± 70 Ma, suggestive of a progressive crustal age trend across the block (Fletcher *et al.* 1984). This compares with an age of 3560 ± 32 Ma for the Pilbara block (Hickman 1981). The later Archaean sequences of South Africa (e.g. the Witwatersrand Supergroup) is a cratonic fluvial sequence whose environment of deposition is quite unlike much late Archaean sedimentation elsewhere in the world and thus offers little possibility of correlation on any basis other than very broad categories of primitive microflora.

The whole of the South African Proterozoic sequence is cratonic and the controls over the various periods of sedimentation and volcanism are unknown hut could conceivably relate to orogenic episodes in adjacent terrains. Whether any of these unconformities or sedimentation episodes correspond to such external events and whether the rock record in these orogenic belts can be matched with any of the events taking place on the craton is unknown but it seems likely that meaningful correlations except by radiometric age dating are a long way off.

Considering the comparison with other shield areas made by Kent & Hugo (1978) (even they would not regard them as correlations) there seems little doubt that in each shield area one can erect a local, self-consistent, stratigraphic scale, but this will be only a lithostratigraphic grouping. It is perfectly conceivable that the deposition of the five major Proterozoic supracrustal groups of the Russian shield in Karelia could represent entirely different periods of time to those seen in South Africa or Australia (Wright 1980). Indeed if the unconformities represent the majority of geological time this is likely to be the case. It might be argued that this may make the construction of a complete Precambrian stratigraphic column *more* likely since more of the time will be represented by rock (although on different continental blocks). However, it also renders impossible correlation by characters found within the rocks, and it is likely that radiometric dating will remain the only means of comparison for many years.

Lithostratigraphy as a basis for chronostratigraphy

The use of lithostratigraphic units in the erection of chronostratigraphic subdivisions of the Precambrian is perhaps the most fundamental of Kent & Hugo's (1978) suggestions. It is, at first sight, reasonable and unobjectionable, especially for the erection of a local stratigraphic time-scale. Clearly the Witwatersrand sediments were laid down over a finite period of time and were preceded by a Dominion Group of different facies and succeeded by the largely volcanic Ventersdorp Supergroup. This succession may therefore be regarded as a definite time sequence in that local area. However, if the problem is considered in more detail it is apparent that the designation of these units as chronostratigraphic units is unjustified.

This part of the South African stratigraphic column is made up of an enormous thickness of sedimentary and volcanic rocks deposited over really quite short a period of late Archaean time. Due largely to its economic importance very detailed lithostratigraphic sequences have been worked out which have enabled detailed correlations to be confidently made over quite wide areas. Since there are well recognized lateral facies changes throughout the sequence it is tempting to regard the major units, above formation level, as having some sort of chronostratigraphic meaning. However, it must be recognized that the major changes that enable the Hospital Hill Subgroup to be distinguished from the Government Reef Subgroup are changes in facies and although there is a consistent sequence along the length of the basin of a lower unit of alternating siliceous and argillaceous beds and an upper unit which is quartzitic and conglomeratic, there is, of course, no means of knowing whether this is a chronological distinction. It is a distinction based upon a change in facies and all such changes are likely to be diachronous; the fact that the lower unit is always of one facies and the upper unit is always contrasting says nothing about their time relationships. No doubt the workers on the Witwatersrand Supergroup are now fairly confident that detailed correlations can be made and certain marker beds are widely used to correlate the various formations more securely than by simple methods of lateral correlation and the law of superposition. However, there is no disputing the fact that the correlations are all lithostratigraphic and even the most widespread sedimentary marker bed may itself be diachronous.

Given that the sequences of formations that are built up to form subgroups cannot be regarded as time bounded sequences, then there is no justification for suggesting that the larger units — the Witwatersrand or even the whole of the Dominion, Witwatersrand and Ventersdrop, may be somehow regarded as representing chronostratigraphic subdivisions. This is particularly the case where the major groups differ lithologically (as in this case) as they are obviously being recognized stratigraphically by lithological criteria and the period during which a particular rock type was laid down would have varied as the environments of deposition varied (and in the case of predominantly volcanic groups as the overall geotectonic environment varied), not necessarily at the same rate or in the same manner from area to area. As the Ventersdorp cannot be resolved geochronologically from the Dominion Reef (Cahen, Snelling *et al.* 1984), correlation with other volcanic/clastic sequences such as the Zoetlief Group of Cape Province (Cahen, Snelling *et al.* 1984) is thus purely lithologically based.

One has only to consider the problem of correlating this sequence onto other cratons to realize the non-chronological nature of the boundaries. In fact very few areas of the world show any unmetamorphosed sedimentary or volcanic sequences of the late Archaean, although many areas (e.g. the Lewisian of Scotland) were very active orogenic zones at that time with complicated sequences of shelf sedimentation, plutonic intrusion, multiple deformation and metamorphism, culminating in a crust stabilizing event at about the end of this period (*c.*2650 Ma).

If Randian and Swazian are to be regarded as world wide chronostratigraphic units then the sedimentation episode of the Scourian Cycle is probably Swazian and the rest of the cycle of activity is Randian. The author does not think anyone would wish to make such a distinction and there is no logical way of correlating any of the events which went to make up the Scourian complex with those taking place in South Africa on a craton which had stabilized some 200 Ma earlier, except by radiometric dating. Even if volcanic sequences were found on other cratons similar in nature to the Ventersdorp there is no compelling reason why they should be correlated with that supergroup, even if they followed a sequence of cratonic basin quartzites. All continental craton sedimentation will give rise to rock sequences that are broadly similar in character, and if they go through the same geotectonic processes of faulting and uplift followed by gradual degradation of the marginal mountains they are likely to contain similar sequences of deposits. Should volcanicity occur on this continental mass it is likely to be of similar type if due to the same overall geotectonic controls, but on other cratons these may be operating at quite different times.

Thus the subdivision of Randian into three units, the Dominion, Witwatersrand and Ventersdorp is purely a local division based upon the local circumstances that prevailed over about 100 Ma (Cahen, Snelling *et al.* 1984) to give three distinct periods of sedimentary and volcanic deposition on a cratonic crust. (The Limpopo (Beit Bridge) Group is distinguished by being the sedimentary unit of a gneissose complex in a neighbouring mobile belt and does not, one would suppose, bear any particular stratigraphic relationship to these other groups. The complex history of that belt (Cahen & Snelling 1984) means that the age of the sedimentary protolithic succession is of unknown age so that this member of Kent & Hugo's Randian has a very doubtful stratigraphic position.)

The clearest analogy with Phanerozoic problems of correlation is with the unfossiliferous desert sequences of the New Red Sandstone of Europe. Here there are dozens of well documented lithostratigraphic sequences with similar successions in different areas but the assignment of almost any unfossiliferous horizon to a chronostratigraphic stage is pure conjecture. Often there are long sequences whose age, whether late Carboniferous, Permian or Trias sic is simply unknown. In the Permian and Triassic correlation charts (Smith *et al.* 1974; Warrington *et al.* 1980) there are almost no 'horizontal' lines indicating that particular formations belong with any certainty to particular stages. Precambrian lithostratigraphic units are, in fact, even more uncertain, as even in the New Red Sandstone there are occasional fossiliferous horizons which give a rough indication of age, and very rare marine horizons which can be assigned to a chronostratigraphic stage.

The apparent ease with which complex lithostratigraphic sequences can be erected and local correlations made, which are perfectly valid and useful, had led to the assumption that widespread 'markers' of lithological origin may be used to correlate throughout the Precambrian between one continental block and another. This method has been brought to its highest art by Choubert & Faure-Muret (1980) who have used lithostratigraphic markers to correlate late Precambrian successions over most of North-west Europe, North Africa and Western America. The word 'art' has purposely been used in the previous sentence because the author believes it to be unscientific. By this method they arrived at the conclusion that the Charnian of central England is about 1000 Ma (Upper Middle Proterozoic) and that the Conception Group of Newfoundland is also of the same age (1300–1000 Ma), (Choubert & Faure-Muret 1980, p. 188 and p. 132) despite the fact that both sequences contain *Charnia* and other members of the Ediacara fauna. This is universally accepted as a very late Precambrian fauna (less than 700 Ma, Cloud & Glaessner 1982) and the radiometric age data from both England and Newfoundland are all perfectly consistent with this interpretation. A wholly ficticious 'rejuvenation' of isotopic dates was resorted to, a process which they invented to explain the wealth of perfectly adequate geochronological data which runs counter to their perceived correlations (Choubert & Faure-Muret 1980, p. 211). This 'rejunvenation' mechanism has, as far as the author is aware, been ignored by all working geochronologists but it must be pointed out that it is completely without any scientific basis and needs to be dismissed along with their correlations. The fact that their correlations are also apparently based on microfloral assemblages (Timofeyev *et al.* 1980) is discussed below. Their general method of correlation is 'based on facies comparisons and on stratigraphic successions. For such comparisons to be valid, the observations must be made in the field by the same person or persons, as the geological language is not sufficiently specific to describe the nuances of these facies' (Choubert & Faure-Muret 1980, p. 212). There is no doubt that the reliance upon lithostratigraphic comparison and dismissal or disregard of accurate radiometric data has led to a highly confusing set of correlations, some of which may be perfectly valid, but since others are clearly wrong it throws in doubt their whole methodology. This has resulted from a complete misunderstanding of the role of lithostratigraphic correlation in the Precambrian. Marker horizons may exist locally, but without independent corroborative data, either biological or geochronological, the equivalence of two similar occurrences of the same type of unit, or the same sequence of units, must not be assumed. This is particularly true when the facies being correlated are part of geotectonic cycles which may repeat themselves many times within a short period. Cycles of activity which take 20–30 Ma (i.e. many parts of orogenic cycles) may be easily miscorrelated in the immensity of the hundreds of millions of years of the Proterozoic.

Lithostratigraphic usage

Kent & Hugo's paper (1978) illustrates another facet of Precambrian stratigraphic usage which is overdue for rationalization. Stratigraphic nomenclature of rock groups in the Precambrian has always been bedevilled by a plethora of names for the major units, viz: system, group, sequence, succession, assemblage and series which have often been used indiscriminately over a long period. Despite making a commendable attempt to modernize the South African Precambrian usage Kent & Hugo (1978) still retain '*sequence*' for some lithostratigraphic units and '*group*' or '*supergroup*' for others.

The authority for this usage is that the term sequence was suggested as an *informal chronostratigraphic* unit for a 'great aggregation of strata lying between major regional unconformities and frequently comprising rocks of several systems' (Hedberg 1971). This is applied by Kent & Hugo to the Karroo, which exactly fits with the above suggestion, but in the Precambrian there seems no way of distinguishing a sequence from a supergroup since systems, or any other chronostratigraphic subdivision, do not exist. It could, perhaps, be argued that an informal term like sequence is appropriate for *all* major Precambrian subdivisions but in that case each of the supergroups and major groups of the South African column should be referred to informal chronostratigraphic sequences, and there would seem to be no greater validity for their newly erected eras than for such sequences — both are simply terms for 'the time during which a particular supergroup was laid down'.

There is no doubt that much of the older confusion and complexity is due to the fact that many named Precambrian rock sequences are metamorphosed, and workers were loath to commit themselves to a standard lithostratigraphic nomenclature. It is proposed below to have a separate nomenclature for those rock bodies which are so deformed and metamorphosed that a lithostratigraphic name is inappropriate, but the corollary to this is that *all sequences which are thought to be in undisturbed stratigraphic continuity should be named according to the standard lithostratigraphic nomenclature*.

The subsequent equation of a Precambrian *era* as the time during which a particular supergroup was deposited has no particular merit; it can only be a local chronostratigraphic unit and is not based on the same rationale as Phanerozoic eras. There is some merit in the informal use of sequence and this would seem a logical way forward, but after its first appearance as a recommendation (Hedberg 1971) it has not been mentioned in subsequent ISSC publications (e.g. Hedberg 1976) and the term seems to have dropped out of favour.

Precambrian biota and Precambrian chronostratigraphy

Recent IGCP projects have been attempting to correlate Precambrian successions, but they have not been noticeably successful. The Middle and Late Proterozoic holds out most hope of correlations being effected by biostratigraphic means (Bertrand-Sarfati & Walter 1981; Vidal & Zoubek 1981) but review volumes published recently (Precambrian Research 1981 v. 15, No. 1 and No.'s 3–4; v. 18, No. 4) reveal no means of correlating the detailed successions described for China, USSR, Australia, Canada and Europe. However, these papers indicate that this is a rapidly developing field of investigation and that rigorous methods, both taxonomic and stratigraphic, are being applied, holding out hope of a marked improvement in reliability in the future.

The slow rate of change within the Precambrian biota may mean that eventually it will be possible to define three or four major changes in the organisms to subdivide Precambrian time into major subdivisions (Eras) but there is no reason to suppose that these will approximate to the ages arbitrarily decided for the Early, Middle and Late Proterozoic boundaries (Sims 1980). A biostratigraphic sequence for the Precambrian must be largely built up by reference to a chronometric scale, as many lithological sequences with biological material have unknown stratigraphic relationships with other lithological sequences. However, there are very few lithostratigraphic sequences which are accurately dated by radiometric means so that the relationships between known biozones are not always clear. The dangers of uncontrolled correlation by microflora is well illustrated by the highly confused results obtained by Choubert & Faure-Muret (1980). Although purporting to use acritarchs and other microflora to correlate between the late Precambrian sequences of north-west Europe, their correlations are so consistently at variance with high quality radiometric data that it can only be assumed that the acritarch zonal scheme which they are using is based upon wrongly dated or poorly dated sequences.

The development of microfloral zonal schemes for the Proterozoic will have to be much more securely grounded on sequences of well dated rocks in undeniable upward sequences if they are to be of use. Unfortunately such sequences are relatively rare. Despite the apparent simplicity of many Precambrian successions and the agreement about the relative age of quite large groups of rock these are most frequently entirely inferential. In the British Precambrian it would be true to say that the relative ages of the Torridonian, Moinian, Dalradian, Mona, Longmyndian and Charnian (Wright 1980) are all unknown, yet their relative positions in the late Precambrian column are all widely accepted. This is almost entirely based now on radiometric data, since they are all separate structural entitites and their actual stratigraphic relations are unknown. Because they are largely from different geotectonic environments it is likely that lithological correlation will never solve their interrelationships. For this reason collecting microfossils from these groups will be unhelpful for the purposes of making a zonal scheme, which should be attempted only on sequences which can be seen in structural continuity.

The paucity of palaeontological data from the Precambrian is partly due to a lack of micropalaeontological study — there is no doubt that a far wider coverage of the Precambrian stratigraphic column will be obtained by micropalaeontologists in the future — but there is also a real lack of data; for example, the number of outcrops of rocks containing the Ediacara faunas is exceedingly small and the problem of finding an '*ideal*' place to make a stratotype for the base of the Vendian or Ediacaran (see below) is correspondingly greater that at the base of a Phanerozoic system where there are literally hundreds of outcrops worldwide where information relating to the base of the system may be found. There may be many more rocks with Ediacaran faunas waiting to be discovered, of course. The relatively recent discovery of an Ediacara fauna in Newfoundland (Anderson & Misra 1968) in a well-studied area with an enormous thickness of late Proterozoic unmetamorphosed sediments, indicates that preservation combined with the existence of an environment favourable for the development of the fauna in the first place, limits very considerably the number of places where an Ediacara fauna is likely to be discovered — plus of course the fact that they are trace fossils rather than body fossils.

Cloud & Glaessner (1982) have, however, suggested that a new system (the Ediacaran) be instituted with its standard reference section the Ediacara Hills of South Australia. However, they do define the new system as 'that geological interval characterized by the soft-bodied, macroscopic marine invertebrates of the Ediacara fauna and allied forms', using a concept for the definition of the system rather than a stratotype sequence. They incidentally reject the term Vendian (which has previously been favoured as a candidate for the latest Precambrian system) on the grounds that it is widely considered to include a late Proterozoic glacial episode which Cloud & Glaessner regard as being below the earliest appearance of metazoans and would thus conflict with their 'concept' definition of the sysem. There would seem to be good grounds for approving of this definition since the major problem in the definition of the Vendian has always heen what character to use to define its base. Since the Vendian has been regarded as including both the metazoan fauna and the tills beneath, in non-glaciogenic sequences or long sequences that contain several glacial episodes the precise position of the base is obscure.

Even more difficulties arise with the use of Riphean. Although the term Riphean has been used in the USSR. for many years as a chronostratigraphic term, its usage has changed with time (sometimes including Vendian as a subdivision, at other being the pre-Vendian, Upper Proterozoic) hut it covers a long period of the Proterozoic (*c.* 1600 – 600) and lower, middle and upper divisions have been suggested. However, no precise definition of these boundaries has been produced which has won widespread acceptance, and there seems little possibility of getting agreement to a chronostratigraphic term which has had such a chequered and confusing previous usage and which cannot be precisely defined.

Certainly the sequence of stromatolites detailed by, for example, Semikhatov (1980), can only be used when the type sequences in which they occur have been accurately dated (and the data on which the dates are based are quoted in an acceptable scientific form).

Thus any advance in chronostratigraphic terminology of the Precambrian must be preceded by international agreement on the method of defining the base of each unit, and, if trace fossils or microflora are to be used, a considerable program of investigation will have to be undertaken and much more extensive dating of the bedded sequences that contain the flora will be necessary, to enable correlation to be based on an outside scale.

Crystalline complexes in the Precambrian

The major problem with trying to subdivide Precambrian rocks, however, lies in the predominance of crystalline gneissose complexes as representatives of mappable Precambrian. Any subdivision of the Precambrian which ignored these rocks will be of limited application. It is not simply a question of part of the rock record being metamorphosed — there is the well-documented dichotomy between the high grade granulite facies shelf-volcanic sequences of the Archaean and the low grade greenstone belt sequences; they seem to have formed in two entirely separate geotectonic environments and it seems unlikely that correlation will be possible between them by tracing undeformed sequences into the 'mobile belt' as may be possible with more recent orogenies. Similar problems will beset correlation between the cratonic undeformed sequences of most of the Proterozoic with the rocks forming in the Proterozoic orogenic belts many of which contain no recognizable Proterozoic sediments (Wright 1980).

These gneissose belts do not lend themselves to normal stratigraphic nomenclature, although increasingly they are being thought of as bodies of rock which have gone through a series of events which can be ordered in time. There is thus a 'stratigraphy', or a sequence of events, to be determined. These events bear little relationship to classic stratigraphic divisions which are essentially periods of sedimentation. In gneissose complexes it is other events, intrusion, deformation and metamorphism, which predominate in the history. One consequence is that the rock units are not simply a result of one process and it may not be possible to recognize lithological units separately from their tectonic identity. In these cases it is suggested that a lithotectonic heirarchy of terms be used in mapping. The use of such terms would imply that a lithostratigraphic term is not applicable because of too strong a deformation or metamorphism and that no time sequence is suggested by the superposition of one lithotectonic unit on top of another. The terms *division* and *assemblage* are suggested roughly equivalent to *group* and *supergroup*. No term equivalent to *formation* has been put forward although *unit* is possible. It is likely that rock names such as *pelite* or *quartzite* would be generally used at the 'formation' level, although terms such as 'gneiss' are not recommended (see below).

Dividing gneissose belts in terms of time has also developed traditional terms and it is suggested that these are also formalized as *cycle, episode* and *phase*. These terms would refer to the development of a metamorphic complex over a period of time in a particular place. Correlation of adjacent gneiss complexes with a given cycle depend not only on the time during which the cycle developed but also whether it was likely to have been in structural continuity with the named cycle. The concept of orogenic or tectonomagmatic cycles has proved to be a fruitful one in the investigation of gneissose complexes and it is felt that a more standarized way of defining them may bring a clearer understanding of the relationships. It must be explained that such cycles are *not* seen as a step towards erecting a Precambrian chronostratigraphic column — they may be related to one but their names should never be used as subdivisions of Precambrian time since they are by definition of local extent and the time of the climax of such cycles of activity would not be expected to have any but a local significance and may be diachronous along the length of a gneissose complex.

The 'chronostratigraphic' scale of the Canadian Geological Survey (Douglas 1980; Stockwell 1982) is apparently defined in terms of orogenic cycles but the end of each cycle is recognized by a major unconformity at the base of a subsequent lithostratigraphic group, which may then become involved in the next orogenic cycle. This method of subdividing Precambrian complexes has always been used in Canada (and many other parts of the world) and may be regarded as one of the classic methods by which our knowledge of Precambrian rocks has been advanced. Harland (1983) suggests that the base of these lithostratigraphic units could become the marker horizons for the erection of a Proterozoic chronostratigraphic scale. The advantages and disadvantages of this development are discussed below, but the proposals in the present paper are specifically against using orogenic cycles as a basis for subdividing Precambrian time. Although gneissose complexes can only be realistically subdivided by recognizing the effects of different orogenic cycles within them, it is hoped that these cycles can be regarded as structures developing at a particular time and place which it will be possible to relate to an entirely independent chronostratigraphic scale.

Chronotectonic Terms

The *cycle* is a concept which is limited in both time and space. In the present state of our knowledge of the workings of the

Earth it seems likely that what have previously been called cycles are sedimentary, metamorphic, intrusive and tectonic events taking place near a continental margin during subduction and sometimes collision. This can only take place (considering plate-tectonic processes) in continuity along one continental margin at a time, so that one cycle is unlikely to be world wide in extent, e.g. the Alpine Cycle could be defined to include the Alpine chain through the Himalayas and beyond but would not include the Rockies and the Andes (although they may be developing at the same time they are a different structural entity).

A cycle is a sequence of metamorphic, igneous and tectonic events related to each other in time and space, perhaps preceded and succeeded by periods of sedimentation (e.g. flysch and molasse), allied in their evolution to the tectonometamorphic event. Problems may arise in naming cycles where two continental margins of contrasted bistory collide. In Britain the Late Precambrian to Early Palaeozoic history illustrates this point well (it also has the advantage of biostratigraphic control in the later events). To the south-east a Celtic Cycle was active from about 750 Ma to Early Cambrian, the climax of the orogenic activity probably being about 650–600 Ma, sedimentation of the Caledonian Cycle beginning with Lower Cambrian rocks unconformably overlying the Celtic volcanic arc. In the north-west the Grampian Cycle was active from about 700 Ma or earlier to Early Ordovician, the climax of the orogenic activity being about 500 Ma and the sedimentation of the Caledonian (or Erian) Cycle beginning in Early Ordovician time. In both areas the Caledonian Cycle reached its climax in Late Silurian/Early Devonian time and was followed by Old Red Sandstone molasse as the last part of the cycle. If sedimentation episodes are to be included in the Caledonian Cycle (and this has been the usage by most workers so far), then the length of the cycle would be different on either side of the Caledonian suture. As continental margins by their nature have independent histories, this is likely to be a general feature of all orogenic belts caused by continental or microcontinental collision.

In any one area the whole of Precambrian time is not likely to be represented by active cycles (i.e. the beginning of one cycle does not necessarily follow immediately after the end of the preceding cycle) so both the beginning and the end of the cycle should be dated as accurately as possible. Although closely connected with orogeny the cycle is not defined as an orogeny (to avoid possible changes in our knowledge of the mechanisms of Precambrian crustal genesis) but the climax of the cycle may be called, for example, the Grenville Orogeny. If only an orogenic episode is recognized then the two become synonymous (cf. Bowes 1968, 1980).

The *cycle* may be subdivided into *episodes* and these episodes into *phases* and possibly phases into *events*. If the early sedimentation of the Laxfordian Cycle is now represented by the Loch Maree Assemblage of supracrustal schists this could be called the Loch Maree Episode of the Laxfordian Cycle (meaning a sedimentary episode). The terms episode and phase, used with any of the names selected for metamorphism, plutons, folds etc., imply a time (or at least a position in a sequence) as well as the specific character of the event.

Dating

More refined analytical techniques have in recent years meant that much more accurate dating of Precambrian complexes has become possible. With the closer co-operation between the geologist and the geochronologist in the selection of appropriate rocks for narrowing down specific events, it has become possible to differentiate between different metamorphic phases within one cycle (Hopgood *et al.* 1983). It is still likely that an orogenic episode may normally only be assigned to a single date (within errors) although early and late phases may be sufficiently distinct in time and have sufficiently distinctive marker events which are radiometrically datable to enable the span of an episode to be dated within one area. This has only recently become possible and older discussions (about the precise meaning of Grenvillian for example) were so bedevilled by the lack of both detailed gneissose 'stratigraphy' and lack of refined geochronological data that they are now of only historical interest.

It would not, however, seem appropriate to define a cycle *rigidly* in terms of ages as these may so easily be subjected to later revision on geological as well as geochronological grounds.

Definition

A *cycle* should thus be defined as a cognate sequence of geological events in a specific (fairly broad) area with an indication of the likely extent laterally and with a type area suggested where the development of the cycle may best be recognized. This should include the rocks which have been dated and although the dates are a significant part of the definition it is the series of events that define the cycle and if later geochronological work suggests that these events all took place at a different time then the age of the cycle must be changed (or the cycle might be recognized as belonging to a cycle previously defined elsewhere). There is no doubt that 'cycles' will be less secure stratigraphic entities than traditional stratigraphic units and much more likely to later revision or rejection.

Orogenic cycles have been widely recognized for a number of years and it is suggested that as far as possible a chronotectonic cycle should be analogous to these, i.e. sedimentary, tectonic and morphogenetic episodes should be apparent. The episodes making up the cycle should thus be defined in terms of a specific character (sedimentation, folding, metamorphism) at a specific time, and a type area for their development be stated.

Lithotectonic terms

Although metamorphic complexes may have several different mappable boundaries drawn through them, none of these are recognizably successive, time-controlled zones. Each type of boundary (tectonic shear zone, metamorphic mineral zones, boundaries between schistose units) may be representative of distinct events in the formation of the complex, but they do not represent on the map a 'time' plane, like the bedding plane of normal lithostratigraphy. The mappable units of metamorphic complexes should thus not be defined in normal lithostratigraphic terms. It is axiomatic that workers in gneissose complexes, although mapping shear 'zones' and metamorphic 'zones' have never regarded these lines as stratigraphic in any sense although the *event* may have a time significance. (It seems necessary to state this here in view of replies given to various questionnaires on Precambrian stratigraphy.)

Unit is suggested for a lithological entity that is mappable and cannot be assigned to the normal lithostratigraphic heirarchy of formation, group and supergroup due to tectonism, metamorphism or both. Individual units may be called by a lithological name, psammite, pelite, marble, metadolerite etc., with or without the term Unit e.g. Dorlin Pelite Unit. A unit does not have to be of uniform lithology (although it would normally be so) but should certainly be a tectonic unity (i.e. should not contain any obvious major tectonic discontinuity within it). Since the origin of metamorphosed rocks is not necessarily sedimentary, both igneous and sedimentary precursors are possible and the term '*Unit*' does not imply anything about its origin.

One or more units may form a *Division*. The Units may or may not be recognized to be in stratigraphic continuity although they would normally not be so (otherwise lithostratigraphic terms could be employed). If the rocks are above staurolite grade metamorphism it seems unlikely that such continuity would be provable and lithotectonic terms would seem most appropriate. Despite beliefs to the contrary, any group of rocks at high metamorphic grade may contain unseen thrusts or isoclinal folds which repeat, invert or otherwise distort the lithostratigraphic sequence. As high grade metamorphic rocks are the hallmark of an orogenic belt and as such structural complications are also typically developed at an early stage in many belts, to suggest that granulite facies rocks (for example) preserve an undisturbed stratigraphic sequence is fatuous. They may, of course, do so, but it is impossible to determine this and a lithotectonic terminology is thus to be preferred. Where banding is probably not depositional (i.e. most gneisses) lithotectonic terms are imperative as lithostratigraphic terms are entirely inappropriate. Many gneisses are regarded (often quite wrongly!) as of igneous origin and the term '*division*' may equally be applied to rocks with an igneous as well as a sedimentary origin.

Several divisions, if associated with each other in a tectonic sequence, may be termed an *Assemblage*. The several Divisions of an Assemblage are normally *not* in provable stratigraphic conformity and may indeed be lateral chronostratigraphic equivalents.

Terms for gneiss precursors

This represents not only one of the most contentious issues but one which, if it can be agreed, will greatly clarify discussions about metamorphic complexes. For most gneisses there are several events in its evolution which are of interest; the time of original formation (e.g. as a sedimentary sequence or an igneous plutonic complex), the time of metamorphism and any subsequent metamorphism, and the time of uplift and emplacement into the structural position within the orogenic belt in which it is now found. Different methods of radiometric dating may reveal the first of these (particularly for igneous precursors), the time of metamorphism (though usually only the last major metamorphic phase) and the time of uplift. During discussions of such determinations a set of terms for the sedimentary precursor and the igneous precursor components of a gneiss complex would be very useful. One possibility is to use normal lithostratigraphic names, such units simply being called, e.g. the Blacktown Group (for the sedimentary precursors) or the Greenbay Granite (for the igneous precursors). This would mean erecting formal stratigraphic names for rock groups which do not, in fact, now exist. I think this would be generally regarded as inadvisable. There seems to be no case for having a hierarchy of terms for two such nebulous entities so the terms *Protolithic Succession* and *Protolithic Intrusion* are proposed to signify those sedimentary and igneous rocks whose deposition or emplacement represents the earliest event recognizable in a crystalline complex. Where the type of precursor is unknown, *Protolith* may be used unqualified. It seems logical to suggest that the geographic name assigned to the Protolith should be the same as that used for the Assemblage or Division mapped. As an hierarchy is not suggested the precursors to the Moinian Assemblage of Scotland, which is divided into the Morar, Glenfinnan and Loch Eil divisions would be designated the Moine Protolithic Succession, the Morar Protolithic Succession, the Glenfinnan Protolithic Succession and the Loch Eil Protolithic Succession. The Ardgour Granitic Gneiss, which would be termed the Ardgour Division in this nomenclature, would have been derived from the Ardgour Protolithic Intrusion and has been converted into a gneiss during an (unnamed) metamorphic episode.

Usage of the term gneiss

It is proposed that metamorphic rock names with textural implications, e.g. gneiss, schist, mylonite, should be taken to indicate, tectonostratigraphically, the time of formation of the *metamorphic* rock, i.e. the time of metamorphism and deformation. Thus a Laxfordian Gneiss is a rock formed during the Laxfordian Cycle by gneissification of pre-existing rocks. The gneissification may have affected older basement as well as newer supracrustal sediments and igneous intrusives. All would be termed Laxfordian Gneisses if now gneissose. In the same way Caledonian Mylonites could include mylonized Archaean, Proterozoic and Palaeozoic rocks. The name then says nothing of the age of origin of its protolith, simply the age at which it became a gneiss or mylonite.

Naturally, there may be cases of polymetamorphism which do not include a second period of gneissification. Thus the Scourian Gneisses may be subjected to a Laxfordian Metamorphism which does not destroy the Scourian gneissosity (they are then still Scourian Gneiss, although the retrogression may well result in a Laxfordian age being obtained). In polymetamorphic terrain it is much too simplistic to ask 'What is the age of this rock?', or 'Is this a Laxfordian rock?'. It is necessary to learn to recognize the multiplicity of features that indicate a complicated history and try to deduce the sequence of these and ultimately correlate these features with a chronologic scale and to use terms for their nomenclature which are clear and logical. Frequently usage has been to use the term gneiss for a whole complex, e.g. Annagh Gneiss Complex and to designate individual rock types within it as gneisses also, e.g. grey gneiss, but sometimes individual mappable gneiss units have been designated by a place name. The usage recommended here would be to call the whole complex the Annagh Division and the individual mappable unit by a place name and the term unit, e.g. Mullet Unit. The period of metamorphism which gives rise to the gneiss would generally be named after the cycle as would the gneiss so formed, so if the Annagh Division was formed during the Mayo Cycle the gneissosity would be produced by the Mayo Metamorphism to give rise to the Mayo Gneiss.

The naming of metamorphosed igneous intrusions

In many crystalline complexes igneous rocks are intruded at many times during their development. Those formed early in the sequence of events become strongly deformed and may be converted to gneisses. The terminology of these is dealt with above. Those intruded later may still he metamorphosed and their mineralogy changed to a metamorphic paragenesis either throughout the intrusion or in part, yet may still retain their cross-cutting relationships. It is clearly illogical and unhelpful to refer to these by their metamorphic rock name (e.g. amphibolite) as the stratigraphically important feature is their igneous intrusive origin. They should therefore be named in the same way as post-tectonic or anorogenic intrusions, e.g. Oxford Dolerite or Oxford Dyke. If the former is unacceptable (because it is not now a dolerite) then metamorphosed Oxford Dolerite should be used, as Oxford Amphibolite would not indicate (necessarily) its igneous and intrusive nature, and Oxford Amphibolite Dyke may be regarded as incorrect, as it was never a dyke of amphibolite but a dolerite dyke that has been converted to amphibolite.

Precambrian stratigraphic subdivision — conclusions

It has been suggested above that normal stratigraphic procedures, i.e. the erection of stratotype sequences, would be unworkable if they were the means of *defining* a Precambrian stratigraphic column. This is due to:

1. A paucity of flora and fauna, so that only very broad biostratigraphic zones are now possible. In any case the fossils are found in a very small proportion of Precambrian rocks.
2. A lack of any other widespread time markers of sufficient accuracy to be used. Various markers have been proposed — tillites, the incoming of red beds, changes in the earth's magnetic field — but at the present time and for the forseeable future it will not be possible to use these as accurate markers.
3. The errors in radiometric dating are often so large in the Archaean and Early Proterozoic (of the order of 30 Ma) that whole cycles of earth activity are likely to be encompassed within the limits of error of a single age determination. The accuracy of measurement has improved with time and with new methods this may be brought down to better than five Ma (U/Pb and Sm/Nd methods are giving dates of this order of accuracy at 2000 Ma at the moment), but the geological errors will always be more difficult to quantify in the complex geological situations of many Precambrian outcrops.
4. The immensity of Archaean and Proterozoic time means that the methods used for Phanerozoic stratigraphy hecome unworkable when applied to rock sequences that represent only a small part of the available chronologic record and do not contain sufficient distinctive characters to allow multiple (or any) correlations between different sequences.

The use of radiometric dates in the definition of chronostratigraphic divisions of the Precambrian is 'unjustifiable' according to most stratigraphers (e.g. Holland 1978, p. 89) and the unconformity is the 'worst possible boundary' for a chronostratigraphic unit (Hedberg 1976, p. 84). Thus the boundary of the Archaean and Proterozoic, agreed recently as 2500 Ma (James 1978) and the proposed local South African chronostratigraphic units (Kent & Hugh 1978), which have met with approval as a *local* chronostratigraphic standard, are not in accord with standard stratigraphic practice.

The proposals to define the major boundaries of the Archaean and Proterozoic by reference to a particular time (James 1978; Sims 1980) are seen simply as a way out of the impasse confronting Precambrian stratigraphers. Because of the lack of diagnostic characters in Precambrian sequences there is no rock-based method available for correlating between sequences (which is the entire basis of Phanerozoic stratigraphy) and radiometric dating is the only available method of correlation. Thus, until certain definable characters have been recognized in the Precambrian rocks, which it is possible either to date or to correlate, then the definition of the major boundaries as a particular date is the only way of having defined terms of world wide use which are immediately comprehensible. It is, after all, only using as the basis of the chronostratigraphic units the only reliable method by which correlations can at present be made.

It is suggested in this paper that eventually the major boundaries should be defined by the character of an event. For example, it has recently been suggested that an *Ediacaran System* be defined by the appearance of the Ediacara fauna of soft bodied animals (Cloud & Glaessner 1982). Such an event is likely to have been rapid compared with errors in radiometric age dating and to have spread all over the world relatively rapidly.

The event which separates the Archaean and Proterozoic is less simple to define, but there is a major body of opinion which regards the processes of crustal evolution of the Archaean as differing in many respects from those found in the Proterozoic. The principal difference is the occurrence of extensive linear orogenic belts. This is such a gross distinction that it will be observed only when large areas of Precambrian crust are considered and it cannot be equated in any way with the boundary-stratotype form of definition. Most workers are agreed that there is a difference between Archaean and Proterozoic even though Archaean and Proterozoic have never been satisfactorily defined. Others will seriously contend that the boundary between the two is (a) a time band rather than a time plane or (b) diachronous — the change taking place later in one area than in another. As there has never been any attempt to define what the houndary is, i.e. what character of the rocks or other change is being used, this seems a very curious position to be in. It seems that those of us working on Precambrian rocks are deciding where boundaries should be by consensus as to where people think they should be, before trying to define what it is we are reaching a consensus about.

Thus the definition of the Middle and Late Proterozoic boundaries as fixed points chronologically (Sims 1980) only has merit as an interim measure, so that we may use these terms as shorthand for post-2500 Ma, pre-1600 Ma, etc., but they will have no other meaning or implication. This should surely be simply a stage in the process of trying to discover major changes in the behaviour of the Earth, or in its biological or magnetic record that are sufficiently distinctive to enable correlation to be made world wide. These changes should then be used as the boundary markers for a chronostratigraphic column that may be applied to local lithostratigraphic successions. Properties that are possessd by igneous rocks (such as magnetic properties) would seem to be the most useful for correlating between the cratonic supracrustal sequences and the mobile belts which occupy so much of the Precambrian shields, so that such methods seem to hold out more hope of eventual adoption for the purposes of Precambrian stratigraphic subdivision than biological correlation.

There is, however, considerable scope for future palaeontological research in the Precambrian and recent attempts at Upper Proterozoic subdivision schemes based on biostratigraphy (Bertrand-Sarfati & Walter 1981; Vidal & Zoubek 1981) indicate the potential both in stratified deposits and also to a limited extent in metamorphosed sedimentary sequences. The problems of interpretation of these microfloras in such rocks are considerable (Zoubek 1981) and it is important to correlate on the basis of the fossils rather than on other criteria as well, or we will be in grave danger of circular arguments to prove, for example, that the Vendian tillites are all the same age.

Many of the arguments put forward in this paper are advanced by James (1978, 1981) who also suggests that biochronology even in the Late Proterozoic may only be accurate to within a few hundred million years. Trompette & Young (1981) have also concluded that 'it is not clear that definition of stratotypes will in fact significantly advance our understanding of the Proterozoic'. Therefore it seems reasonable, for the time being, to use the terms Archaean and Early, Middle and Late Proterozoic for broad divisions of Precambrian time and to define them in terms of radiometric ages. There is general agreement that the Archaean Eonothem shows many differences from later rocks but until these can be defined more accurately, or at least a definition internationally agreed upon, there is no basis for deciding where this boundary should be placed by other means than an agreed radiometric age.

With regard to subdivisions of the Proterozoic, although widely subdivided there has heretofore been no logic in such a subdivision as there are no major differences recognized between the three subdivisions. The only subdivision which is widely recognized is at the system level, where the Vendian was virtually accepted as the last system of the Precambrian, although its lower boundary has not been accurately defined. General usage accepts a tillite sequence as the lowest 'feature' of the Vendian. It remains to be seen whether a more closely defined system, the Ediacaran (Cloud & Glaessner 1982), will win more general acceptance.

A recently published geologic time-scale (Harland *et al.* 1982) proposes a set of chronostratigraphic names from a wide variety of continents. They suggest that these divisions will eventually be defined 'where there are good stratal characteristics for correlation' and this seems the logical way forward. However, it seems premature to suggest names for divisions before those good stratal characteristics have been found and without stating what they might be. They give a detailed discussion of the merits of various names for parts of the Late Proterozoic (Riphean, Sinian, Vendian, Sturtian, Ediacaran, etc.) pointing out that many of the names have been used with different meanings by different authors and then suggest a subdivision of the Late Proterozoic which is in reality simply another set of 'meanings' for these names without any attempt to define the new names in terms of stratal characteristics.

More recently Harland (1983) has suggested that the Proterozoic chronostratigraphic scale may be gradually evolving by usage (in Canada) to approximate to the original definitions of various Canadian lithostratigraphic groups. The major boundaries are in fact the classic unconformities (e.g. Keeweenawan, Huronian/Algonkian) which have been used to subdivide the Precambrian of the Canadian Shield for nearly a century. Such a scale would have the advantage of being defined in rock and Harland suggests that such reference standards could be correlated with sequences elsewhere in the world by 'magnetic, isotopic, climatic, biostratigraphic, geochemical or any other means'. Such standards, based on major unconformities, still do not have any correlatable characteristic and this seems to be an identical solution to that proposed by Kent & Hugo (1978 – see above) and there would seem to be no basis for recognizing the boundary outside the type area. Indeed if an unconformity is found elsewhere of approximately similar age the boundary will tend to be taken at the unconformity regardless of its true position.

It would seem more logical to defer any naming of Proterozoic eras or periods until distinctive differences have been recognized either in microbiota or other geological criteria, and it would be pointless to agree to names for Early, Middle and Late Proterozoic time based upon the agreed dates (Sims 1980) as such names will be necessary when actual changes seen in rocks are used as the basis of subdivisions. The use of local chronostratigraphic terms – e.g. the eras of Table 1 (Kent & Hugo 1978), which are based on lithostratigraphical sequences – have no chronostratigraphic use outside the immediate area of definition and seem to add nothing useful to Precambrian stratigraphy and it is suggested that the informal term '*Sequence*' be used if a chronostratigraphic equivalent to Supergroup or Group is desired.

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Some key rules for the calibration of the numerical time-scale

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SUMMARY: The building of an acceptable numerical time-scale, in whole or in part, is a complex matter. At the present time, it is often impossible to obtain even a nearly correct figure on which everyone can agree. This is often due to the selective rejection of data which, although considered unsatisfactory by some are accepted by others, and by the existence of data which appear at the time to be mutually incompatible. However, it is possible to draw a series of recommendations which could help in obtaining a result less open to criticism.

As a general scheme, a fully documented time-scale (a certainly megalomaniac aim) can only be obtained after a long itinerary of research comprising essentially four stages: (1) the collection of published data; (2) the evaluation of individual data; (3) the combination of data; (4) the drawing of conclusions: this paper summarizes the author's attitudes to these problems.

Collection of published data

There are two approaches to the collection of available data. The easiest is to search for and quote original papers. However, with some old results this leads to the consideration of results which, in the light of recent developments, particularly during the last decade, are of dubious value. It is better therefore to try to re-evaluate, when possible, old data using modern standards for stratigraphy, geochemistry and analytical methodology. This approach tries to realize a completely new re-assessment coupled with the inclusion of new data. It has been attempted in the past on two occasions: the joint meeting of the Geological Societies of Glasgow and London in 1964 and my own recent work (Odin 1982a). In both resulting documents, one may find a large number of possible calibration points more or less fully assessed or reassessed. The common problem for both of these works was to obtain pristine and unbiased information. The most important lessons to be learnt from these efforts seem firstly to be exhaustive, and secondly to select the most recent information on a given subject.

Exhaustivity

This aim can rarely be perfectly attained; however, the resulting selection has to try to *review all the most complete and sufficiently accurate studies*. A complete study needs to clearly examine the problems quoted below. It is sometimes useful to quote a result used in the past, even if someone personally considers the datum unreliable, just to show *why* it seems to be unreliable. On the other hand, unpublished results can hardly be quoted without details. If one needs to quote new results, a full statement and discussion is needed. The proposed synthesis then becomes the basis for publication of new informative results.

Selection of most recent results

Any geochronologist interested in the application of age determination to stratigraphy knows examples of the necessity of re-evaluating old results in the light of new calibration procedures or analytical results, new stratigraphical discoveries, or new interpretations of the significance of the data. The best criterion for assessing the value or otherwise of an earlier published datum is the latest opinion of the authors themselves *on their own results*. The best way to know of such opinions is, therefore, to question the original authors. In the authors' recent work (Odin 1982a) this approach was applied and it resulted in a rich collection of new information modifying many previous conclusions. The author does not intend to suggest that the best possible opinion on a series of radiometric data is necessarily that of the authors of the measurements but only that, when different results or interpretations have been successively given by these authors, the latest interpretations and results must be considered to be better than previous ones.

Individual evaluation of the data

This second stage cannot be achieved without access to complete information relating to stratigraphical problems, geochemical problems, and analytical problems (see Odin 1982b, p. 3–16). Stratigraphical problems occur especially with the more recent samples (post Jurassic) for which the absolute analytical precision is often very good compared with the biozonal correlation uncertainties. For this reason, the radiometric data obtained from areas where the type localities were defined will have more weight simply because correlation and extrapolation considerations are not necessary. In any case, the presence of useful diagnostic fossils in, or near, the dated formations should be clearly indicated and discussed, as well as other geochemical or geophysical lithologic characteristics sometimes used for correlation in recent years (see Odin *et al.* 1982). One may note that specific difficulties occur when stratigraphic definitions are frequently changing, especially, for example, the recent multiple biozonal redefinitions of stage boundaries within the Tertiary Era. The best solution may be to try to return to the original lithostratotype definition as the first order reference of correlation, if only to speak a common language.

The geochemical problems are those which may modify the interpretation of an apparent age. They relate the non-conformity of the geochronometer to the theoretical model of accumulation of radiogenetic isotopes in a perfectly closed, initially empty, 'box'. They have been subdivided into *genetical problems* which concern the moment of closure and the actual isotopic ratios at that time, and *historical problems* particularly disturbances which affect the theoretical model during the time span from the 'time zero' to the time of measurement. A detailed discussion of those problems becomes possible when a large number of analytical results have been obtained on a single formation using different kinds of geochronometers and methods of dating. Disturbances do not influence the different radioactive-radiogenic systems in the same way and it may thus be possible to identify such disturbances. Consequently, an age obtained on different geochronometers using different methods of dating can be weighted much more heavily compared with single mineral, single method ages.

The analytical problems include an heterogeneous series of consideration such as the calibration of the system of measurement, decay constants, methods of calculation, analytical reproducibility, etc. and also the *representativity* of the selected aliquot with regard to the whole formation investigated. It is well known that frequently, the analytical error quoted only concerns the short term analytical precision. The nature of the analytical error must therefore be carefully defined. A datum should be considered as complete only if the above quoted indications are readily available so as to permit other workers to discuss and independently evaluate the proper weight to be given to the particular point of calibration.

Combination of data

Schematically three main methods have been used in the past to try to build a numerical time-scale.

1. *The computer method* as developed by Armstrong (1978) consists of listing the numerical data wbich are related in three ways (maximum ages, minimum ages, and contemporaneous ages) to a given boundary. This, obviously, eliminates most of subjective considerations. But the weights of the diverse numerical data are artifically homogenized. This can give good results where many data are available but where they are scarce may bias the probable truth.

2. *The graphic method* makes the *a priori* admittance that the time-scale to be constructed does not depend only on the radiometric dates but also on a second, independant, variable viz, a 'known' relative duration between two or more portions of the time-scale. This relative duration is deduced from three main methods of *extrapolation* (or interpolation): sediment thicknesses (proportional to the time); biozone durations (of similar length); and the palaeomagnetic record of the ocean floor (a constant rate of spreading). This graphic method suffers from two problems. (a) It depends on the reference chosen and may he complicated by factors such as: extreme variation in thickness of sediments of the same age from one basin to another; the biozone duration of (for example) molluscs, foraminiferas, or nannofossils seem to vary separately; ocean floor spreading rates seem to vary in both time and place. The result is, therefore, highly variable from one study to another and moreover, no definitive proofs of such regular evolutions are presently available. (b) It is very difficult to determine objectively the actual uncertainties which should be attached to the resultant numbers; this problem in particular does not appear to have been appreciated by some of the authors.

The authors' own position is that it is of prime importance to determine the uncertainties so as to know if the obtained numbers are usable or not for independent scientific considerations of rates of evolution or the duration of geological events etc. In reality these are the important data which result from the establishment of an accurate time-scale. In any case this method gives to the selected points of anchorage, from which extrapolation is made, a very high weight; very few points in the whole time-scale are presently sufficiently well established to serve as such anchors.

3. *The exclusive radiometric method* consists of the determination of some stratigraphic boundary ages exclusively by using weighted radiometric dates. The more data one has, the more restricted should be the interval of time in which one may locate the boundary. However, in some cases, the abundance of dates leads to an increase in the required interval so as to include two proposals not fully in agreement; in other cases, a single good analytical result may lead to the proposal of too precise an interval of time which may be questioned later when new data are obtained. The main problem when using this method obviously is that diverse boundaries remain undated; but this method is the only one acceptable as giving a tool actually useful for geologists. My fourth rule is to recommend a systematic priority for the weighted radiometric dates.

The author considers it fundamental, when building the time-scale to use, as a first order constraint, the 'exclusive radiometric method' after an adequate weighting of the different data. It permits the definition of some boundaries without use of a subjective extrapolation system. In this exercise, it is of high importance to note clearly the studies from which the preferred number is deduced, as well as those from which the extreme acceptable numbers are obtained.

The derivation of a complete scale is usually done by combining the three methods. But this should he achieved only by taking into account the following proposals.

Distinction of the extrapolated ages

When the use of extrapolated ages seems the only way possible, these numbers must clearly be distinguished as extrapolated numbers. The author proposes, for example, to systematically show them in parentheses so as to indicate a low confidence level compared with dates obtained exclusively from radiometric data. An estimate of the uncertainty of these dates should be given in the same way as when using direct radiometric dates. Any new radiometric datum obtained later may supersede this tentative estimate.

Elimination of the lowest possible number of studies

As indicated above, the dates obtained from the diverse formations are to be weighted differently depending on the number of measurements done, on the number of geochronometers measured, and on the number of radiometric methods used. A particularly good criterion of confidence is when several laboratories have measured similar ages. However, due to the usually insufficient number of available data it appears better to retain as many of the data as possible and to reject as few as possible. For example, provided correct estimates of the actual analytical uncertainty are given, it is not a good solution to systematically eliminate *a priori* results obtained from one specific kind of geochronometer. Finally, it is rarely a good solution to eliminate many apparently inferior results just because one date or one formation appears to be highly reliable. These remarks would appear to be somewhat obvious, but experience shows that these rules are frequently rejected. In case of conflict it seems better to leave the situation open to further discussion. Three examples of common *a priori* rejection criteria may be mentioned here.

1. *The acid volcanics whole rock Rb-Sr ages.* It is commonly considered that whole rock Rb-Sr apparent age obtained on acid volcanics reflect rejuvenation by later alteration, especially by temperature overprinting. However, if a specific study proves that no such event has occurred after the extrusion of these rocks, there is no necessity to systematically doubt such results. The literature shows that some acid volcanics gave Rb-Sr whole rock apparent ages similar to contemporaneous granites, in formations as old as Cambrian (see NDS 249 and Gale, this volume, for further examples).

2. *Glaucony K-Ar ages.* It is well known, according to recent studies, that deeply buried, or slightly tectonized, or little-evolved glauconies have a clear tendency to preferentially lose their argon. However, a careful study of only slightly buried and highly-evolved glauconies, collected from undisturbed continental platforms, clearly show that apparent ages consistent with bentonites or volcanic flow ages may be obtained at the least until Jurassic time. No definitive good comparisons are available for older times, but the recent development of knowledge of the geochemical behaviour of this kind of chronometer (see review of recent data in Odin 1982) clearly show that results useful for time-scale calibration may be obtained from glauconies.

3. *Zircon fission track ages.* The first systematic use of zircon fission track ages in calibrating old parts of the time-scale is recent (Ross *et al.* 1982). Some confusion exists in this subject. The rejection of these data by recent authors is not related to the principle of this method, but to two specific problems. The first is that the analytical errors calculated in this particular study were almost certainly underestimated, and that 1σ errors were compared with 2σ errors generally used by other geochronologists (see Gale & Beckinsale, 1983). The second is that the retentivity of the tracks during a very long period of time is questionable. The inadequate state of knowledge in this matter is well illustrated by the very uncertain blocking temperature for zircon fission tracks, sometimes estimated at 180°C and sometimes at a temperature much lower, or higher, as well as by the absence of agreement on the decay constant and the absence of a well accepted rock age standard. However, all specialists agree that future refinements of the method is probable, and that confirmatory results obtained from similar rocks in different laboratories will obviously give weight to the method. The present position therefore is that, if a correct estimate of the analytical problems remaining with these ages is made, it will be difficult to refine the present numerical time-scale of the Palaeozoic based on other methods of dating due to the large analytical error which must be accepted. New studies will certainly prove in the near future the importance and interest of this pioneer work.

Bias when data are available from a single side of a dated boundary

The comparison of age estimates made in the early stages of time-scale development, with the solutions proposed today using new results, shows a tendency to a systematic bias when the early data documented only a single side of the dated boundary. In practice, a boundary essentially dated on the old side tends to be estimated too young because possibly rejuvenated rocks have influenced the conclusion, while a boundary essentially dated on rocks from the young side will tend to be estimated too old. Consequently, a good estimate will only be obtained when a nearly similar number of data are available on both sides of the boundary. An increase of the time interval in which the boundary may be located should perhaps be made on the side where most data are available to counterbalance this tendency.

Bias due to the systematic selection of the oldest results

It is commonly accepted by geochronologists that diverse processes of alteration usually lead to a preferential loss of the radiogenic isotopes relative to the radioactive ones. Consequently, when such an opening of the system is suspected it is customary to accept the oldest apparent age measured as the best or minimum estimate, of the actual time of closure of the system. The systematic application of this rule, when in reality the system has not been open during its history, may lead one to overestimate the actual age of the dated formation. Several factors can also result in measured apparent ages being *older* than the age of formation of the system under investigation. These are, analytical bias, choice of an incorrect initial isotope ratio and preferential loss of radioactive isotopes. These possibilities have been reviewed in recent years but two specific examples will be recalled; (1) When using glauconies, it has been proved that, in many cases, the initial apparent ages of this geochronometer is positive (Odin & Dodson 1982). The selection of the older analytical result is consequently the wrong way to obtain the actual age of an undisturbed formation. (2) When using altered biotites from bentonites, Obradovich & Cobban (1976) have shown that apparent ages older than that of the ash emplacement may be measured as well as younger ages.

There exist, fortunately, adequate criteria with which to resolve the possibilities of too old ages; they consist of the examination of the field data together with geochemical studies or sophisticated geochronological studies on several geochronometers. It is therefore important that full analytical and petrographical details are obtained and indicated when quoting a new age estimate of a stratigraphically well defined formation. Moreover, although justifiable in many cases, an assessment of a minimum age as being correct cannot be accepted without very full discussion.

Concluding proposals for a numerical time-scale

One can easily imagine the implications of each proposal quoted above on the numbers suggested as presently fixing the age boundaries of the stratigraphic units. Some of these proposals will influence the selected preferred number; but most of them will influence essentially the *confidence level* of an age. In order to give a realistic picture of the situation for a given boundary, it seems to be much better to try to locate a boundary not at a point (i.e. the preferred number) but in an *interval of time* which attempts to include most of the possibilities generated by the available data, in other words; *definition of an interval of time is better than of a number.*

The result of an estimate will not be *one* number but *three* numbers; the maximum possible age, the minimum possible age and the preferred age. This can be written either

'T + x_1 − x_2' or 'T_1 to T_2'. T being ages in years, x being numbers in years showing the uncertainties on the old side (either T_1 or T + x_1) and on the young side (either T_2 or T-x_2). There is no necessity, in this proposal, that x_1 and x_2 should be equal.

An example of a summarized age estimate

The example discussed below could only be fully exemplary if it is viewed as part of a larger study including the original results and recent reassessments. This age estimate depends on radiometric dates, and not on graphical or other less rigorous extrapolation arrangements, and concerns the Llandeilo-Caradoc boundary. The use of radiometric dates from both sides of a studied boundary to estimate the interval of time in which it probably lies appears at present to be the most realistic system. As far as possible, this system was used by the different authors when building the syntheses gathered in Odin (1982).

The proposed Fig. 1 does not include any pre-estimate of the age or duration of the stages or zones which are just shown in their relative stratigraphical succession. The figure also includes a rather complete series of dates because, in order to avoid any *a priori* assumptions on the results to be reached, it is better to reject the minimum of constraints, provided that an estimate of the actual stratigraphical, geochemical and analytical uncertainties are made.

The items used are summarized in Table 1. All of them have already been quoted elsewhere in this volume or in Volume 2 the authors' own review (Odin 1982). However, one may note that the very precise recent results obtained by Williams *et al.* (1982) on the Kinnekulle bentonite minerals were preceded by interesting studies, the last of which was that by Baadsgaard & Lerbekmo (1982). Although analytically less precise, these results obtained in an independent laboratory and quoted 'new' in the table, support and give a high weight to this calibration point which, alone, could lead to propose a boundary older than 445 Ma. This point, together with other results recently published, enables the author to locate the boundary in an interval of time around 450 Ma: older than previously proposed (433–443 Ma: Gale 1982).

Looking at the constraints available *above* the studied boundary, one may note first, that the mid Caradoc Tyrone limestone has given apparent ages at 443 ± 10 and more recently at 454.1 ± 2.1 Ma (Kunk *et al.* this volume). The last error bar is clearly related to analytical precision. A more realistic complete analytical uncertainty must include a minimum of ± 1% of the age due to calibration, especially necessary when using the 40Ar-39Ar method due to the absence of good interlaboratory standards. The resultant figure of 454 ± 7 Ma is in analytical agreement with the K-Ar age of 443 ± 10 as well as with the age of the probably

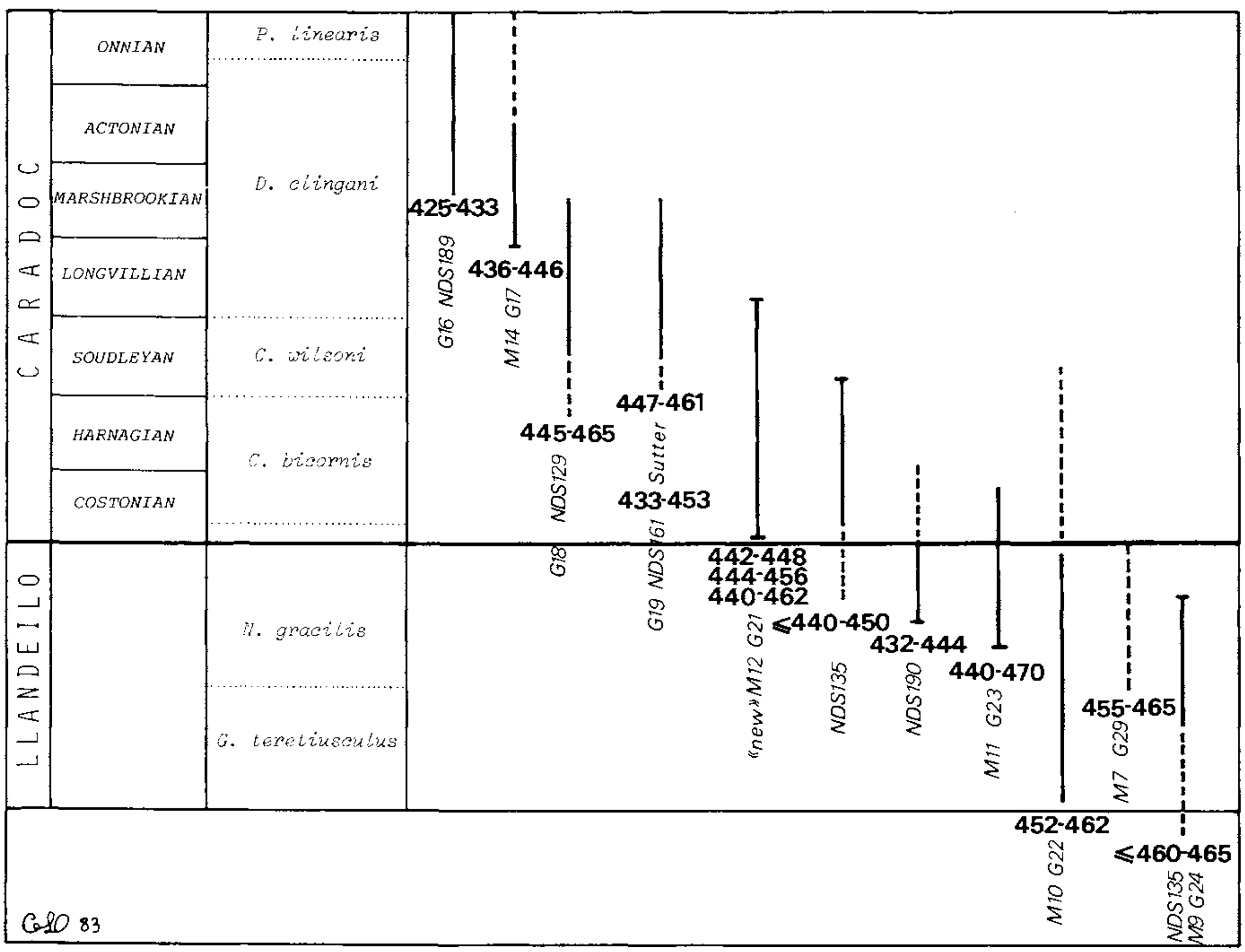

FIG. 1. Presently available radiometric ages versus their relative stratigraphic position around the Llandeilo–Caradoc boundary.

TABLE 1. Geochronological results useful for estimating the interval of time in which the Llandeilo–Caradoc boundary probably lies. In the list of items, NDS represents the number of the abstracts gathered in Odin 1982a; G. represents the number of the items used by Gale in this volume and M represents the number of the items quoted by McKerrow *et al.* in this volume for comparison.

Item	Formation	Dating method	Age (Ma ± 2σ)	Stratigraphy
NDS 189 = G.16 —	Eskdale granite	Rb-Sr, w.r. 10 pt	429 ± 4	Late Caradoc
— G.17	Oliverian Syenite	Rb-Sr, w.r.	441 ± 5	Post lower clingani z;
NDS 129 = G.18 —	Carters limestone bentonite	K-Ar, 4 biotites + 1 sanidine	455 ± 10	Middle Caradoc (or Early)
NDS 161 = G.19 —	Tyrone limestone bentonite	K-Ar 1 biotite	443 ± 10	id id
Sutter = (G.20) —	id	39/40 Ar	454 ± 7	id id
— G.21 = M.12	Kinnekulle bentonite	Rb-Sr biotite	445 ± 3	post gracilis -pre clingani z.
		K-Ar biotite	450 ± 6	id
New — —	Kinnekulle bentonite	K-Ar, 1 biotite + 3 sanidines	451 ± 11	id
NDS 135 : — —	Tormitchell conglomerate	Rb-Sr, w.r. 5 pt	≤ 451 ± 5	pre-*wilsoni* z.;
			≤ 440 – 450	
NDS 190 : — —	Threlkeld microgranite	Rb-Sr, w.r., 13 pt	438 ± 6	Late Llondeilo (or post)
— G.23 = M.11	Bail Hill volcanics	K-Ar biotite	455 ± 15	gracilis to post gracilis z.
— G.22 = M.10	Borrowdale volcanics	Sm-Nd	457 ± 5	Llondeilo (to Early Caradoc)
— G.29 = M.7	Bay of Island gabbro	40/39 Ar	460 ± 5	gracilis zone (or older)
NDS 135 G.24 = M.9	Benan conglomerate	6 Rb-Sr, w.r.	≤ 470 ± 5	pre-gracilis
			≤ 460–465	

correlative Carters limestone at 455 ± 10 Ma (see table).

The maximum apparent age of the pre-*wilsoni* granite boulders collected from the Tormitchell conglomerate leads Longman *et al.* (1982) to propose an age of about 440–450 Ma for Early Caradoc times.

The results from the Oliverian syenite are less constraining due to a large stratigraphical uncertainty. However, the present results show, if the dates are all correct, that the boundary must be older than 447 Ma, the minimum age of the Tyrone limestone (40/39 datum); and younger than 450 Ma, the oldest age proposed for the Tormitchell conglomerate.

Looking at the dates from samples stratigraphically located *below* the boundary, the constraints are as follows. The Benan Conglomerate granite clasts apparent age leads Longman *et al.* (1982) to suggest a numerical age of 460–465 Ma, or younger, for the boundary because the granite was probably extruded within the Llandeilo. The Bay of Island gabbro also suggests an age younger than 455 Ma for a Late Llandeilo or possibly older formation. The Borrowdale volcanic Sm-Nd date (457 ± 5 Ma) shows that these, more probably Llandeilo (possibly Early Caradoc) extrusive rocks, are older than 450 except if the Sm-Nd age is inherited. The Bail Hill volcanics, rather precisely located in the stratigraphy, just show that their imprecise apparent age (recalculated by Gale, in this volume) does not permit one to diminish the interval of time in which the boundary can be located. The final constraint is obtained from the Late Llandeilo Threlkeld microgranite from which an age of 438 ± 6 Ma was obtained according to the summary of Rundle (1982) although it is not completely impossible that this rock was emplaced slightly later during the earliest Caradoc times. This datum provides a maximum age for the boundary at about 444 Ma.

We have therefore a conflict between probably mid Caradoc ages (the Tyrone and Carters limestones) at more than 447 Ma which, both together, would suggest an age of 450–460 Ma for the boundary, and ages at less than 450–440 Ma (Tormitchell conglomerate and Threlkeld results) for the same boundary.

Consequently, without any hypothesis as to the relative durations of series or zones, the presently available radiometric data lead to the conclusion that the Llandeilo-Caradoc boundary must lie in the interval of time 450 ± 8 Ma, without rejecting any of the available results. The maximum and minimum ages of this interval are equally possible depending on the preference given to one or another set of results. The future will show if this estimate may be more precisely expressed; but it is clear that this boundary will remain more stable than many others, less documented, and for which just one new dating may change the previous age estimate.

The present summary must, however, be seen as a review following an individual assessment or reassessment of each data set quoted in the table. At the present time, no definitive proof of incorrect age assessment due to stratigraphical, geochemical, or analytical problems has been shown for any of the data used. The extreme values proposed just exclude one date on the old side, and two ages on the young side.

Conclusion

In conclusion, although attention has been recommended to many points in the difficult exercise of building a time-scale, the author must underline that the problem is usually made easy by the geologists who publish complete geochronological results. On the other hand, the author is fully aware that it is nearly a utopian hope to obtain results on which everyone will agree; he has tried! A completely unquestionable estimate will frequently necessitate an excessive increase in the error margins. The experience (i.e. the various solutions accepted by the people who have tried to achieve the synthesis of a time-scale) shows that, to some degree or another, it will always remain possible to modify or refine some part or another of the monumental multicontributor job that needs to be done.

This volume has tried to answer parts of the many remaining questions. However, for the future, the key documents have to include, together with the tentative syntheses, more new and original data, and periodic re-examinations of the foundations of geochronology applied to stratigraphy. These foundations consist of both methodological research to explain the behaviour of geochronometers, and compilations of complete abstracts of the available data.

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Late Precambrian and Cambrian geological time-scale

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SUMMARY: The suggestion by the Subcommission on Precambrian Stratigraphy (of the International Union of Geological Sciences) that the beginning of Late Proterozoic times should be defined in the geochronometric scale at 900 Ma is accepted; this probably correlates with the chronostratic scale at the Middle–Upper Riphean boundary. The Upper Riphean–Vendian boundary on the chronostratic scale seems to correlate with approximately 650 Ma. The chronostratic Precambrian–Cambrian boundary is poorly correlated with the chronometric scale on a global basis at present but may be more accurately related in the next decade or so. In the meantime a range of 530–600 Ma can be considered but no conclusions are offered here.

An attempt is made to give the present state of knowledge and techniques, with only a selection of radiometric data, especially those bearing more or less directly on the calibration of chronostratigraphic units. The geographic coverage is not intended to be complete.

At present there are two contrasted ways of dealing with definitions of major units of time.

An ideally integrated geological time-scale, which is still only a hope for the future, would in the authors opinion, be composed of standard chronostratigraphic divisions based on rock sequences (evidence) and accurately calibrated in years.

The two differing kinds of scale (Harland 1975) are;

1. A chronostratic (chronostratigraphic) scale mainly aimed at producing a scale of continuous rock sequences with standardized reference points selected in stratotype sections.

These complete and internationally acceptable sections with reference to a particular boundary are the global boundary stratotype sections and points between stratigraphic units, e.g. the Silurian–Devonian boundary stratotype section at Klonk in Bohemia, Czechoslovakia which is well curated and accessible.

The order of procedure in the Phanerozoic (or 'Holozoic' — Glaessner 1984) Eon is to establish the chronostratic scale first and then to calibrate it by relating it to the chronometric scale. The beginning of the Phanerozoic Eon is here conventionally (and arbitrarily but there is much international agreement) taken to be conterminous with the beginning of the Cambrian Period. The Precambrian–Cambrian Boundary is also defined on the chronostratic scale.

2. A chronometric scale defined and based on units of duration — years or million of years (Ma). The Early-Medial-Late (Lower-Middle-Upper) Proterozoic I, II, III units suggested (but not yet ratified) by the IUGS Subcommission on Precambrian Stratigraphy with limits at 2500 Ma, 1600 Ma and 900 Ma are proposed as part of such a chronometric scale. Periods in the Precambrian Supereon (Glaessner 1984) are also being put forward with chronometric limits.

As research proceeds, progressive and regular revisions of the integrated geological time-scale are required. Efforts have recently been made by Harland *et al.* (1982) and Gale (1982).

In this article the main time-divisions of the Late Proterozoic and Cambrian such as the Middle-Upper Riphean, Upper Riphean–Vendian, the Vendian–Cambrian boundary and the time-span of the Cambrian Period are discussed. The Cambrian–Ordovician Boundary is left to other authors in this volume who are concerned with the Ordovician Period and the authors note their figure of *c*.515 Ma for this time-boundary.

None of the above boundaries have yet been fixed or correlated on a world-wide basis although it is generally agreed that there should be a separation of Late Proterozoic time (Upper Proterozoic or Late Proterozoic or Proterozoic III) from 900 Ma up to the beginning of the Cambrian period but the correlations are poorly established.

In discussing late Proterozoic sequences emphasis must be given to the USSR craton, followed by China and Australia, where the most significant steps have been made towards a chronostratigraphic scheme, based on palaeontological data and correlated with radioisotope dating. However, it must be pointed out that the base of the Upper Riphean (≡?Proterozoic III) in the type-area, the Ural Mountains, is an unconformity and the base of the Vendian is a tillite horizon.

One of the consequences of establishing by international agreement global boundary stratotype sections and points guided by biostratigraphy which may cause conflict with classic and entrenched methods is that previous practices and compromises in establishing the boundaries of periods, eras and eons should not be allowed to go unquestioned. This particularly applies to the Precambrian–Cambrian Boundary. These practices and compromises include the use of unconformities and changes of tectonic grade and style and of facies of special stratigraphic significance when the unfossiliferous rocks above and/or below the unconformity are not capable of being diagnosed on the basis of biostratigraphy.

It is widely agreed that all Phanerozoic boundaries (including the Precambrian–Cambrian/Proterozoic–Phanerozoic Boundary) are to be *defined* chronostratigraphically (*not*

TABLE 1. Chronostratic scale usage in Phanerozoic Eon contrasted with chronometric scale usage in subdividing Precambrian Supereon

chronostratic scale →	Phanerozoic		(?570 Ma)	
		III	900 Ma	chronometric scale
		II	1600 Ma	
	Proterozoic	I	2500 Ma	
		III	2900 Ma	
	Archaean	II	3500 Ma	
		I		

TABLE 2. Subdivisions of the Late Proterozoic

Eon	Era		Period	Epoch	Age in Ma
PROTEROZOIC		SINIAN(Z)			590
			Vendian (V)	Ediacaran	
				Varangian	630
	Pt_3				670
			Sturtian (U)		
					800
	(?)	RIPHEAN (R)			900
	Pt_2				1050
			Yurmatin (Y)		
					1350
			Burzyan (B)		
	(?)				1600
	Pt_1				

TABLE 3. Late Proterozoic succession in the USSR

	Cambrian		
		570 Ma	
	Vendian		Stratotype is Moscow syncline.
			Ediacaran metazoa.
	▲ ▲ ▲ ▲ ▲ ▲ ▲ ▲	650–680 Ma	Laplandian Tillite (Uncon. at base; change in microfossils).
Upper Riphean	R_4 Kudashian		Marked by incoming of Yudomian
		c.680 Ma	assemblage of stromatolites and microphytolites.
	R_3 Karatau	c.920 Ma ± 50	change in stromatolite assemblages:— used to previously be taken as the R_2/R_3 boundary for some workers.
		1000 ± 50 Ma	Unconformity between R_2/R_3 in the Urals: 1100 Ma
Middle Riphean	R_2 Yurmatin		date from a diabase below the unconformity.

Notes:
The dates shown in the middle column are K-Ar glauconite determinations. In addition the following points may be noted:—
(i) Keller (1979) noted a possible lower age of 620–600 Ma for the Ediacaran fossils in the USSR – this would seem to be younger than Australian occurrences (see below).
(ii) K-Ar glauconite datings in Vendian: 610–560 Ma (Valday Group); 510–610 Ma (Volyn basalts)
(iii) K-Ar glauconite dates of 660–665 Ma (from just below the Laplandian Tillite) and 610 Ma (from just above it). The age of the tillite is put in the 630–650 Ma range by Chumakov & Semikhatov (SI1).
(iv) Kudashian (R_4) — its lower boundary is established where the Upper Riphean stromatolite assemblage and microphytolites are replaced by a fourth assemblage. This change is marked by a disconformity in parts of Siberia and middle Asia. The fourth assemblage of microphytolites is, however, recognized in the Upper and even Lower Riphean in the Volga–Uralian area.
(v) Vendian and Riphean are *chronostratigraphic* units: there are no reliable radiometric data for dating the lower boundary of the Vendian.
(vi) If one accepts Pringle's (1973) date of 653 ± 23 Ma for the Nyborg Formation some of the tillites are younger and others older than 650 Ma. In view of the uncertainty about the age of the base of the tillite units and because of the difficulty of correlation between basins, tillites cannot be regarded as good chronostratigraphic markers.
(vii) It should be noted that the Varangian epoch has two glacial episodes not only in Finnmark but elsewhere in the North Atlantic region (Mortensnes (top); Nyborg Formation, Smalfjord Formation (base) and Russian workers correlate the two tillites in the European USSR with these glacial episodes.

chronometrically) and guidance in most of the younger part of the geological column is by biostratigraphy (Cowie *et al.* 1984).

Ideally, the chronostratigraphic position of the Pϵ-ϵ Boundary should be placed between (as closely bracketed as possible) an internationally accepted earliest Cambrian faunal assemblage zone above with a similarly accepted latest Precambrian zone below. This is not widely achieved as yet but the recommended global stratotype section and point in south China comes close to it and other possibilities are east Siberia and Mongolia with the poorly accessible northern Siberia as a hope for the future.

Late Proterozoic, 900–?570 Ma

The time-scale

The IUGS Subcommission on Precambrian Stratigraphy has made the following provisional recommendations for the chronometric division of the Precambrian (Sims 1980; James 1981).

The figure of ?570 Ma is here (in Table 1) put in brackets because James as a member of IUGS – IGCP Project 29 Working Group on the Precambrian–Cambrian Boundary Working Group is well aware that the Pϵ-ϵ Boundary can only have *one* definition — chronostratic *or* chronometric — and it has been decided since 1974 that the Working Group will define it chronostratigraphically by a physically fixed boundary stratotype point in a section which is globally validated.

In this article we are concerned only with Proterozoic III (Late Proterozoic) and its boundary with Proterozoic II (Middle Proterozoic). As a compromise between Australian, Russian, Chinese & Scandinavian terminology the authors propose the following subdivisions of the Late Proterozoic in Table 2. A similar scheme is used by Harland *et al.* (1982).
In two special issues of *Precambrian Research* (**15**, 1981 and **18**, 1982) – 'Upper Precambrian correlations' in 1981 and 'Geochronological correlation of Precambrian sediments and volcanics in stable zones' in 1982 – referred to here as SI1 and SI2 – the following topics were discussed:
1. Middle–Upper Riphean Boundary
 (i) unconformities
 (ii) stromatolite stratigraphy
 (iii) radiometric data
 (iv) microfossils

2. Upper Riphean–Vendian Boundary
 (i) metazoa
 (ii) Yudomian stromatolites and microphytolites
 (iii) Varangian tillites (Varangian has priority over

TABLE 4. Late Proterozoic succession in China (see also Tables 7, 8 and 9)

Cambrian System		
	610 Ma	
(?R_4)	*Sinian System* — Rb/Sr (Toushantuo Fm.): 700, 691 Ma Rb/Sr (Nantuo Fm.) : 739, 714 Ma Acanthomorphs, Prismatomorphs, *Sphaeromorphites*	
		*c.*800 Ma
(?R_3)	*Qingbaikou System* — K/Ar glauc. (Jing'eryu Fm.): 853, 862 Ma Mega-plants, *Inzeria-Linella* assemblage	
		*c.*1050 Ma
(?R_2)	*Jixian system* — K/Ar glauconite from upper Formation: 1134, 1152, 1011 Ma micro-plants, Dasycladaceans, large conical stromatilites in upper part	

Notes:

(a) In Jiangnan and the Upper Yangtze and also in Qinling, the Sinian system was preceded by the Xuefeng orogeny, while the Wuling orogeny antedated the Qingbaikou system.

(b) Metazoa (medusoid affinity) occur in the Wuhangshan Group (? base of the Sinian or top Qingbaikou) – does this fit in with USSR or Australia?

(c) Tillites:
- (i) Chang'an glacial epoch – 800–760 Ma = Sturtian of Australia
- (ii) Nantuo glacial epoch – 740–700 Ma (Ma Guogan *et al.* 1980) = Marionoan & early Varangian
- (iii) Luognan glacial epoch – 640–600 Ma

All these tillites occur in the Sinian system: (iii) is the only candidate for correlation with the Laplandian, but it seems too young, Chumakov (SI1) correlates it, however.

(d) Cai Xuelin *et al.* (1980) report that low-grade meta-sediments (Hui Li System) are covered unconformably by the Sinian system. A quartz-dolerite intrusion pre-dating the unconformity gives an age of 910 Ma, and basalt in the lower part of the Sinian system gives 759 Ma.

TABLE 5. Adelaide geosyncline succession, Australia

	Cambrian		
		Disconformity	
Late	Ediacaran		Ediacaran metazoa, Stromatolite, *Tungussia* cf. *julia*
Adelaidean	Marionoan		
		— 690–680 Ma Rb/Sr: 750 ± 53 Ma	Stromatolite assemblages suggest R_4/Vendian boundary
	Sturtian ▲ ▲ ▲ ▲ ▲ ▲ ▲		Sturtian glacial epoch not long before *c.*750 Ma (but could be much older) i.e. two episodes at 800 & 790 Ma.
Early			
Adelaidean			
	Torrensian		Stromatolites: *Baicalia burra* and *Tungussia wilkatann*
Base?	Willouran		Beda volcanics. Stromatolites — *Acaciella, Gymnosolen*

Varegian, Varangerian, Veregian and Laplandian)
(iv) radiometric data
(v) microfossils

In the following account the authors summarize the main points in SI1 and SI2 and review, on a geographical basis, the evidence for dating the boundaries between the Middle and Late Proterozoic and between the Riphean and the Vendian.

USSR (Chumakov & Semikhatov *in* SI1; Keller 1979)

The Late Proterozoic succession in the USSR is established as follows (Table 3) with the stratotype of the Riphean suberathem in the Bashkir anticline in the Urals (Shatsky 1945).

China (Chen *et al.*, in (SI1))

The Sinian Suberathem (615–1950 Ma) has now been replaced by a number of systems with the youngest, immediately below the Cambrian system, being the Sinian System (see Table 4). As the same name should not be used for two different meanings Sinian Suberathem is suppressed. In the Yangtze River region the Sinian System rests on an igneous–metamorphic complex which includes the Huangling granodiorite. On this rock an age of 860 ± 50 Ma has been obtained from U-Pb measurements on zircon and apatite (Ma Guogan *et al.* 1980). Six stromatolite assemblages are considered by Chinese palaeontologists to show many similarities with the USSR and Australia (Bertrand-Sarfati & Walter *in*

TABLE 6.

Eon	*Era*	*Period*	*Epoch*	*Age & Region* (≡ Stage) Yunnan, S. China	SE Siberia, USSR
Phanerozoic					
	Palaeozoic				
		Cambrian			
				Qiongzhusi	Atdabanian
			Early Cambrian (part)	Meishucun (Zones II & III)	Tommotian
Precambrian–Cambrian Boundary					
				Meishucun (Zone I)	Yudomian
			Latest Precambrian (part)		
		Sinian/ Vendian/Ediacaran		Dengyingxia	
	Riphean				
Proterozoic					

SI1, p. 356). There is no record of the use of microphytolites. Metazoa (Sabelliditidae, *Micronemaites*) are found in the Dengying Formation which is mainly of Sinian System age but is Cambrian at its top (Table 7).

Conclusions. From this discussion the base of the Vendian in China cannot yet be identified.

Australia (Preiss & Forbes *in* SI1)

The Adelaide geosyncline provides the main evidence as detailed in Table 5.

Other basins are:

1. Amadeus: post-1053 Ma (date of metamorphic complex) succession with stromatolites which are mainly new forms. Walter (1972) regards them as a typical Late Riphean assemblage.
2. Officer: post-1037 Ma succession with glacial deposits (?=Sturtian).
3. Kimberley: Succession with ?Sturtian and Marinoan glacials (672 ± 70, 670 ± 84 Ma ages from shales above the glacial deposits).

Jenkins (1981) recognizes an Ediacaran Period embracing the time interval characterized by soft-bodied metazoa of which the earliest are at about 640–620 Ma. The type-area is in the Flinders Ranges, South Australia. The lower boundary of the Ediacaran lies just above the Marinoan glaciogenic horizon. On this basis the Ediacaran Period overlaps the late Varangian glacial epoch but the Varangian epoch includes by definition both glacial epochs. Jenkins notes 'a close equivalence between the Vendian and the Ediacaran' but this is not followed by Harland *et al.* (1982) nor Cloud & Glaessner (1982).

Cloud & Glaessner in 1982 examined possible age brackets for their proposal for an Ediacaran Period as follows:

1. Rb-Sr whole rock isochrons from shales of lower to middle Ediacaran age (correlated with the lower Brachina Formation of their type area): 672 ± 84 Ma, in South Australia.
2. Beds with medusoids (*Kullingia concentrica*) above the uppermost varangian tillites are post 653 ± 23 Ma which is the recalculated Rb/Sr whole rock isochron age for the tillites (Pringle 1973) in Scandinavia.
3. Rb/Sr whole rock isochron on the Holyrood granite which is intrusive into the Precambrian Conception Group of Newfoundland, Canada. This date is recalculated by modern methods (Gale 1982) at 585 ± 15 Ma: see below.
4. From North Carolina U-Pb concordia age on zircons from slightly metamorphosed pyroclastics which are of Ediacaran equivalence – 620 ± 20 Ma.
5. In Charnwood Forest, England (i) Rb/Sr isochron from diorites intruding strata with an Ediacaran type of medusoid: 540 ± Ma, and (ii) Rb/Sr isochron from diorites: 546 ± 22 Ma.

These English dates are much younger than those given above (1–4) and must be examined critically in the light of discussion later in this paper on the Precambrian–Cambrian Boundary (Ediacaran–Cambrian *or* Vendian–Cambrian Boundary).

Conclusion. The dating of the Marinoan glacials (between 750–676 Ma) seems too old for the base of the Vendian. The 750 ± 53 Ma date comes from the Tapley Hill Formation just below the Marinoan glacials.

Europe (Choubert & Faure-Muret 1980)

1. *Armorican Massif.* Most of the Brioverian has Lower and Middle Riphean microfossils. The Upper Riphean and Vendian only occur in Normandy.
2. *British Isles.* Accepting the correlation of the Port Askaig Tillite (= Mortensnes glacial epoch) with the Varangian tillites (inter-tillite beds, Nyborg Formation dated at 653 ± 23 Ma by Pringle 1973) then it is possible to view part of the Dalradian Supergroup as Vendian and this is supported by the discovery of Vendian acritarchs, worm burrows and the stromatolites *Aldania* and *Jurasania* (Downie 1975). There is an earlier (?Sturtian) tillite in the Appin Group.

Conclusions: the Late Proterozoic

1. There is a lot of agreement on a date for the beginning of the Late Proterozoic (III): Russian, Chinese and African workers place it at roughly 1000 Ma. This is a *maximum* age and provisionally the IUGS Precambrian Subcommission's suggestion of 900 Ma for the initial houndary for the Late Proterozoic can be followed without serious difficulty. As noted earlier some Russian workers used to correlate the chronostratic R_2–R_3 boundary with 920 Ma because of a change in stromatolite assemblages at about that time.
2. The use of the Varangian glaciation to fix the base of the Vendian has only limited (European) value at the present time. Correlation of glacial epochs in China or Australia with the Varangian are premature. In most areas neither radiometric data nor the fossil record are sufficiently detailed to permit correlation of the R_4–Vendian boundary. Until better correlation is achieved 650 Ma is an acceptable date for this boundary. Good sections in Svalbard give good biostratigraphic control.

Precambrian–Cambrian boundary

Principles and general comments

The Precambrian–Cambrian Boundary is assumed here to be accepted by many geologists (perhaps the majority) as conterminous with the Proterozoic–Phanerozoic Eons Boundary and the Proterozoic–Palaeozoic Boundary.

The 'Base of the Cambrian' is a term sometimes loosely used and is a lithostratigraphic and local concept subject to diachronism but can never be older than the Precambrian–Cambrian Boundary. The Precambrian–Cambrian Boundary is a chronostratigraphic concept for an instant in time and cannot therefore be subject to diachronism. The latter term is used below because of its greater fundamental value.

The international definition of the Precambrian–Cambrian Boundary in terms of a global stratotype section and point was close to agreement in early 1984. The IUGS-IGCP Working Group on the boundary has decided by a 80% majority of its Voting Members to place it at the base of Bed 7 (within the Meischucun Stage at Fossil Zones contact I/II) in the section at Meishucun, Jinning County, Yunnan Province, China (~ 75 kms south of Kunming, latitude 25°N, longitude 103°E). The Working Group decision has yet to be ratified by the International Commission on Stratigraphy of IUGS, however.

Rocks of late Precambrian age assignment depend for validation of their truly Precambrian internationally agreed Proterozoic chronostratigraphic age on a complex of correlative factors which may not be available in a particular locality, area or region. An important reference standard will be the internationally agreed global stratotype section and point: a lithostratigraphic equivalent selected to define the chronostratigraphic Precambrian–Cambrian Boundary. This stratotype is and must be unique and has been selected with great care so that sections and points elsewhere can (as far as is possible and reasonable) be correlated with it by all the means available in the geological sciences. (see Table 6)

From readings in the literature it is evidently important to distinguish between sedimentary strata which appear to be the oldest Cambrian deposits in a region such as the British Isles (or a comparatively small continental area such as north-west Europe) and the earliest Cambrian deposits on a world scale. The Precambrian–Cambrian Boundary as now developed in its definition is a concept with international and intercontinental utility when expressed in terms of a single, unique global stratotype section and point.

Current working decisions include (Cowie 1981):

1. The Precambrian–Cambrian boundary stratotype point should be placed as close as is practicable to the base of the oldest stratigraphical unit to yield Tommotian (*in sensu lato*) fossil assemblages.
2. The exact geographical and stratigraphical positions of the recommended boundary stratotype point are given above.
3. The guiding criterion is that the Precambrian–Cambrian boundary should be approximately located in the chronostratigraphic scale near the evolutionary changes which are signalled in the rocks by the appearance of diverse fossils with hard parts. It is emphasized, however, that we are defining and placing the boundary stratotype point in the rock and it will then be a reference point which should remain fixed despite the possibility of fresh discoveries in the rocks stratigraphically below or above.

With the present level recommended by the international IUGS/IGCP Working Group for the Precambrian–Cambrian Boundary obviously some regions, including the British Isles and north-west Europe, may well find that their basal Cambrian deposits and their included faunas and floras plus associated geochronometric indications are not assigned to the earliest Cambrian as defined on a global scale.

Evidence surviving in many parts of the world shows that marine transgressions occurred over large parts of the Earth's surface during late Precambrian and/or early Cambrian times. They were, of course, diachronous and grossly diachronous as a global process. There are many parts of the world, however, where sedimentary rocks identified as late Precambrian are followed conformably by sedimentary rocks identified as early Cambrian. Two factors bear especially on the problem:

1. In certain regions rocks which are conventionally and provisionally assigned a Precambrian age are distinguished from rocks which are in a similar manner assigned a Cambrian age, on the grounds of position below and above a very obvious marine transgression. In the absence of fossils or other methods of dating (chronostratigraphic and/or chronometric) there is a clear danger of circular argumentation.
2. The correlation of the geological time-scale is extremely coarse at this stratigraphical level and errors are possibly quite large (millions or tens of millions of years may be involved). Even when based on the best available methods and results in biostratigraphy, geochronometry, magnetostratigraphy, sedimentological cycle interpretations, palaeoclimatology and perhaps most important but most difficult — palaeo-oceanography, there is still only inexact correlation.

In any discussion of the Precambrian–Cambrian Boundary it is, in the authors' opinion, essential to consider Precambrian radiometric dates as well as those obtained from Cambrian rocks. Where possible, dates from the late Precambrian rocks in any region should be sequential in value with dates found close to the boundary with the Cambrian system and in the early Cambrian rocks. There is no theoretical or practical basis for a gap or an overlap when the geochronometric and chronostratigraphic scales are collated: time is continuous even if sedimentation, magmatism and tectonism are not.

USSR

Soviet geochronologists have established a great number of dates many of which relate to late Precambrian and early Cambrian strata and some of these preliminary results were used with qualifying remarks in Cowie & Cribb (1978). These earlier determinations, based mainly on K-AR dates, have not always been possible to evaluate fully in non–Soviet countries because of the difficulty of obtaining access to the constants used, the errors which should be quoted and the geochemical/radiometric information needed but they *have* been published, some in English, and *can* be obtained. In the 1970s it was the authors' (and the editors') liberal policy to consider all available dates regardless of their full substantiation. This policy has been criticized as weakening the conclusions of papers given at the Geological Time-Scale Symposium in Sydney in 1976 (at the International Geological Congress of that year) due to:

(a) lack of statement of errors
(b) failure to use uniform decay constants
(c) inclusion of data with poor stratigraphy and/or poor geochronology.

The reasons for the lack of (a) and (b) is dealt with above for the present topic and can be ascribed to lack of opportunity, time and money for library research. The criticism in (c) is rather subjective as no detailed analysis has been given but benefits mainly from hindsight after 10–20 years of further work. Harland (1983) comments on this but the authors also comment that data used in 1976 at the time of the Sydney 25th International Geological Congress were not then demonstrably bad, the evidence for such a judgement was not available, and is still not in most cases available to the authors to make them *scientifically* demonstrably bad in 1983. They cannot, of course, be given much weight at present because the data and determinations are still incomplete for us. The authors agree with Harland (1983) that it is to be hoped that 'a consensus will increasingly be possible as to what constitutes determinations good in all respects'. Without such an accepted consensus individual determinations should not be lightly dismissed.

The technical equipment, expertise and knowledge continues to improve. Soviet geochronology cannot and should not be neglected. Keller & Krasnobaev (1983) recently published on 'Late Precambrian geochronology of the European USSR, giving the Riphean lower boundary as 1650 ± 50 Ma, the Vendian lower boundary as 650 ± 10 Ma and the Cambrian lower boundary as 590 ± 10 Ma.

Full data are not included by Keller & Krasnobaev in their article in English to avoid over-extending the length of their paper (Keller *pers. comm.*, November 1983) but the data are readily available from the Soviet authors for a scientific assessment.

IGCP Project 196 'Calibration of the Phanerozoic Time Scale' has been supplied with Siberian and Russian Precambrian–Cambrian boundary section samples for glauconite studies. Material from the same sources are being investigated by I. M. Gorokhov in Leningrad using both Rb/Sr and K/Ar methods on glauconite. Semikhatov is preparing a review paper on the global geochronology of the Precambrian–Cambrian boundary with special emphasis on USSR and China (Semikhatov *pers. comm.*, November 1983). So further Soviet results could soon be available using a variety of radiometric methods and the latest apparatus. This will again demand close attention from stratigrapher-geochronologists who are concerned with international geology and its standards.

China

In recent years the late Precambrian Sinian System rocks and the early Cambrian rocks have been intensively studied in

TABLE 7. Yangtze area, Hubei Province, China; late Proterozoic–early Phanerozoic stratigraphy calibrated with isotopic ages in millions of years (Ma). Author code: (1) Ma Quogan *et al.* 1980, (2) Zhang *et al.* 1982, (3) Xing *et al.* 1982, (4) Zhang *et al.* 1984, (5) Xue 1984, (6) Xing *et al.* 1984, (7) Ma Quogan *et al.*, *pers. comm.* 1982, (8) Compston, *pers. comm.* 1983 — ion probe crystallization age, (9) Luo, *pers. comm.* 1984.
It can be assumed with some confidence that, in most cases, the dates given in later papers should be given more weight than those given earlier which were not quoted again.

System	Formation	Member	Isotopic ages (Ma) authors in brackets (see caption) and method
CAMBRIAN	Shuijintuo		(573 ± 7(2) (6) Rb-Sr; 613 ± 23(1) (3) (6) Rb-Sr (568 ± 12(4) (6) U-Pb 573 ± 32(4) U-Pb (572 ± 14(4) Rb-Sr 415–485(2) Rb-Sr (570 ± 4(2) Rb-Sr (574 ± 20(7) Rb-Sr 565–490(2) Rb-Sr
	hiatus at least in places		
	Dengying	Tientzushan	(602 ± 15(2) (4) (6) Rb-Sr (460 ± 9(2) Rb-Sr
			615 ± 20(2) (3) (4) Rb-Sr
SINIAN	Dengying	(Baimatuo) (Shibantan) (Hamojin)	614 ± 18(2) Rb-Sr
	Toushantuo		(700 ± 5(2) Rb-Sr, 580 ± 25(2) Rb-Sr (691 ± 29(7) Rb-Sr, 580–420 (2) Rb-Sr (727 ± 9(2) Rb-Sr, 460–340 (2) Rb-Sr
	Nantuo (with tillites)		
	Datangpo		(728 ± 27(7) Rb-Sr, 500–360 (2) Rb-Sr (608 ± 15(2) Rb-Sr
	Gucheng		
	Lientuo		(740 ± 16(7) U-Pb
PRESINIAN	Huangling Granite		(805(11) K-Ar (823 ± 7(9) K-Ar (806 ± 4(7) (9) Rb-Sr (819 ± 54(1) (2) Rb-Sr (860 ± 50(2) (9) U-Pb (819 ± 6(8) (9)

many parts of the People's Republic of China where rocks below and above the proposed level of the Precambrian–Cambrian Boundary are well exposed. Because of its significance in international and intercontinental correlation special study of that part of the geological column from Sinian tillites to Cambrian trilobites has been undertaken by Chinese geologists. Some of the isotopic age data have been published (Ma Guogan *et al.* 1980; Zhang *et al.* 1982; Compston & Zhang 1983) or is in press (Xing & Luo 1984; Xing *et al.* 1984 Zhang *et al.* 1984; Xue 1984). The recent papers are entirely in English. Earlier ones may have short abstracts in English but the main, detailed, body of results is in Chinese. A brief further summary has been published by Compston & Zhang in 1983.

The data obtained, using Rb-Sr, U-Pb and K-Ar radioisotope methods, were from mainly two centres of research (others are now involved) – Yichang, Hubei Province, China and the Research School of Earth Science, Australian National University in Canberra. The stratigraphic settings of the dated material have the merit of close stratigraphic control in well-studied relatively continuous successions close (or not so far) from the recommended global stratotype for the Precambrian–Cambrian Boundary (Pє-є B). The Precambrian–Cambrian boundary successions are now well known in many parts of China. The faunal assemblage of Tommotian (*sensu-lato*) type (Cowie 1978) is found in these sections (Zones II & III, which are part of the Meishucun stage) giving the earliest Cambrian age now stipulated by the recommendations of the IUGS-IGCP Working Group on the Precambrian–Cambrian Boundary. (see Table 6)

The four main areas are (1) East Yangtze gorges (Hubei Province), (2) Guizhou Province, (3) East Yunnan Province and (4) Sichuan Province which all lie in central or eastern China in inhabited, geologically well-studied terrain.

East Yangtze gorges

This relatively complete succession in the Sinian and Cambrian Systems has in places, however, a large time-gap (without angular unconformity) with possible or probable non-sequence and/or condensed sequences in the Cambrian succession *above* the claimed position of the Precambrian–Cambrian Boundary. The Precambrian rocks are followed conformably by Cambrian rocks. Abundant shelly fossils of Tommotian (*sensu lato*) assemblage (Meishucun Stage) are found here (in the Tientzushan Member), as in all the other three areas above, giving good biostratigraphical control. Trilobites occur in the Shuijintuo Formation indicating an Atdabanian age. In Tables 4 & 7–9 the ages have been calculated using the decay constants proposed by the IUGS Subcommission on Geochronology at the 25th International Geological Congress (Steiger & Jäger 1977). Stratigraphic details and geochronological data are given in Table 7. Complete information relating to the age data are still in process of publication but probably sufficient is already known from the references and personal communications to merit discussion of the resulting calibration of the chronostratigraphy without taking up a definite position regarding its reliability which must await publication of the full details and data.

The most significant date so far known to the authors relating to the Pє-є Boundary is from argillaceous limestone from the upper part of the Dengying Formation (Table 7). Shelly fossils are found in the same subdivision (Unit 6 of the member). According to Compston (*pers. comm.* 1983) the rock is unmetamorphosed and the fine-grained clay fraction (*c.*1.5 microns) has yielded a Rb-Sr isochron age of 602 ± 15 Ma (Ri = 0.7091 ± 0.0003) (about 33 samples). The analysed material was subjected to a comprehensive mineralogical examination (Clauer 1976), no detrital minerals such as kaolinite, 2 M illites and K feldspars were present and the 1 M illite crystallinity indices were high indicating a depositional or early diagenetic origin. Calcareous leachates were extracted to give the sea water ^{87}Sr: ^{86}Sr ratio.

From the Shuijintuo Formation (which yields trilobites including *Hebediscus*) dark shales with lenses of limestone, similarly controlled Rb:Sr studies gave (Compston, *pers. comm.* 1983) Model III isochrons of 572 ± 14 Ma (Ri = 0.7088 ± 0.0008) and 573 ± 7 Ma (Ri = 0.7091 ± 0.0003). The former result was based on analyses made in the Yichang Laboratory, China while the latter age resulted from the analyses of the same samples at the Australian National University, Canberra under the direction of Professor W. Compston. The high crystallinity indices of the analysed 1 M illites > 5.75 allow the conclusion to be drawn that the age of *c.*573 Ma dates the time of completion of early diagenesis when the 1 M illites finished growing *in situ*. The significance of this age is further strengthened by U:Pb isochron ages on whole rock samples of uraniferous black shales which have given $^{207}Pb/_{204Pb}$: $^{231}U/_{204}Pb$ = 573 ± 32 Ma, and $^{206}Pb/_{204Pb}$: $^{238}U/^{204}Pb$ = 568 ± 12 Ma. Uranium is fixed on carbonaceous matter during sedimentation or during early diagenesis, whereas Rb is fixed as illites. Thus two quite different geochemical systems are involved to give ages identical within errors. The results strongly support the interpretation that both are dating the same event — early diagenesis (W. Compston *pers. comm.*): an age of *c.*575 Ma thus probably dates the Quiongzhusi (Atdabanian) Stage, the next stage of the Lower (Early) Cambrian above the Meishucun Stage Zones II & III (Tommotian) (Table 6).

The combination of the U-Pb age corroborating the Rb-Sr age eliminates the possibility that the Rb-Sr age might be too old on account of inherited illite.

Guizhou Province

The Niutitang Formation in Zhijing, Guizhou Province is equivalent to the Shuijintuo Formation of the Hubei Province Yangtze gorges area and gives a Rb-Sr isochron age of 569 ± 12 Ma from black shales (level of mineralogical control is unknown for this age). The stratigraphy is summarized in Table 8.

East Yunnan Province

In Table 9 estimated dates with estimated confidence errors for the member boundaries have been omitted but are given by authors as:

1. Base of Xiaowaitoushan Member 610 ± 10 Ma (Zhang *et al.* 1984).
2. Base of Zhongyicun Member 605 ± 15 Ma (Zhang *et al.* 1984).
3. Mid-Dahai Member 595 ± 15 Ma (Zhang *et al.* 1984).

The Precambrian–Cambrian global stratotype point (as recommended to be adopted) is between points 2 & 3. It must be noted, however, that some of the ages may be too high or too low (see below & Table 7).

The Badaowan Member is well exposed in two sections about 60 km apart which can be closely correlated. At Wangjiawan (about 100 km south of Kunming) it indicates an age of 588 ± 13 Ma and at the locality Meishucun (also south of Kunming) it gives an age of 587 ± 17 Ma. The determinations were on whole rock utilizing a computer with a modern mass spectrometer and come from well above the Precambrian–Cambrian Boundary. In these cases, however, there was no extraction of illite and no crystallinity index is available, the ages may thus be somewhat too high due to the presence of inherited detrital minerals (compare with dates in Table 7).

TABLE 8. Precambrian–Cambrian stratigraphy and calibrating isotopic date, Guizhou Province, China. For authors (4) (6) and (9) see Table 7.

Cambrian Niutitang Formation	569 ± 12 Ma (4) (6) (9) Rb-Sr
hiatus	
in some places only	
Cambrian Dengying Formation (upper part)	
Sinian Dengying Formation (lower part)	

A recent unpublished report from Luo Huilin (Kunming, Yunnan, China) states (without the essential details which will be made available) that an isotopic age of 580 ± 8 Ma has been determined from a horizon at 5 m above the Pє-є Boundary global stratotype point in the Meishucun section (Zone I/II contact) Jinning County, Yunnan. This is given in Table 9.

Other good sections showing probably complete stratigraphic successions from the late Precambrian Sinian System

TABLE 9. Precambrian–Cambrian stratigraphy and calibrated isotopic dates, East Yunnan Province, China. The recommended global stratotype section and point for the Pє-є Boundary is situated in this province at Meishucun, Jinning County at the level shown below.

Periods	Lithostratigraphy		Isotopic dates (Ma) & method & authors (Table 7)	Biostratigraphy — stages & zones & lithostrat. units		
		Yuanshan Member Units 13–15 (72 m)		ZONES IV & V	(Units 13–15)	QIONGZHUSI STAGE
CAMBRIAN	Qiongzhusi (Chiungchussu) Formation	Badaowan Member Units 9–12 (54 m)	588 ± 13(4) (6) RbSr 587 ± 17(4) (6) (9) RbSr 612 ± 36(3) (6) RbSr 603 ± 31 (3) (6) RbSr 580 ± 8(5) (6) Rb-Sr 588 ± 13(4) (5) (9) Rb-Sr 585 ± 15(5) (6) Rb-Sr	ZONE III	(Units 9–12)	MEISHUCUN STAGE
		Dahai Member Unit 8 (1.1 m)		ZONE II	(Units 7 & 8)	
SINIAN–CAMBRIAN BOUNDARY			580 ± 8(9) RbSr PRECAMBRIAN–CAMBRIAN BOUNDARY			
	Yuhucun Formation	Zhongyicun Member Units 3–7 (11.6 m)		ZONE I	(Units 1–6)	
		Xiawaitoushan Member: Units 1 & 2 (8.2 m)				
SINIAN (Precambrian)		Bayanshao Member				DENG-YINGXIA STAGE
		Jiucheng Member				

Zone I : *Anabarites–Circotheca* Zone (A-C)
Zone II : *Paragloborilus–Siphogonuchites* Zone (P-S)
Zone III : *Sinosachites–Eonovotatus* Zone (S-E)
Zone IV : *Parabadiella* Zone (P)
Zone V : *Eoredlichia* Zone (E)

into the Cambrian System with the earliest Cambrian shelly fossil assemblages from above the horizon currently selected for the Precambrian–Cambrian Boundary by the IUGS-IGCP International Working Group are found (a) in Western Sichuan Province (b) South-western Shaanxi Province and (c) Xinjiang Province in NW China. This illustrates the wealth of stratigraphically controlled sections in China across the Precambrian–Cambrian Boundary with numerous radio-isotopic potentialities.

North-west Europe and Africa

Isotopic age determinations of great current interest and relevance come from (Odin *et al.* 1983):

1. France : Vire-Carolles granite, 540 ± 10 Ma
2. Morocco : Anti-Atlas syenite, 534 ± 10 Ma
3. England : Ercall granophyre, 533 ± 12 Ma
   Rushton schists, 536 ± 8 Ma

The Vire-Carolles granite (Pasteels & Doré 1982; Doré 1984)

The Vire-Carolles granite intrudes the Upper Brioverian Slates of the French *Massif Armoricain*. Pasteels & Doré (1982) state that 'no fossils have been found in the Upper Brioverian flysch'. The Vendian microfossils reported from the Brioverian apparently occur in the Middle and/or Lower Brioverian. The Brioverian was folded during the Cadomian phase and is about 6000 m thick and the possibility that it includes unfossiliferous Cambrian as well as fossiliferous Precambrian strata cannot be excluded and is examined below.

The mainland Upper Brioverian does not yield body macrofossils but the offshore island of Jersey exposes rocks which are correlated with the mainland Upper Brioverian on lithological grounds. Duff (1978) postulated a late orogenic (post-533 ± 16 Ma) phase of 'Cadomian' deformation in Jersey. Squire (1973) found trace fossils (e.g. *Sabellarites* in these beds. These trace fossils may, in the opinion of the authors (although some of the information is unpublished and uncertain), indicate a late Precambrian or Cambrian age on palaeontological grounds but *Sabellarites* is known also from the Ordovician system in Canada (Squire 1973; Crimes & Anderson 1984). In central Brittany Brioverian strata have yielded *Planolites*.

Microfossils from the late Proterozoic Brioverian rocks have recently been commented upon by Mansuy & Vidal (1983). Specimens are comparable to late Proterozoic microfossil taxa *Sphaerocongregus* and *Bavlinella* reported from localities in North America, Scandinavia, Greenland and elsewhere. There seems no doubt as to their Precambrian age. The exact stratigraphical level of the rocks yielding these microfossils from France is not given or discussed by Mansuy & Vidal so the possibility that other higher Brioverian strata were still being deposited when the Cambrian period commenced must be considered with impartiality, even though Mansuy & Vidal give a geochronometric age for the Brioverian as 670–640 Ma in the locality of provenance of their fossils. At this locality (Quibou in eastern Brittany) the microfossils came from black siliceous rocks interstratified in a Lower and Middle Brioverian sedimentary sequence which is intruded by plutonic rocks dated at 670 Ma and overlain by volcanics dated at 640 Ma. The age of the Upper Brioverian on the other hand, which is unfossiliferous, seems unknown on either a chronostratigraphic or a chronometric basis and may be partly Cambrian and partly Precambrian. It also seems fair to comment that the whole of the Brioverian could be Vendian (P€) age giving an age for the P€-€ Boundary which may be much less than 540 Ma if time allowance is made for the unconformity at the top of the Brioverian as well as for Vendian times and assuming that 540 Ma is the correct age for the Vire-Carolles granite (see below).

Unconformably overlying the granite on the mainland of France are a Red Conglomerate formation and arkoses which are probably continental and contain no fossils. A *Schistes et calcaires* formation overlying the Red Conglomerate Formation yielded a putative worm which is now identified by Doré (1984) as ?*Coleoloides* sp. This genus is known from the Atdabanian stage but has also been referred to younger horizons of the Cambrian. Brasier *et al.* (1978) comment that 'The value of *Coleoloides* tubes for correlation is as yet untested'. However, the *Schistes et calcaires* may be correlated with the *Coleoloides* Zone of the United Kingdom and Newfoundland (probably Lower Atdabanian).

Two sections in Normandy are of particular importance (Doré 1984):

1. *Carteret.* (i) A fauna of Atdabanian age in the Saint-Jean-de-la-Riviere Formation includes Archaeocyathids, Trilobites (*Bigotina*), *Aldanella, Indianites, Epiphyton, Botomaella, Renalcis* and stromatolites. (ii) Below this two levels occur — a higher with *Allonia, Chancelloria*, and *Eotheca primitiva* and a lower with *Planolites, Skolithos, Phycodes* and *Taphrhelminthopsis*. No base is seen. These two lower faunas belong to the Carteret Formation which Doré interprets as of Tommotian age but in the opinion of the authors might be of Atdabanian age.
2. *Zone Bocaine.* (i) An Atdabanian age fauna of *Circotheca* and *Fordilla* equal to (i) of Carteret. (ii) Below this *Scolicia* and *Cochlichnus*. (iii) Near the base of the cover rocks – *Monomorphichnus, Taphrhelminthopsis, Helminthopsis, Planolites, Phycodes* and ?*Coleoloides*.

These faunas from (ii) and (iii) are assigned by Doré (1984) to the Tommotian age but could in the authors opinion be of Atdabanian age and this would be especially true of *Coleoloides* (if correctly identified) because of arguments presented below. *Coleoloides* could indicate an Atdabanian age.

(iv) below (iii) is an unconformity with below it Upper Brioverian Flysch intruded by the granodiorate of Vire which gives an isotopic date estimated at 540 ± 10 Ma on monazite and a whole rock isochron of 615 ± 10 Ma, both dated by Rb/Sr methods.

The *Coleoloides* Zone was until recently considered to be the basal zone of the Lower Cambrian succession in south-eastern Newfoundland but Bengtson & Fletcher (1983) now recognize two assemblages, the upper one being a *Coleoloides typicalis* assemblage of Atdabanian age with a lower *Aldanella attleborensis* assemblage which is equivalent to the Tommotian stage with a typical content of *Aldanella attleborensis, Heraultipegma* n.sp., and *Fomitchella* cf. *acinaciformis*. If correlatable Massachusetts occurrences are included then this Tommotian *Aldanella attleborensis* assemblage can be taken to include *Lapworthella* n.sp. and *Anabarites tripartitus* while the *Coleoloides typicalis* assemblage contains trilobites as well. Correlations across the Atlantic also support the Atdabanian age of the *Coleoloides typicalis* assemblage and there is no indication that *Coleoloides* spp. occur in beds of Tommotian age.

The marine sediments also contain trace fossils of arthropods (*Monomorphichnus*) and gastropods (*Taphrhelminthopsis*) (Doré 1984) which may be Atdabanian or Tommotian. They are not yet chronostratigraphically diagnostic (Crimes & Anderson 1984) but may be even younger in range as may the heteractinellid sponges and hyolithids. The significance biostratigraphically of these and other trace fossils in the late Proterozoic and early Phanerozoic eons is being studied globally and at present their biostratigraphic zonation has not been established with regard to the correlation of the recommended Precambrian–Cambrian Boundary.

Other Cambrian faunas in this part of France are of uncertain stratigraphical relationship to the strata unconformably overlying the Vire-Carolles granite and there seems to be no certain palaeontological evidence that early Cambrian beds older than the trilobitic Atdabanian strata are present in the whole region or if they are that they can be related to the strata resting on the granite. Further work is required on the stratigraphy and palaeontology: the area is a difficult one geologically.

The age of 540 ± 10 Ma for the Vire-Carolles granite was adopted by Pasteels & Doré (1982) only after considerable discussion of a large body of generally discordant U:Pb, Rb:Sr and K:Ar data and setting aside the alternative date of 615 ± 10 Ma (Rb/Sr whole rock). It is based on the Pb:Pb ages of 'high temperature acid-washed aliquots' of monazites one of which yielded a concordant age of 554 Ma. It is pertinent to note that Pasteels & Doré (1982) themselves comment that 'The significance of the monazite age (540 ± 10 Ma) remains open to discussion in the present case. It cannot be ascertained that it corresponds to the age of the granite emplacement and crystallization, as in other reported cases (e.g. Gulson & Krogh 1973). In any case, monazite, because of its frequent occurrence in Mancellian granites and euhedral shape in some cases, may be regarded as a primary constituent yielding, if not a crystallization age, a closure time. This closure corresponds, according to all available data, to a rather high temperature'.

From the monazite ages in Pasteels & Doré (1982, Table 1) of 550 ± 10 Ma, 542 ± 9 Ma, 547 ± 10 Ma, 543 ± 17 the age of 540 ± 10 Ma is taken by Pasteels & Doré (1982) but there may be disagreement with this (W. Compston, *pers. comm.* 1983). Whether one looks at the monazite data in isolation or at the total body of geochronological data the age of intrusion of this particular igneous body is somewhat ambiguous. The various K:Ar and Rb:Sr determinations strongly suggest Palaeozoic disturbance(s). Although it might be more prudent to adopt the one concordant monazite age — *c*.554 Ma (RG 100.307) — as the best estimate of the age of cooling of this igneous body the total stratigraphic and geochemical uncertainties greatly diminish the value of this body in the definition of the Cambrian time-scale.

The stratigraphic levels in the Normandy successions which are not yet clarified are:

(a) the Atdabanian–Tommotian boundary;

(b) the Tommotian–Vendian boundary (Precambrian–Cambrian Boundary) which could be:

(i) above the unconformity (that is the unconformity between the Brioverian flysch and the cover rocks)

or (ii) below the unconformity

or (iii) at the level of the unconformity.

These questions clearly affect the value and interpretation of the isotopic data of 540 ± 10 from the Vire-Carolles granite in calibrating the chronostratigraphically defined Precambrian–Cambrian Boundary based on a global stratotype section and point (holostratotype).

Considerations of tectonic grade or style or sedimentation are not pertinent in this context – the desired factors in biostratigraphy (which is the guiding principle in defining the position of the Precambrian–Cambrian Boundary) is palaeobiological. In the Normandy successions we have a clearly Atdabanian fauna with trilobites and a clearly Vendian fauna with microfossils at some level in the midst of the Brioverian succession. The Precambrian–Cambrian Boundary is based on a global stratotype section and point (holostratotype).

Considerations of tectonic grade or style or sedimentation are not pertinent in this context – the desired factors in biostratigraphy (which is the guiding principle in defining the position of the Precambrian–Cambrian Boundary) is palaeobiological. In the Normandy successions we have a clearly Atdabanian fauna with trilobites and a clearly Vendian fauna with microfossils at some level in the midst of the Brioverian succession. the Precambrian–Cambrian boundary level lies between. The great progress in study of small shelly fossils, other macro- and microfossils, trace fossils and other palaeobiological evidence in the years since the visit of the Working Group on the Precambrian–Cambrian Boundary in 1974 is manifest. The Normandy succession seems to be the most favourable in continental Europe for the establishment and calibration by chronometry of a regional stratotype of the Pє-є Boundary if further progress can be made.

It is not as yet *proved* stratigraphically to the highest requirements that the intruded Upper Brioverian is Precambrian, it may be Cambrian in its upper part, and in any case, as no characteristic Tommotian fossils are found, the unconformably overlying sediments may be younger than the earliest Cambrian and belong to the Atdabanian Stage. The Vire-Carolles granite may be intruded into Cambrian strata and represent an age above the Precambrian-Cambrian Boundary as may be internationally defined by IUGS.

Anti-Atlas syenite, Morocco

Results of this U-Pb attempted dating of the Precambrian–Cambrian Boundary in Morocco have recently been discussed by Lancelot (1982), Gale (1982) and Odin *et al.* (1983). There appears to be some doubt in the geochronometry (W. Compston, *pers. comm.*[1]). There is no uncontrovertible evidence of Tommotian (early Cambrian) age fossils in this region (Rozanov & Debrenne 1974): the archaeocyathids found are Atdabanian in age and are probably not earliest Atdabanian either (Debrenne & Debrenne 1978). The trilobites found stratigraphically below the archaeocyathids cannot be shown by biostratigraphic considerations to be older than Atdabanian: Tommotian trilobites are not known to occur anywhere (Sdzuy 1978). Putative trilobites claimed from the Tommotian of east Siberia can be discounted as a probable misidentification (Cowie & Rozanov 1983).

In the High Atlas the Bou Ourhioul rhyolite of the Ouarzazate Group has been dated by U-Pb on zircons at 578 ± 15 Ma (Jeúry *et al.* 1974).

[1] Three of the four analysed zircon fractions plot practically in the same place on the Concordia diagram. Thus for all intents and purposes the Discordia line is defined by two points which are very close together. Furthermore the fourth datum point which effectively defines the Discordia line is the least radiogenic fraction and would be most susceptible to any error in the common lead correction.

TABLE 10. Schematic succession in Morocco

		(*Schisto-calcaire* series ((	Archaeocyathids and trilobites
and/or	Atdabanian ?Tommotian	(*Calcaires superieur* series	Archaeocyathids and trilobites in upper part
and/or	?Tommotian ?Vendian	*Lie de vin* Series (Adoudounian)	Stromatolites
	Vendian	*Calcaire inferieur* Series (Adoudounian)	
	Vendian	Precambrian III (Ouarzazate volcano-sedimentary group)	

The *Lie de vin* series of the Adoudounian Stage contains stromatolites which could well indicate a Precambrian age (probably Vendian) according to Schmitt (1978 & *pers. comm.* 1983) and Choubert (1984). Bertrand-Sarfati (1981) disagreed with Schmitt's earlier work when he identified and described the stromatolites and suggested a Tommotian age but Schmitt verbally disagreed with Sarfati's 1981 conclusions in 1983. Many biostratigraphers do not favour the use of stromatolites to differentiate late Precambrian from early Cambrian strata and as there are no other grounds in biostratigraphy available in the case of the *Lie de vin* Series it can be Vendian and/or Tommotian in age. Lancelot (1983) and Doré (1983) stated 'Adoudounian is a rather informal equivalent of the Vendian.'

The latest opinion of Choubert (1984), a geologist working in Morocco, is that the Jiucheng Member of the Yuhucun Formation (Precambrian Sinian system of Yunnan Province, China) with its *Vendotaenia, Chuaria* (?) and acritarchs (Table 9) can be correlated with the marine beds of the *Serie Lie de vin* of Morocco with its stromatolites such as *Linella avis* (*Parmites*), *Tungussia inna, Tifounkeia* which indicate a Vendian age. The *Calcaires superieur* Series of Morocco above do not contain 'small shelly fossils' but display stromatolites such as *Aciciella angepina* which is known from beds assigned to the Lower Cambrian in Australia but where their Tommotian or Atdabanian age is not substantiated.

The association of stromatolites with diagnostic Tommotian–Meishucun small shelly fossil assemblages has not been recorded or established.

Although Choubert's interpretation of *both* the Jbel Boho syenite and the trachytic lavas occurrences as volcanic is commonly accepted the suggestion is made here that the possibility of the dated syenite (534 ± 10 Ma) being intruded into the lavas should be re-examined in case it occurred later than previously thought. There seems to be no published evidence that the syenite is actually in contact with sedimentary strata and that these strata conformably or unconformably overly the syenite. According to Choubert (1952) field observations suggested that the syenite represents a feeder to the trachyte lavas and that the syenite and trachyte are more or less contemporaneous; such evidence is, however, notoriously difficult to demonstrate. A further visit could be worthwhile.

The absence of the earliest Cambrian (Tommotian) faunas in Morocco, on results reported to date, mean the position of the Precambrian–Cambrian Boundary in the geological column in this region on the basis of stratigraphical criteria is poorly defined and controversial. The 534 ± 10 Ma date may not be bracketed stratigraphically between Precambrian and early Cambrian series and does not unequivocally delineate the age of the Precambrian–Cambrian Boundary as claimed by Gale (1982). The Moroccan section is indeed one of the best sections spanning latest Precambrian and Early Cambrian times but the absence of diagnostic fossil evidence at the critical levels which could be equivalent to the Tommotian–Meishucun (II & III) Stages is a serious disadvantage. Further discoveries and field work could elucidate the situation in this relatively accessible region. If the Jbel Boho syenite with its recommended age of 534 ± 10 Ma is emplaced no higher than the *Calcaire inferieur* Series (Precambrian: Vendian) and its associated trachytes are interbedded with that series then the Pє-є Boundary could be much younger than 530 Ma leaving a very short duration of only 10–15 million years for the Cambrian period with little time for the many Cambrian events (including biological evolutionary changes) which are known to have taken place.

Ercall granophyre (533 ± 12 Ma) and Rushton schist (536 ± 8 Ma & 667 ± 20 Ma)

The Ercall granophyre (533 ± 12 Ma) is a small (800 m × 300 m maximum dimensions) intrusion which is probably faulted against Permian rocks on its north-western margin and against Uriconian tuff and agglomerate on its northern margin (Geol. Surv. GB map compiled by Hains 1978).

A more accurate determination of the Rb-Sr isotopic age of the granophyre has recently been made (Beckinsale *et al.* 1984) and is given above. Further research in Canberra, Australia with U-Pb analyses seems to corroborate the date (W. Compston *et al., pers. comm.* 1984). According to a preliminary abstract from Compston and his colleagues these U-Pb analyses have been made on two multi-grain samples of zircon by isotope dilution and on 15 single zircon grains by ion microprobe. $^{207}Pb/^{206}Pb$ ages agree at 565 ± 7 Ma — this older date may be spurious due to a content of older zircon xenocrysts while most of the other grains appear to be concordant at 531 ± 5 Ma (2σ). 531 ± 5 Ma can therefore be interpreted as a well-substantiated isotopic age for the Ercall granophyre. Rb-Sr dates in close agreement with U-Pb dates on the same rock body have a special value and this applies not only to the Ercall granophyre radiometric dates but also to the similarly substantiated Cambrian dates from the Shuijintuo Formation of the Yangtze area, Hubei Province, China.

The south-eastern margin of the granophyre abuts against the Wrekin Quartzite: in the present state of paucity of outcrops the contact of the granophyre with the quartzites is little exposed. Published records and maps are equivocal but it has been a common assumption that the Wrekin Quartzite is unconformable on the granophyre. This seems to be an error partly based on an incorrect chronostratigraphic age

TABLE 11. Precambrian–Cambrian stratigraphy, Shropshire, England

Cambrian	Lower Comley Sandstones and Limestones Wrekin Quartzite	up to 200 m thick
unconformity		
Longmyndian (Precambrian)	Wentnor Group: sedimentary rocks Stretton Group: sedimentary rocks	up to 6000 m thick
Uriconian (Precambrian)	Gneisses, Schists, Andesite, Tuff, Agglomerate, Rhyolite	
Intrusive igneous rocks	1. Felsite 2. Granophyre (Ercallite)	

assignation of the Ercall granophyre as contemporaneous with the Uriconian rocks. Elsewhere in Shropshire there are good grounds for assuming an unconformable relationship between Lower Cambrian strata and the Uriconian. The contemporaneity of the Ercall granophyre and the Uriconian volcanics is probably not a valid assumption.

The local rocks and succession are usually accepted as follows (Table 11):

In England there are many non-sequences and angular unconformities in rocks ranging from putative late Precambrian to undoubted Lower Cambrian strata with trilobites. A summary of Lower Cambrian transgressions and related factors has been published hy Brasier (1980).

The Uriconian seemed for long to be undoubtedly Precambrian on the time-scales accepted in the past and particularly as the provenance of three samples of rhyolites from the Wrekin and their derived dates were taken as reliable, ranging from 677 ± 72 Ma to 632 ± 32 Ma (Fitch *et al.* 1969). Baker (1971) cast some doubt and Patchett *et al.* (1980) disagreed entirely with these results by Fitch *et al.* given in 1969 and substituted new estimates of 540–560 Ma on the basis of Rb-Sr dates which tbey claimed were more accurate.

The Longmyndian was coloured as part of the Cambrian System in early British Geological Survey maps. Stubblefield (1956: Fig. 1 and p. 5), in his diagram to illustrate history of research, shows Longmyndian as Cambrian by the authority of various authors until 1873 when both Hicks and Sedgwick still referred the unit to Lower Cambrian. Lapworth in 1898 placed it in the Precambrian on grounds which might not be accepted without question today. In the Cwms (2.5 kms east of Church Stretton) the basal Wrekin Quartzite appeared in earlier times to lie upon some purple beds (faulted between Eastern Uriconian rocks) which were thought to be equivalent to the Western Longmyndian stratigraphic unit. These outcrops are no longer visible and records of excavations are uncertain/unknown. Purple arkose at the base of the Hartshill Formation and at the base of the Lickey Hill Quartzites suggests that if and when seen unfaulted the base of the Wrekin Quartzites could be arkosic and purple. The previously reported purple beds in the Cwms may thus have been basal Wrekin Quartzite. It seems as if no contact of Wrekin Quartzite or other undoubted Cambrian strata with Longmyndian strata has been seen without doubt. Otherwise no rocks older than Upper Ordovician rest directly on the Longmyndian (Pocock *et al.* 1938). Attempts have been made to correlate the Longmyndian Wentnor Series rocks with the Torridonian rocks of NW Scotland on lithological, sedimentological and magnetostratigraphic grounds. Greig *et al.* (1968) seemed to agree with this correlation on magnetostratigraphic grounds. Radiometric dating of the Torridonian suggests a minimum age of 800 Ma and it may be more. It seems probable that both Torridonian sedimentaries and Monian metamorphics may well he older than the Longmyndian. Recent dates on rocks from the Longmynd (Bath 1974), from illitic shales of the Strettonian Series, give well-defined isochrons of 452 ± 31 Ma and 529 ± 6 Ma. Bath interpreted these determinations as dating the final movements of ions by pore waters and that, by extrapolation of $^{87}Sr/^{86}Sr$ ratios back to reasonable (assumed) initial values, the time of deposition was likely to be about 600 Ma and could represent Cambrian or very late Precambrian. He further suggested that the deformation of the Longmyndian took place very soon after deposition and the putative Cambrian and younger isochrons may indicate subsequent mild metamorpohic events. If the Ercall granophyre is not in fact part of the Uriconian then this interpretation does ameliorate the problem of its relationship with the Longmyndian.

According to Beckinsale *et al.* (1984c) the geochemical data demonstrate that the Ercall granophyre cannot be simply an intrusive equivalent to the Uriconian rhyolites; also the chronostratigraphic age of the Longmyndian could be elucidated by new micropalaeontological work now being undertaken by the British Geological Survey (Beckinsale *pers. comm.* 1984). Recently published results on microfossils from the Longmyndian succession (Peat 1984) describe and illustrate nematomorph cryptarch assemblages (trace fossils) which appear not to conflict with the author's assumed Precambrian age for the Longmyndian strata.

No Ercall granophyre pebbles have been found in, or associated with, the Wrekin quartzite beds. At the present stage of research it seems reasonable to assume that the Ercall granophyre is not of the same age as the Uriconian rocks and there is now agreement with the author's suggestion in 1982 that the Ercall granophyre is intrusive (and faulted) into the Wrekin Quartzite and the Uriconian rocks.

The geological setting of the Rushton Schist is different from the rocks of the Ercall and the Wrekin hills; it is extremely poorly exposed in low ground to the east and on present outcrops the relationship with the Lower Cambrian sandstones/quartzites is not at all clear in this much faulted region. How it is physically related in the field to the Uriconian, Longmyndian and Ercall granophyre rocks is not clearly established but records indicate it is unconformably overlain by the Wrekin Quartzite although at present previous evidence is obscured by rubbish dumping.

It can be postulated that the Rushton Schist date of 536 ± 8 Ma (as against what is considered to be its true age of 667 ± 20 Ma) may be useful as evidence that this younger Rushton Schists date obtained on the eastern side of the Wrekin Fault was reset by the Ercall granophyre intrusion on the western side of the Wrekin Fault. This possibility, which assumes that no other thermal resetting influence which operated is at present hidden beneath younger rocks, suggests that no major transcurrent movements on the Wrekin Fault (≡ Church Stretton Fault) have occurred since the intrusion of the Ercall granophyre about 533 ± 12 Ma before the present (Beckinsale *et al.* 1984).

The thermal resetting of the Rushton schist date to 536 ± 8 Ma from 667 ± 20 Ma (Beckinsale *et al.* 1984c) seems to inevitably raise the entire question of possible thermal resetting of English and Welsh dates in this part of geological time by igneous activity such as the Ercall granophyre (533 ± 12 Ma).

The Wrekin Quartzite formation:

1. contains trace fossils (*Diplocraterion*, Brasier & Hewitt 1979) but the only body fossil (found by Odin) is a fragment of a horny brachiopod (according to A. W. A. Rushton, *pers. comm.*): it can be correlated lithologically and in stratigraphical position with the Malvern Quartzite formation which occurs about 100 km to the south of the Wrekin. The Malvern Quartzite contains '*Obolella*' *groomi* and *Camenella baltica* typical of the basal beds of the Home Farm Member of Nuneaton (Brasier & Hewitt 1981) which may be near the Tommotian–Atdabanian boundary but are most probahly of Lower Atdabanian age.
2. is overlain conformably by the Comley Sandstones and Limestones formation which gives abundant fossils of Early Cambrian age at certain horizons above the base which indicate an Atdabanian age.
3. correlates lithologically with the Hartshill Quartzite to the east of Shropshire in the Midlands of England but the Wrekin Quartzite may be younger (Brasier *et al.* 1978; Brasier *et al.* 1981; Brasier 1982).

Diachronism in these arenaceous deposits must be considered as a factor in correlation schemes.

The Hartshill Quartzite Formation of the Nuneaton region in the Midlands of England is of special importance because of its significance in biostratigraphic correlations with other parts of the world. In 1978 Brasier *et al.* correlated the lower part of the Hartshill Formation with the top of the Tommotian Stage of the East Siberian Cambrian but Matthews & Cowie in 1979 cast doubt on this correlation. In 1981, however, Brasier revised his estimate of the age of the lower part of the Hartshill Quartzite on the basis of further palaeontological and palaeoecological studies. There now seems to be evidence only for a correlation with certainty to the Atdabanian Stage with some chance of late Tommotian and less or none of Vendian. Both the low-diversity trace fossil assemblage and the body fossil assemblage may be condensed and/or transported and/or reworked and are probably diachronous facies faunas.

Brasier (*in* Cowie 1982) stated that the 'earliest English shelly faunas may be approximately of *Schmidtiellus mickwitzi* or *Fallotaspis* age': these trilobites indicate an Atdabanian age.

From these dates in Shropshire of the Ercall granophyre and the Rushton Schist (533–536 Ma) taken together with evidence from the English Midlands (the Nuneaton diorites and Charnwood Forest igneous relationships with soft-bodied metazoan fossils of Ediacaran/Vendian age) (Cribb 1975) it can be claimed that there is strong evidence that there was an episode of igneous activity in England between at least 530 and 550 million years ago.

Cambrian rocks in the whole region may be no older than Atdabanian or just possibly the youngest part of the Tommotian.

The only data available at present for the geochronometric duration of the Cambrian part of the Meishucun Stage (Zones II & III) are from China (modern dates are not available for the probably equivalent Tommotian stage of Siberia). Estimation of the radioisotope dates for the lower and upper boundaries of this earliest part of the Cambrian period place them at about 595 Ma and about 575 Ma giving a duration of about 20 Ma (Tables 7 & 9). This figure may seem excessive to many (including Snelling 1982) when it is related to the thickness of strata known (although this varies) but there could be time unrepresented by strata due to sedimentational hiatuses.

The Ercall granophyre (533 ± 12 Ma) can only be seen in outcrop at present to intrude and postdate the upper part of the up to 50 m regional thickness of the Wrekin Quartzite (late Tommotian–early Atdabanian in chronostratic age). Before erosion took place at a higher level above the present hill the granophyre probably then cut younger Cambrian beds then present due to depositional or tectonic dip of the strata. The extra height above the present summit for this to have occurred need only be of the order of 60 m. It is also possible that today the subterranean intrusive margins of the granophyre could cut the Atdabanian trilobitic upper part of the Comley Series or even, possibly, the Medial Cambrian. Distances involved are quite small. The youngest Cambrian outcrops, at the junction with the Ordovician Tremadoc Series, are at present only about 600 m eastwards horizontally from the eastern intrusive margin of the Ercall granophyre. The presence of a camptonite intruded into the Tremadoc Series at Maddocks Hill (outcropping about 900 m east of the eastern margin of the Ercall granophyre) is also of interest — other igneous bodies may be hidden in the region.

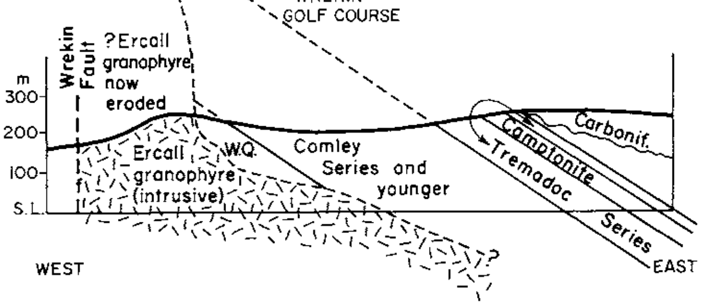

FIG. 1. Geological relationships in the Ercall-Maddocks Hill area of Shropshire, England. Vertical scale = horizontal scale at 1:10,560. WQ = Wrekin Quartzite Formation.

A speculative estimate in numerical terms for the calibration of the chronostratigraphic scale on current British evidence could be as follows:

1. Relying only on present observable outcrops the Ercall granophyre (533 ± 12 Ma) is seen to cut some horizon in the Wrekin Quartzite not far from the top of the maximum known thickness elsewhere of 50 m, levels which are stratigraphically Atdabanian in age. The Atdabanian Comley Sandstone outcrops only 50–70 m from the eastern margin of the intrusive Ercall granophyre and has a regional dip to the east of 35° (Beckinsale *et al.* 1984c). This could therefore mean that the Ercall granophyre was, on present outcrops, intrusive into possibly the latest Tommotian, or probably earliest trilobitic Atdabanian stages of the Comley Series. This gives a minimum chronometric calibration age for the latest Tommotian or earliest Atdabanian of 533 ± 12 Ma. Assuming from Chinese dates that the total Tommotian Stage (?≡Zone II & III of the Meishucun Stage) had a duration of 20 ma (Tables 7 & 9) then the Precambrian–Cambrian Boundary could be, say, 22 Ma earlier than the Ercall granophyre dates, i.e. 555 Ma. The duration of the earliest Cambrian stage is, of course, debatable.

2. If the eroded higher levels of the Ercall granophyre had in the past intruded into the previously overlying main part of the trilobitic Atdabanian Comley Series then the Pϵ-ϵ Boundary date could be adjusted by, possibly, a total of 25 Ma added to the Ercall granophyre date. This speculative estimation gives a date of 558 Ma for the Precambrian–Cambrian Boundary. Intrusion into Middle Cambrian or later, which is speculative, but not impossible, would give a mid-point age for the Precambrian–Cambrian Boundary of, say, 565 Ma. The aerial extension of these hypotheses though probable can only be speculative, the subterranean situation, however, could be tested by boreholes or geophysics.

Patchett *et al.* (1980) gave a Rb-Sr whole rock isochron age of 558 ± 16 Ma from felsic tuff samples from the Eastern Uriconian volcanics exposed in Leaton Quarry (about 3 kms north-east of the Ercall hill on the western side of the Wrekin Fault). This can be taken as a lower side of the bracket to date the Precambrian–Cambrian Boundary, i.e. from 542 Ma to 574 Ma allowing for the date's errors.

If the above arguments were correct the 6000 m thick Longmyndian may be narrowed in duration to 3 Ma or less *if* it was post-Uriconian and pre-Cambrian as has been often postulated.

Beckinsale *et al.* (1984a, 1984b) reported new Rb-Sr dates of 549 ± 19 Ma from the Sarn igneous complex intrusion in North Wales and 542 ± 12 Ma from the Parwyd gneisses in the same region, both these rock suites are considered to be part of the Mona Complex in North Wales which is reported to include Cambrian rocks with fossils (Downie 1975; Muir *et al.* 1979). It seems quite unproven that the Mona Complex is a basement to the Cambrian basin in the whole of North Wales: the evidence is lacking.

Both the Sarn igneous complex and the Parwyd gneisses can be considered to be regionally unconformably overlain by Arenig strata of Ordovician age and are consequently pre-Arenig. Whether these igneous and metamorphic rocks were formed or laid down or metamorphosed in Cambrian or Precambrian times seems unprovable at present even invoking petrogenetic and structural arguments. The chronostratigraphic position of the Coedana granite of Anglesey with its age of 603 ± 34 Ma (Beckinsale & Thorpe 1979) and other ages for rocks for putative inclusion in the Mona Complex — 542 ± 17 Ma — may reflect igneous and metamorphic events contemporaneous with sedimentation in North Wales in Cambrian times (Beckinsale *et al.* 1984b). It can be further suggested that the Cambrian rocks of the Mona complex are older than any of the other Cambrian rocks of the main crop in North Wales (including the Harlech Dome sedimentary rocks) but the position of the Arvonian volcanics below the Cambrian grits in the borehole at the centre of the Harlech Dome raises other problems. It seems a hazardous assumption to make bracketing calibrations of the Precambrian–Cambrian boundary on the basis of collations of radioisotope dates from contrasting facies, basins and igneous situations from Nuneaton through Shropshire to North Wales–Anglesey (Beckinsale *et al.* 1984 a, b, c). The chronostratigraphic position of the Precambrian–Cambrian boundary in these regions is basically controversial and unknown.

It can be emphasized, in agreement with Beckinsale (1984) that, like the position in Normandy, Morocco and elsewhere in the world, the biostratigraphic control is still inadequate — and using similar criteria to the above it is possible that both the Uriconian volcanic rocks and the Longmyndian sedimentary rocks may belong to the Cambrian period if new evidence, with palaeobiological or other correlations in chronostratigraphy becomes available. Alternatively finds of Precambrian diagnostic fossils could place the rocks of provenance unequivocally below the Precambrian–Cambrian Boundary.

The Middle East: The Arabian–Nubian massif, Sinai and Dead Sea

The crystalline basement of the Arabian–Nubian mass is overlain unconformably by a sedimentary cover: the oldest part of the cover consists of continental sandstones with intercalated marine beds which yield fossils of Early Cambrian age as defined in chronostratigraphic terms. The chronometric Rb-Sr age determinations suggested to Bielski (1982) that the Lower Cambrian section may be younger than 530 Ma.

A thermal event at 530–540 Ma affected the basement rocks which are truncated by a regional unconformity. An erosional surface (a peneplain) developed on which the fluviatile Nubian Sandstone was deposited; intercalated marine shales and dolomites contain trilobites in southern Israel and in Jordan, along the line of the Gulf of Elat and the Dead Sea.

Bielski (1982, Table I) correlates these faunas with those of Morocco. The basement ranges in age from 800–580 Ma. Volcanics cutting and unconformably overlying the basement rocks were sampled and gave a Rb-Sr whole rock isochron of 548 ± 5 Ma which was interpreted as an age of extrusion.

A younger granite (Mandar Granite of southern Sinai) intrudes into granites of 600–580 Ma age and itself gives an age of 529 ± 9 Ma. This younger granite is not visibly related to the unconformity: 'the peneplain in this region is missing' (Bielski 1982). It may therefore be Cambrian on chronostratigraphic grounds; there is no certainty that the interpretation by Bielski that the peneplain has been destroyed by the erosion of 1000 m to 2000 m of rocks overlying the Mandar granite is correct. The thicknessess of strata assumed to have been there are great and so are the distances from observable peneplanation. The chronostratigraphic age of the Mandar granite's intrusion is hypothetical.

No thermal effects have been found in the sediments above the peneplain: the authors would question if these effects would be detectable anyway in the predominantly sandstone strata.

The biostratigraphic correlation of the Lower Cambrian of the Middle East with Morocco is not precise in either the Dead Sea or in the southern Negev; the latter is assumed to be the younger by Bielski (1982, Table I). On the basis of the trilobite fauna listed, the Lower Cambrian strata must be assigned to the Atdabanian Stage and may not belong to the oldest part of the stage.

It seems, therefore, that the only stratigraphically possible date for the regional base of the Cambrian is the date of 548 ± 5 Ma which gives a maximum age for the lower part of the Atdabanian Stage. Allowing for the possible duration of the basal Cambrian stage plus, possibly, a part of the lowest Atdabanian Stage) a rough estimate for the age of the Precambrian–Cambrian Boundary instant could here be 565 Ma or more.

North America

Holyrood granite, S. Newfoundland, Canada

Gale (1982) gives the recomputed date for the Holyrood granite as 585 ± 15 Ma. The Holyrood granite is overlain unconformably by Lower Cambrian strata and *Hyolithes* and *Coleoloides* were collected a few metres stratigraphically above the nonconformity (Lambert 1971). These fossils could represent the Atdabanian Stage. The age of the earliest Lower Cambrian strata here could, on current biostratigraphical arguments, be much younger than the Precambrian–Cambrian Boundary instant. It would seem that the *maximum* age of the Vendian–Tommotian boundary could be 600 Ma (585 + 15) on this isotope dating and no reliable estimate of the *minimum* age is permissible on the present evidence. The Conception Group is believed to be younger than the granite and this sedimentary unit is usually considered to be Vendian and some way below the Vendian–Cambrian boundary in age. There seems, however, to be some ambiguity in the historical geology and regional mapping.

Hoppin Hill granite complex, Massachusetts, USA

Gale (1982) gives the recomputed date for the Hoppin Hill granite as 553 ± 10 Ma and claims it as a maximum age (563 Ma) for the Precambrian–Cambrian Boundary. The Hoppin Hill granite-gneiss is overlain by unmetamorphosed Lower Cambrian slates containing *Obolella* from beds above about 4 metres of basal quartzite. The isotopic date has been considered to be too low in age due to weathering and should be > 553 Ma (Lambert 1971). *Obolella* probably gives a biostratigraphic indication of Atdabanian Stage. In Comley, Newfoundland and elsewhere *Obolella* sp. (*sensu strictu*) seem to appear in the *Callavia* Zone (late Atdabanian).

These two dates from North America are not closely definitive for the Precambrian–Cambrian Boundary but could be indicating a time between 575 and 560 Ma.

Burin Peninsula, Newfoundland, Canada

Isotopic dates from volcanics underlying early Cambrian in the Burin Peninsula of Newfoundland have been reported in abstract (Krogh *et al.* 1983) and require published details to be made available for their full assessment. They are U-Pb zircon dates and are reported to range from $623^{+1.9}_{-1.7}$ Ma to $606^{+3.7}_{-2.9}$ Ma. There is no doubt that the overlying earliest Cambrian with its *Aldanella attleborensis* assemblage (Bengtson & Fletcher 1983) correlates with the Tommotian Stage of Siberia.

The Cambrian period

Apart from the putative dates for the lower boundary of the Cambrian period discussed above the only dates which can be quoted are a computed age of 540 ± 14 Ma for the Middle Cambrian St. David's Series from the United Kingdom (Harland *et al.* 1982) and a fission track date from the Middle Cambrian of the USA (A. R. Palmer, *pers. comm.* 1983) of ~ 540 Ma. Both these dates are not yet fully substantiated (Beckinsale 1984 *et al.* a, b, c).

It can be noted that the Geological Society of America's *ad hoc* Time Scale Advisory Committee to 'encourage uniformity ... in the citation of numerical ages for chronostratigraphic units of the geologic time scale' has established a Decade of North American Geology 1983 Geologic Time Scale (Palmer 1983). For the Cambrian Period the Precambrian–Cambrian Boundary is given as 570 Ma, the Early–Middle Cambrian Boundary as 540 Ma, the Middle–Late Cambrian Boundary as 523 Ma and the Cambrian–Ordovician (Trempealeauan–Tremadocian) Boundary as 505 Ma.

It may be worth commenting that undue and premature reliance on geochronometry in calibrating late Precambrian and Cambrian chronostratigraphy should be avoided: this particularly applies to the use of graphical plots of factors involving geochronometry, chronostratigraphy, biostratigraphy, palaeobiology and palaeomagnetism.

Some examples are: (i) faunal diversity curves using isotopic dates (Sepkoski 1978); (ii) hard ground evolution through geological periods (Palmer 1982); (iii) magnetostratigraphy relating to tectonic plate movements (Kirschvink 1978); (iv) the opening of oceans (e.g. Iapetus) (Wright 1974; Anderton 1980).

Conclusions

1. The age of the Precambrian–Cambrian Boundary calibrated in millions of years still has a wide range of uncertainty. At present the possible range could be 530 to 600 Ma before present.
2. The figures for the possible age for the Pє-є Boundary extracted and summarized above are by no means comprehensive and also are, perforce, drawn from only a few scattered points on the surface of the globe: only a small sample from the many uncalibrated outcrops of late Precambrian–early Cambrian critical stratigraphic levels have, ideally, the latest Precambrian and earliest Cambrian strata containing fossils (and magnetostratigraphic correlations) in maritime Canada, north-west Canada and Alaska, south-west USA and Mexico, Tierra del Fuego of Argentina and Chile, Australia, eastern and northern Siberia, Kazakhstan, Mongolia, China, northern India, Iran, the Middle East, eastern and western Europe, Greenland and other areas. Without new reliable dates there can be no global consensus on calibration of the chronostratigraphic scale by chronometry near the Precambrian–Cambrian Boundary and indeed in the Cambrian period.
3. Diachronism in the arrival in regional successions of the

earliest Cambrian (e.g. Tommotian–Meishucun Zones II & III) faunas cannot at present be proved or disproved but is a distinct possibility: a tentative suggestion made already by one of the authors (Cowie 1964).

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Radiometric dating of Late Precambrian times

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Introduction

Between about 1979 and 1981 new radiometric data relevant to the numerical age of the Precambrian–Cambrian boundary became available. Based on this it was suggested in a recent review (Odin *et al.* 1983) that the numerical age for that boundary was probably much younger than usually accepted, lying probably in the interval of time 520–540 Ma rather than at about 580–600 Ma. This new estimate has provoked discussions which essentially concern the stratigraphical location appropriate to these modern dates. In addition, it has been argued that this new view of the location of the Precambrian–Cambrian boundary did not take into account all the available geochronological results. We consider this last remark as unfounded at the time when our review was submitted for publication (at the beginning of 1982). However, one result of our review seems to have been to rejuvenate interest in the subject so that there is now more data to be discussed. Although the scarcity of fully published results still permits one at present only to propose tentative conclusions, some facts seem well enough established; conversely, in our opinion, other data are certainly not as soundly based as some authors aver. The present comment tries to give complementary information to enable others to judge both our own arguments as well as those of our colleagues.

A fundamental new contribution to this problem is in progress, i.e. a formal definition of the Precambrian–Cambrian boundary within the Meishucun stage of China (apparently now strongly advocated by the IUGS-IGCP Working Group voting members, but not yet ratified by the International Commission of Stratigraphy of IUGS). Given this new advance and the fact that there has so far been no analysis of the stratigraphic relations between this possible new definition and the outcrops that have been radiometrically dated, except in China, it might appear that we could hardly present in this comment any definitive number to characterize the Precambrian–Cambrian boundary. In spite of this limitation, we consider it valuable to examine critically some of the geochronological and stratigraphic data recently used in the assessment of this part of the time-scale.

We will re-examine first the data taken into account in Odin *et al.* (1983) for which recent discussions have given new geochemical or stratigraphic information. We will give later our position on old or new data presently not available in detail but sufficiently often quoted as to need discussion.

The data from the Massif Armoricain

We have difficulty in accepting some of the geochemical arguments presented against the use of the data from the Massif Armoricain for numerical calibration of the Precambrian–Cambrian boundary.

Let us consider first the radiometric data. J. Cowie remarked at the Ist annual meeting of the IGCP Project 196, London, in November 1983, that the apparent age of the monazites of the granodiorite of Vire (540 ± 10 Ma, Pasteels & Doré, 1982, NDS 121) may have been rejuvenated. This seems a very uncertain hypothesis in the light of recent results.

Since the publication of the paper by Odin *et al.* (1983), another age of 540 Ma appeared in the literature concerning the Massif Armoricain. Peucat in his published thesis (1982) gives new U-Pb and Rb-Sr results obtained from an outcrop 50 km away from the granodiorite dated by Pasteels. The age of 540 Ma appears to be that of the metamorphism of the migmatites of Saint-Malo. The details are as follows:

- 542 ± 62 Ma: whole-rock Rb-Sr isochron from anatectite granites.
- 535 ± 5 Ma: U-Pb cooling age or crystallization age of the monazites from the same granites.
- 541 ± 5 Ma: U-Pb crystallization age of zircons from a granitic vein contemporaneous with anatexis.
- 555 ± 25 Ma: U-Pb age of sphenes from the same vein.
- 536 ± 14 Ma: loss of Pb from detrital zircons (2100 Ma old) from a paragneiss (U-Pb).

According to Peucat 'l'âge de 541 ± 5 Ma peut être retenu comme l'âge du métamorphisme pluri facial et de l'anatexie survenue dans le Massif de Saint-Malo'.

These results agree perfectly with the age of the monazites of the granodiorite of Vire collected at Villedieu. In this outcrop, the granodiorite does not present the common facies of that granodiorite but a whiter one, richer in muscovite. Although this facies of Villedieu, somewhat hydrothermal, was produced at the same time as the end of the metamorphic evolution of the Massif of Saint-Malo, it must be emphasized that the regional 'Cambrian' sediments are *absolutely not affected*. Thus, the monazite apparent ages, fully quoted and appropriately discussed in Pasteels & Doré (1982), correspond to an upper limit for the age of the above deposits. They must be younger than 530 to 550 Ma, an interval of time which, if not the crystallization age, is the closure time of monazite at a rather high temperature.

Let us consider now the stratigraphical data and first the age of the Brioverian. It was also suggested that it is not proved that the late Brioverian sediments intruded by the granodiorite of Vire are not Cambrian. This is to ignore the clear faunal and floral opposition, in Armorica, between the Brioverian sediments and the Cambrian sediments (i.e. 'Cambrian' *sensu lato*).

The Armorican Brioverian contains a microflora (Cyanophyta) recently described by Mansuy (1983). The observed taxons (*Sphaerocongregus, Palaeocryptidium, Bavlinella, Favosphaera*) have no actual stratigraphic value in themselves; however, the characteristics of this microflora seem coherent with those obtained from the late Proterozoic series from Northern Europe (Mansuy 1983; Mansuy & Vidal 1983). Track fossils are extremely rare in the late Brioverian sediments. The atypical burrow found by Squire (1973) in the late Brioverian from Jersey and related to *Sabellarites* is not considered by Odin *et al.* (1983) as a fossil useful for stratigraphical calibration, but as an exceptional trace of life of a metazoan in the Upper Brioverian. It has been suggested by some that the age of the post-Cadomian, post-granitic cover is possibly of Atdabanian age and not of Tommotian age because it does not yield characteristic fossils. If we extrapolated this kind of argument to the levels below, we would be led to a nice sophism: the very old Icartian gneiss could be of Atdabanian age! Other authors, particularly Cowie & Rozanov (1983) make completely different deductions from comparable arguments: on page 136 of this paper and concerning the bed 8 of the Yudoma formation at Ulakhan-Sulugur (East Siberia) these authors consider that the absence of the rich Tommotian faunal assemblage is not mainly due to an inappropriate facies but reflects the evolution of life at the Precambrian-Cambrian boundary. Why cannot it be the same in the Massif Armoricain?

Returning to the Cambrian of Normandy the following observations are relevant:

1. In the Western 'zone bocaine', above a granodiorite and the over-lying azoic arkose and conglomerates, the terrigenous sediments have shown trace fossils; *Monomorphichnus*, *Planolites* and *Helminthopsis* have been observed by F. Doré. These traces are described in the recent ichnological literature as soon as late Vendian times, or in rocks for which one questions the age to be Vendian or Tommotian. The 'fossils' themselves are very rare in that area in the basal layers of the 'Cambrian' and they are limited to tubes related to worms (*Coleoloides*). To reach well characterized Cambrian it is necessary to go up well above in the series.
2. In the Central 'zone bocaine' (section of the Orne River valley at Saint-Rémy), the fossiliferous 'Cambrian' with *Fordilla*, hyolithids, only appears with the 'Schistes du pont de la Mousse' formation. Obviously, that could be Atdabanian or Tommotian as well.

Below, there are nearly a thousand metres of strata where life is represented by colleniform stromatolites (no stratigraphical meaning for us: F. Doré) and trace fossils (*Scolicia*, *Monomorphichnus Taphrhelmintopsis*) for which the world wide appearance remains to be studied (Vendian or Tommotian); but one must note that *Monomorphichnus*, a presumed trace of an arthropoda and possibly a trilobite, is at present known as soon as the Tommotian.

At a small distance away, these strata, of doubtful age, lie just above the granodiorite of Vire.

In conclusion, the specialists of the Massif Armoricain look at the problem as follows:

1. In Normandy, the Vendian–Tommotian boundary is to be sought in the levels located *between* the fossiliferous Cambrian above and the Cadomian discordance below; there is no reason for suggesting that the Vendian–Tommotian boundary should coincide with the Cadomian discordance.
2. In the present state of knowledge of the faunal assemblage, it is most likely that one should relate the levels above the granodiorite of Vire to the very early Cambrian. There are no definitive arguments to reject a late Vendian age. In any case the granodiorite was certainly intruded before the Atdabanian age.
3. There is absolutely no argument to relate the levels below the peneplain, that is the flysch facies regionally located in the Upper Brioverian, to Cambrian times.

The data from Morocco

We have no new information concerning the geochemical nature of the data published by Lancelot (1982), but in the light of recent discussions several points about the stratigraphical control of the Moroccan series need comment.

The archaeocyathids of Atdabanian age (not earliest but possibly early Middle) found in Tiout (Morocco) by Debrenne & Debrenne are located in the *upper part* of the Calcaire supérieur formation. The appearance of these archaeocyathids is due to a regional ecological factor according to Debrenne & Debrenne (1978) and F. Debrenne (pers. comm. 1983). The trilobites found below these first archaeocyathids are similar to the genus *Bigotina* and *Bigotinops* (Debrenne & Debrenne 1978) and are always in the *upper part* of the Calcaire supérieur which, accordingly, is of lower Atdabanian age.

A second correction concerns the quotation in Odin *et al.* (1983) of a remark by Cowie (1981) in which the Tommotian Petrotsvet Formation on the Aldan in Siberia was said to contain both Tommotian small shelly fossils and archaeocyathids *and* trilobites generally similar to those from Morocco (Cowie 1981 referring to Fyodorova *et al.* 1979, a paper in Russian not accessible to us). It is now known from the specialists (J. Cowie, F. Debrenne, pers. comm. 1983) that this opinion was based on a misidentification and that no trilobite remains were found in the Petrotsvet Formation.

Concerning the stromatolite remains, Odin *et al.* (1983) quoted the opinion of Schmitt (1978) who considers that the stromatolites from the Série Lie de Vin from Morocco (below the Calcaire supérieur formation quoted above) indicate a Precambrian age. However, this interpretation has been attacked by J. Sarfati, who published documented evidence refuting arguments to show that there is no ground to consider the stromatolites from this formation as Vendian. On the contrary, she greatly emphasized their *Cambrian character* and finally proposed to relate them and the Série Lie de Vin, in which they lie, to the *Tommotian* (Bertrand-Sarfati 1981). We do not know of any published refutation of this expert opinion. This conclusion, consequently, slightly modifies the figure in Odin *et al.* (1983), lowering the Vendian below the Série Lie de Vin. In any case, the syenite intruding the Calcaire inférieur formation (Dolomie inférieure of Choubert), below the Série Lie de Vin, can be referred to the Tommotian as the youngest possibility, being covered by the Série Lie de Vin of probable Tommotian age itself.

Concerning the lithological relations between the sedimentary sequence and the volcanics, the question has been raised whether or not the dated syenite and related volcanics are interlayered in the sequence. Here we can only refer and give confidence to the present knowledge of the field geologists: there is no alternative interpretation for the correlation of the trachyte-sediments in the Jbel Boho area. The field evidence seems clear to the specialists; we recommend the reader to consult the geological map: 1/200.000 Ouarzazate, Alougoum, Telouet Sud, published by the Service géologique du Maroc: Notes et Mémoires 138 (1970). The trachytic horizons are shown in concordance with the dolomies inférieures 30 to 45 km away from the volcano in several directions. Our Fig. 1 shows the sequence. We do not know of any specialist supporting the hypothesis of an hypabyssal intrusion and certainly not our colleague G. Choubert (pers. comm. 1983–1984).

However, it is clear that at present nobody knows where the Precambrian–Cambrian boundary, as suggested in China, lies in the Jbel Boho sequence. Most stratigraphers agree that it is below the Série Lie de Vin and somewhere in

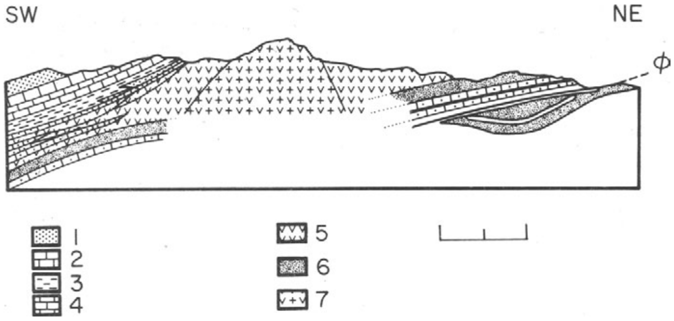

FIG. 1. Schematic section of the Jbel Boho volcano, Morocco (after G. Choubert, pers. comm., simplified). 1. Sandstones; 2. Calcaires sypérieurs; 3. Série Lie de Vin; 4. Dolomies et Calcaires inférieurs; 5. Trachyte; 6. Stratified andesites, tuffs and lapilli; 7. Syenite. The scale shows 2 km. The actual stratigraphic relations clearly show that the core of the volcano (syenite) is not intrusive: the trachytic and andesitic rocks, regarded as emissions of the volcano by Choubert, are interstratified in sediments below the Série Lie de Vin.

the Calcaires inférieurs of the underlying Ouarzazate series, which were deposited (according to radiometric data) during a time interval of more than 40 Ma.

Summarily, the dated Moroccan outcrop does not permit one to be absolutely sure that the obtained apparent age of 534 ± 10 characterizes a level below the *earliest* Tommotian, the present common view; however, this possibility cannot be ruled out. In effect, the apparent age characterizes a stratrigraphic level lithologically *well below* early Atdabanian faunas (i.e. the trilobites and archaeocyathids of the upper part of the Calcaire supérieur formation) as well as *below* stromatolites now accepted as Tommotian in age (Série Lie de Vin). According to present knowledge, the apparent age of 534 ± 10 Ma, geochemically undisturbed and in good rapport with many other results obtained from older rocks, is a *maximum age* for somewhere in the Tommotian (Fig. 2).

On the dating of the Ercall Granophyre and related rocks

Concerning the geochemical context of the Ercall Granophyres, its Rb-Sr whole-rock isochron apparent age was suspected to be a reset age. A series of measurements of U-Pb zircon ages was therefore recently undertaken in order to check the Rb-Sr age. The abstracted results were published recently (Compston *et al.* 1984). These authors measured an age of 531 ± 5 Ma which was interpreted in this abstract as a 'well substantiated isotopic age for the Ercall granophyre'. These U-Pb results do not disagree therefore with the Rb-Sr apparent age of 534 ± 12 Ma published earlier (Patchett *et al.* 1980), and no resetting seems to have occurred.

Compston *et al.* (1984) note, somewhat tautologously, that the relevance of this age for calibrating the time-scale depends on the actual relations of the dated rock with the fossiliferous sequence, which we now discuss.

The stratigraphic data published in Patchett *et al.* (1980) need revision. We may report first a relevant observation. In July 1983 one of us (G. S. Odin) was in the quarry and collected several curious structures in the quartzites. One of these pieces of rocks was submitted to Dr Rushton. His comment is as follows: 'the block contains a fragment of what

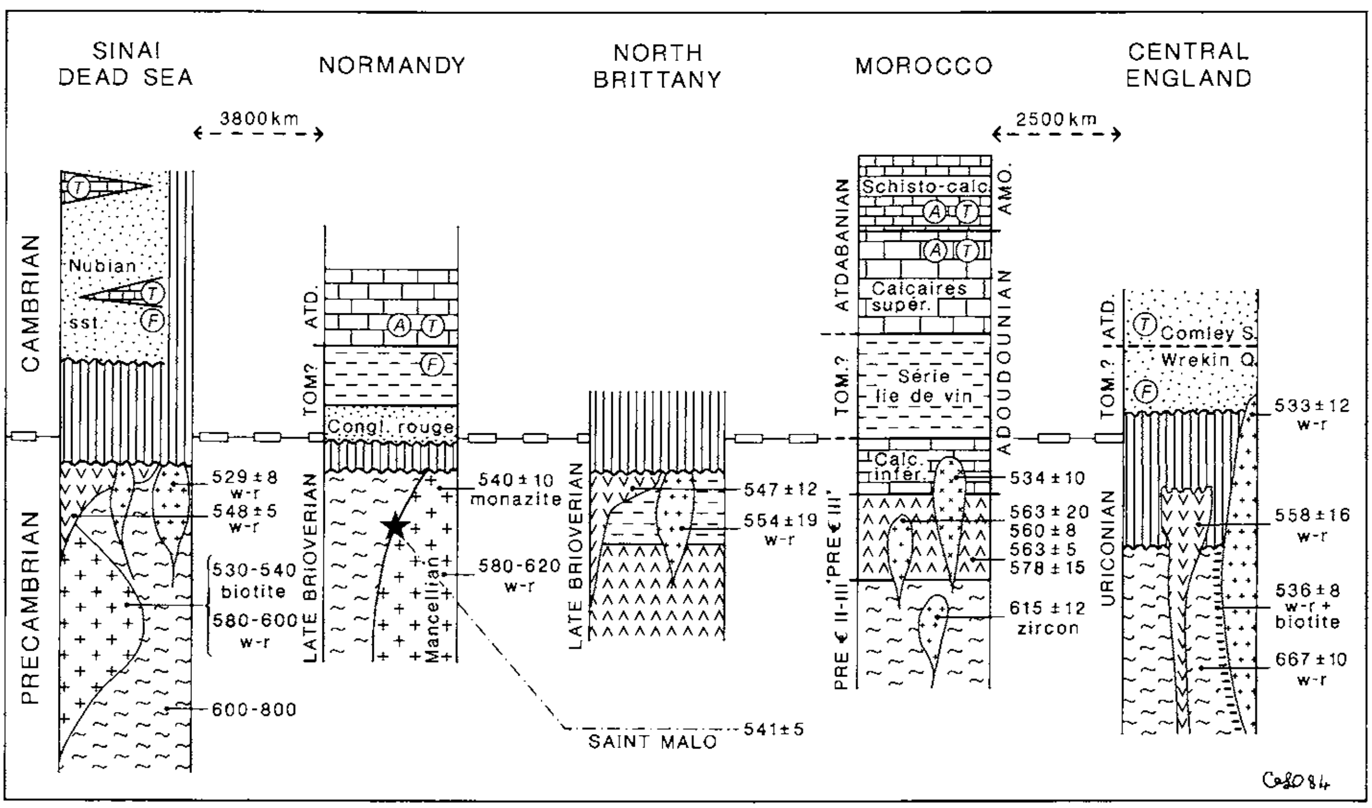

FIG. 2. Radiometric dates from NW Europe, Morocco and Sinai located according to their presently known stratigraphical location. (modified from Odin *et al.* 1983). The correspondence between the Precambrian–Cambrian boundary shown here and the newly proposed boundary in China remains to be reassessed; what is actually known at present is the location of the undoubted trilobitic Cambrian (i.e. the Atdabanian stage). A = Archaeocyathids; AMO = Amouslekian; ATD = Atdabanian; F = worms, trace fossil or skeletonized faunas; TOM ? = probable Tommotian; w.-r. = whole-rock Rb-Sr isochron age.

I think was a horny brachiopod, the nearby fragment that looks bryozoan-like is certainly inorganic, being a curious coincidental interface between quartz-grains and white powdery matrix' (pers. comm. to P. Toghill). The block will be submitted later for further study, but it is of interest to note that the Wrekin Quartzite may contain body fossils.

During the first annual meeting of Project 196 in London, Beckinsale reported new field observations according to which, the Ercall Granophyre was *intruded* into, and not overlain by, the Wrekin Quartzite (Beckinsale *et al.* 1984), as previously thought. A similar opinion is accepted by A. R. Palmer (pers. comm. April 13th, 1984) after a visit to the outcrop by several of his colleagues following the Precambrian–Cambrian boundary conference in Bristol, 1983. This obviously modifies the previously proposed interpretation of the radiometric data obtained from this granophyre from a probable maximum age to a minimum age of deposition of the Quartzite which is itself of imprecise stratigraphic age, but is most probably Cambrian. The overlying Atdabanian Comley Sandstone is not intruded by the Granophyre in the present outcrop. The date of 533 ± 12 Ma for the granophyre may therefore remain a maximum for the Atdabanian trilobitic Cambrian of the Comley Sandstone. One could obviously speculate that this sandstone has also been intruded, but this is not at present demonstrated in the field; on the other hand that the Comley Sandstone is not intruded is not proved either. We can only wait for new data on this point.

The radiometric age of 533 ± 12 Ma is therefore a minimum age for the base of the Cambrian, if the presence of skeletonized body fossils is accepted in the Wrekin Quartzite. We have no firm indication whether this age is a minimum or a maximum age for the Atdabanian fossiliferous formation overlying this Quartzite. The figure proposed in Odin *et al.* (1983) is to be modified accordingly (cf. Fig. 2).

On the dating of pre-Lower Cambrian Rocks from Sinai

We have no new data concerning the geochemical uncertainties related with the dating of the plutonic and volcanic rocks from Sinai. It will be recalled that diverse radiometric chronometers were used and that a coherent set of results was obtained measuring the age of a moment prior to the peneplanation. Most of the rocks dated from Sinai are lithostratigraphically located below a peneplanation phase which eroded all previous rocks.

The youngest age is given by the Mandar granite: apparent age 529 ± 8 Ma. One could in principle question the location of this granite below the peneplain, which is missing in this area. However, M. Bielski (in Odin *et al.* 1983) clearly stated that it has been destroyed by the erosion of 1–2 km of rocks overlying this granite. Therefore, the Mandar granite must be older than the peneplanation; it is lithostratigraphically below the peneplain and we see no field data whatever to suggest that the Mandar granite is Cambrian, as has been conjectured elsewhere in this volume.

The other Rb-Sr isochron apparent age available is 548 ± 5 Ma. (Amram Quartz Porphyry). These two data seem therefore to be maximum ages for the first fossil control located above these volcanics (shown locally either as post peneplain flows or as debris flows: Bielski 1982) as well as, by correlation, above the Mandar granite.

The biostratigraphic control is given by trilobites correlated with a mid Atdabanian from Morocco. Therefore, the radiometric ages available are actually maximum ages for part of the early Cambrian; we have no firm indications if they may be related to a level older than the Atdabanian. Until now, we have essentially quoted in this paper data which, in all rigour, seem to be coherent enough to estimate, if not the age of the base of the Tommotian, that of a time located before or not far from the base of the Atdabanian at 530–540 Ma. In spite of an older estimate proposed since about 1964 on dubious grounds, we have not seen any good arguments in the earlier literature capable of contradicting this new estimate. The Chinese numbers now found in numerous abstracts or briefly quoted in more important recent papers are clearly older, and are discussed below.

On the radiometric dates from China

A first general point about these dates concerns the availability of the data. It seems clear to every geochronologist that one cannot give full weight to numbers for which no properly documented publication is available. The example of the Russian results has properly been emphasized elsewhere (Gale 1982). Numerous numbers exist in the Russian literature, but, to our knowledge, no detailed analytical results and geochemical considerations are *published*, which at present precludes a correct use of these dates in discussing the Precambrian–Cambrian boundary. A necessary precondition for the assessment of radiometric dates is obviously the availability of details about the *analyses* of the samples: not only the radiogenic and radioactive isotope contents, but also the reproducibility and calibration of the apparatus; the geochemical constraints, for example, the mineralogy of the dated minerals and the history of the deposit; and indeed details about the stratigraphical relationships. For the Chinese dates we essentially have short abstracts, personal communications or preprints and this is clearly still insufficient, but more detailed results will apparently soon be available. The above remarks partly answer a very surprising comment recently made by a respected specialist in the calibration of the time-scale, i.e. W. B. Harland (1983, p. 397), who seems to doubt that there exists a consensus between geologists 'as to what constitutes determinations good in all aspects'. The answer is perfectly clear and simple in its principles. It has already been analysed in detail by one of us (Odin 1982): a determination good in all respects is a datum for which we have good information allowing us to discuss all possible uncertainties, analytical, geochemical, stratigraphic etc., related to a calculated age. We suggest that as long as sufficient information is not available, radiometric dates cannot serve as good key points for time-scale calibration. However, because the Chinese dates are already so often quoted we will make some preliminary remarks concerning the apparent opposition between them and the fully published results discussed above. The opposition is on the one hand an age of 530–540 Ma for a moment located below, or not far from, the base of Atdabanian. On the other hand the base of the Tommotian, if defined inside or at the base (Rozanov 1984, p. 24) of the Meishucun stage in China, is estimated at 615 ± 10 Ma (Ma *et al.* 1980; Zhao *et al.* 1980; Chen *et al.* 1981; Xing *et al.* 1982).

It has been argued elsewhere that the data obtained from

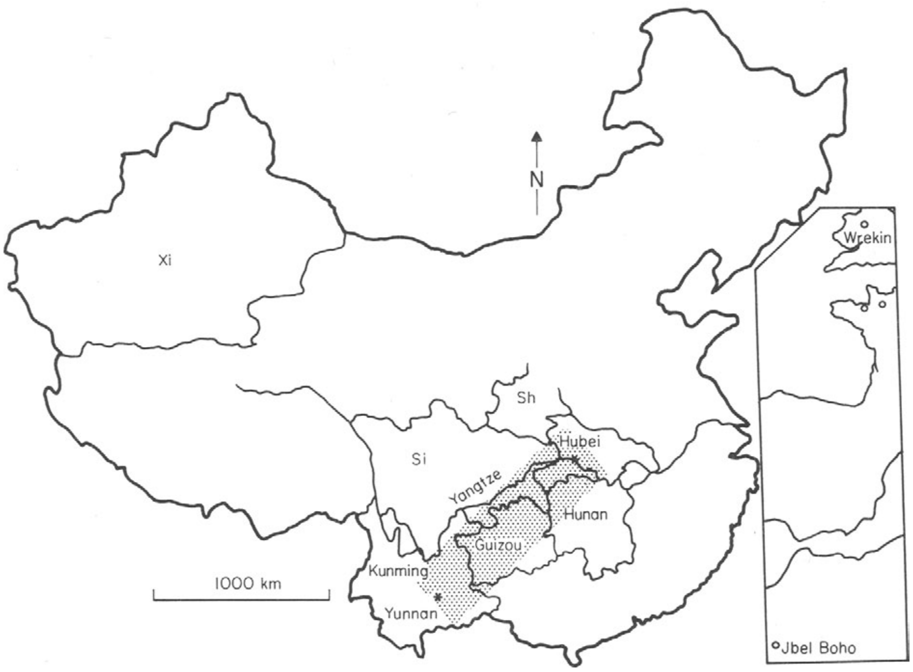

FIG. 3. A map of China showing the four Provinces where shales were dated (dotted area). The star in the Húbei Province shows the locality of Lientuo in the East Yangtze Gorges where many samples were dated. The star in the Yunnan Province shows the area where the Badaowan member was dated and where the newly proposed base of the Cambrian is located. For information the provinces of Xinjiang (Xi), Sichuan (Si) and Shaanxi (Sh) have been drawn; they also show outcrops of the earliest Cambrian. On the right hand side the *same scale* has been used to locate the outcrops dated in NW Europe and Morocco (circles).

Western Europe were from a comparatively limited sector of the world which could have been submitted to a common resetting event. In arguing against this we note first that there is more than 2000 km between the Wrekin Area and the Anti-Atlas of Morocco and nearly 4000 km between Normandy and the Sinai, while there is only about 1000–1200 km between the Yangtze Gorges outcrops (Hubei Province) and the dated outcrops 100 km south of Kunming, Yunnan Province (Fig. 3). Moreover, all the dated Chinese sediments are from a single shield of homogeneous geology (the Chinese Shield) which is not the case in NW Europe and NW Africa. We therefore suggest that the remark on possible geological factors in common applies better to the Chinese data than to the others.

If we pick some of the Chinese numbers available to estimate the age of the boundary we observe that, near the boundary, there are two analytical ages: 614 ± 18 and 602 ± 15 Ma. They are *both* located slightly above the boundary according to the table of data of Zhang *et al.* (1982). The precise numbers available lithostratigraphically well below the boundary, are of about 700 Ma. We conclude that, *if the numbers quoted represent a time of deposition*, then the base of the Tommotian is slightly older than 600 to 630 Ma.

If we accept both series of measurements, i.e. 530–540 for below the Atdabanian, 616–635 for the base of the Tommotian then the Tommotian stage has a duration of more than about 80 Ma. This is to say that the Tommotian stage, whose base is defined in China (and which has recently been added below the trilobitic Cambrian system), and which forms the first of the three Siberian stages composing the early Cambrian epoch = ex-Georgian (Doré 1982), would be 2.5–8 times larger in duration than the whole (as formerly defined) trilobitic Cambrian system if we accept a Cambrian–Ordovician boundary either at about 495 or 520 Ma. We have no fundamental arguments against that possibility.

However, results from the Chinese Atdabanian stage with trilobites also include Rb-Sr shale apparent ages of 573 ± 7 Ma (Yangtze Gorges), 569 ± 12 Ma (Guizou Province) and possibly 588 ± 13 Ma (Yunnan Province and similar U-Pb shale ages). The stratigraphical age corresponding to this later date from Yunnan (Tommotian or Atdabanian) is shown in Fig. 4 below. This means that, compared with the pre-Atdabanian rocks gathered in Odin *et al.* (1983) and reassessed here above, apparent ages different by more than 10% are found for rocks of nearly similar stratigraphic attribution. We actually accept all the quoted Chinese analytical data as *analytically* reliable, though there seems to be nothing published in enough detail to prove it. Therefore, there is necessarily a *geochemical* problem. There seems to be no serious possibility of systematic geochemical problems for all the European, Moroccan and Middle East plutonic and volcanic rocks dated by different geochronological methods and different laboratories. On the other hand, it appears that all rocks dated from China are either *clay* size fractions, whole-rocks from shales, or shaly limestones and this needs to be discussed.

The dating of shales, as whole-rocks or clay fractions, can hardly be considered as the usual means to obtain the age of the time of their deposition. For the illitic clays under discussion, accepted as the most favourable case, it is well known by the specialists that — by definition — the radiometrically measured apparent age *cannot* be the age of deposition. Without going into the details of the different applications of the Rb-Sr method to sedimentary whole-rock and clay fractions, we may consider that there are in fact two possible main situations. Illitic clays are *not* formed in the sea during deposition; therefore, their age of crystallization

	Yangtze Gorges (Hubei Province) mostly		Yunnan Province	
Cambrian	Shuijintuo (trilobites)	573±7 (○) 565–490 (*) 574±20 (+) 613±23 (++) 568±12 (+) 415–435 (○○)	Qiongzhusi stage	
			Meishucun stage	588±13 (+) 612±36 (++) 587±17 (+) 603±31 (++)
	Dengying	614±18 (○) 602±15 (+) 460±9 (○○)		
			Precambrian–Cambrian definition	
Sinian	Toushantuo	700±5 (○) 727±9 (○) 691±29 (+) 580±25 (○○) 580–420 (○) 460–340 (○○)	Dengying	
	Nantuo			
	Datangpo	728±27 (+) 608±15 (○) 500–360 (○○) (Hunan Prov.) (Hubei Prov.)		
	Gucheng			
	Lientuo	740±16		

FIG. 4. Radiometric results on Chinese shales: (+) from Ma *et al.* 1982; (○) from Zhang *et al.* 1982 on bulk clay size fractions more than 1.5 μm; (○○) idem but less than 1 μm in size; (++) from Xing *et al.* (1982); (*) the apparent ages of 565–490 Ma were obtained on samples at 35 km from those in the left column on similar fractions by the same laboratory; Zhang *et al.* (1982) added: (1) that there exists a variable proportion of detrital minerals in the bulk acid insoluble fraction; (2) that the age of the diagenetic component is ambiguous at 565 to 490 Ma. The dates of about 588 Ma Yunnan were from shales from the Badaowan member. Note that the age of 568 ± 12 Ma is quoted as a U-Pb whole-rock analytical age and is taken as an important criterion of the coherence and meaningfulness of the results.

represents either the *inherited* apparent age of the neighbouring eroded sediments or the age of the more or less *post-depositional* diagenetic influence which modified the inherited clays into illites by burial at depths of several thousands of metres. A mixture of both possibilities is not rare.

In both the simplest cases, as perfectly stated in Zhang *et al.* (1982), the age of *illite growth* may be approached using adequate size fractions which unfortunately cannot be assimilated to mineralogically pure clay fractions but only to enriched clay fractions.

It is not easy to distinguish geologically the two possibilities (unless we know the true numerical age of deposition and if we compare it with the apparent age measured). Fundamentally, an illite apparent age is therefore either a minimum age (diagenesis) or a maximum age (inheritance) of deposition, as clearly recognized by Zhang *et al.* (1982). It would, consequently, be an error to interpret the apparent ages from China as giving the age of deposition, though this has wrongly been assumed by most Chinese authors until now.

Even if we were to presume that the analyses are perfect and that the stratigraphy is certain, the unsolved geochemical problems destroy all possibility of using these data for time-scale calibration at the moment. Moreover we must remember that, at present, we know of no radiometric dates from shales actually accepted as key points for the calibration of the Phanerozoic time-scale.

During the meeting in Nikko (Japan), Dr Zhang clearly stated that the interpretation of the analytical results is not as simple as suggested by the more recent presentations.

We reiterate that only an objective report of *all* the analytical data as summarized especially by Zhang *et al.* (1982) may be considered the correct *information* to be discussed. Zhang *et al.* gave results which are *not* in the sequence quoted by others and therefore correctly report the result of their work. Zhang *et al.* also gave their own interpretation of their results and selected those dates which they considered useful for time-scale calibration; however, Zhang *et al.* clearly state the uncertainties of their conclusions. We consider at present that a selection of the results and their interpretation as time of deposition must be founded on a study of the analytical and geochemical details, when these are eventually published.

We, however, may note as an example that Zhang *et al.* interpret both the Datangpo age of 608 ± 15 and the Dengying age of 602 ± 15 (see Fig. 4) as the time of *diagenetic growth* of the illitic minerals, yet the stratigraphic deposition dates of these two formations are separated by about 100 Ma according to them. We therefore emphasize that a time duration of the order of 100 Ma after deposition is not considered impossible for that geochemical reaction (diagenesis). In this sense, Rb-Sr illite ages must be accepted as *minimum* ages of deposition as clearly stated by Zhang *et al.* (1982). Moreover, the difference between the ages considered reliable and those *interpreted* as registering the effect of a weak thermal or chemical event by Zhang *et al.* is, to our knowledge, that the first ones are measured on sediment size fraction of 1.5 μm or more and the second ones are measured on a clay size fraction below 1 μm, which certainly has some foundation but depends also on the process used for disaggregation of the rock.

A second point is that Zhang *et al.* are absolutely clear in quoting the presence of detrital minerals in the Shuijintuo formation: (1) 'near Changyang ... the bulk acid-insoluble fractions show that they contain a variable proportion of older detrital minerals, (2) "older" minerals are present in the Lientuo samples also', i.e. in the Yangtze Gorges (Hubei Province), (3) 'the presence of detrital minerals is evident from petrographic and X-ray diffraction studies'. Zhang *et al.* conclude on the one hand that the detrital minerals contribute little to the total Rb and radiogenic ^{87}Sr in comparison with that of illite but also that 'the isotopic dating does not exclude the possibility that the illite in one or more horizons might be detrital'. Apparent ages from shales may therefore be higher than the time of deposition. We present our apologies to Zhang, Compston and Page for quoting nearly all the information gathered in their very informative abstract, but we think it necessary to point out that their original data and interpretations are not fully reported in more recent presentations. We should emphasize that nevertheless Zhang and his colleagues clearly would at present favour an age of more than 600 Ma for the base of the Tommotian.

Conclusions

Not one of the recent results quoted in the literature indicate an age of 570–600 Ma for the Precambrian–Cambrian boundary and this intermediate proposal appears unfounded. To accept an age in this interval would imply that not one of the set of data here discussed is correct; a very pessimistic conclusion since there seem at present to be no better data for the calibration of this boundary.

A rigorous assessment of the data from Western Europe, Morocco and Sinai does not definitely prove that the base of the Tommotian is younger than about 530 Ma. It does, however, show some coherence at the worst for an early and pre-Atdabanian time located at 530–540 Ma and would entirely agree within the uncertainty margins with a basal Tommotian very near this numerical age.

Given our present knowledge of the geochemistry and stratigraphy of the areas studied the acceptance of an age of 610 Ma or more as suggested by the Chinese dates, if accepted as the time of deposition of the sediments, would imply a duration of more than 60 Ma:

1. between the pre-mid Atdabanian rocks of Sinai and the base of the Tommotian (within the Meschucun stage);
2. between the most probably pre-Tommotian but certainly pre-Atdabanian volcanic rocks of Morocco and the base of the Tommotian;
3. between the pre-Atdabanian rocks of the Massif Armoricain and the base of the Tommotian.

We see no conclusive arguments for or against these hypotheses but we believe that they remain today unfounded and would lead to considerable problems with the usual stratigraphic attributions in several areas, unless there is any strength in the suggestion that there may be considerable regional diachronism in the arrival of the earliest Cambrian faunas.

Finally we have to emphasize that the old estimate of 570 Ma or 590 Ma so frequently quoted in the literature was deduced in the past for the first trilobitic levels, i.e. the base of the Atdabanian. It seems obvious today, independently of the new stratigraphic definition proposed by the stratigraphers, that this estimate was too old by 40 Ma or so if we accept the results obtained from various sources and first emphasized in Odin (1982). This appears very important and it seems necessary to give full weight to this modern data, and to recognize that the earlier recommendations for the date of the Cambrian–Precambrian boundary (e.g. Cowie & Cribb 1978) were founded on very poor early data. The full and detailed publication of the very extensive work now in progress on Chinese formations is eagerly awaited, especially because it can help to answer such important questions as: the length of the pre-trilobitic Cambrian; the usefulness and accuracy of dates from shales within the Phanerozoic; and stratigraphic knowledge of the earth's history near the Precambrian–Cambrian boundary in several areas.

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The Ordovician, Silurian and Devonian periods

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SUMMARY: A reassessment of published age determinations, including some recent dates on Middle Devonian, Late Silurian, basal Caradoc and middle Arenig rocks, enables the construction of a more precise time-scale. The most accurate points on our scale are: basal Frasnian, 368 ± 8 Ma; early Ludlow, 421 ± 2 Ma; end of early Llandovery, 431 ± 6 Ma; early Caradoc, 450 ± 7; middle Arenig, 482 ± 5 Ma. Interpolations from these data indicate that the base of the Devonian is 412 ± 5 Ma, and the base of the Silurian is 435 ± 7 Ma. The end of the Devonian is less precise, we estimated it to be at 354 + 10, −5 Ma (depending on stratigraphic and sedimentological estimates for the duration of the Frasnian and Famennian stages). The base of the Tremadoc is perhaps 513 ± 10 Ma. Dating by K/Ar, $^{40}Ar/^{39}Ar$, U/Pb and Sm/Nd all show good agreement, but some Rb/Sr isochrons on acid volcanic rocks lie off our time-scale line. Most fission track dates on zircons from bentonites have very large errors; a few with the smallest errors are shown on our graph, but none of them have been precise enough to use in the construction of the time-scale.

Introduction

Advances in technology and analytical skill in isotope geochemistry have led to a situation where many modern isotopic ages on rocks with good stratigraphic control have 2σ errors less than 3%, so that many less accurate data have become redundant. There are, however, likely to be more improvements in the future, and the present scale can be expected to be superseded by a more precise scale in the next few years. We estimate that our correlation with the stratigraphic column is unlikely to be more than 7 Ma in error between the middle Arenig and the early Ludlow, but outside this range the errors are greater.

In recent years, ages have been determined by a variety of methods: K/Ar, $^{40}Ar/^{39}Ar$, Rb/Sr, U/Pb and Sm/Nd. It is interesting to note that ages by all these methods provide generally consistent results (all the dates used have been converted, where necessary, to the constants recommended by Steiger & Jäger 1977). A large number of less precise, but stratigraphically exact, fission track ages are now available; however, the 2σ errors on most of these are very large, and only a few with the smallest errors are shown in Fig. 1.

Since the review by McKerrow *et al.* (1980), new data have resolved some, if not all, of the discrepancies revealed by Gale *et al.* (1980), though there are still some divergent opinions and interpretations (Gale 1982; Odin 1982). Some of these depend on the weight attached to Rb/Sr isochrons from acid volcanic rocks, especially the Stockdale Rhyolite (Gale *et al.* 1979). In the main body of this paper we review 28 items on which we have constructed our time-scale (Fig. 1) or which we include for comparison. Many dates employed in McKerrow *et al.* (1980) are now redundant because of more precise data. Most of these have been omitted from the present paper.

The construction of Fig. 1

Most graphs are constructed with definite numerical scales along both the x and y axes; this is not the case with Fig. 1, where only the horizontal (x) axis is numerical. The vertical (y) axis is a stratigraphic time-scale, showing periods, series, stages and zones; the precise duration of each of these time divisions is unknown. In fact the whole object of this paper is to determine, as far as possible with the evidence available, what estimates can be given on the duration of these stratigraphic divisions. Thus in the course of preparing this figure, we have constructed a series of graphs, each with slightly differing vertical scales, until we obtained a scale which allowed a straight line to pass through almost all the rectangles representing the analytical errors (2σ) and the stratigraphical uncertainties in the data we use. Since this paper was presented orally in May 1982, new dates and new interpretations of some existing dates have resulted in some further adjustments to the relative durations of the stratigraphic units; and similar adjustments are the reasons why our present estimates differ from previous time-scales (Table 1).

In the construction of the stratigraphic units shown in Fig. 1, we have used the three Geological Society correlation charts (Cocks *et al.* 1971; Williams *et al.* 1972; House *et al.* 1977) together with some later revisions (Ziegler *et al.* 1974; Bassett 1976; House *et al.* 1979) and other published and unpublished revisions. These references have also been used in the updating of the stratigraphic ranges of some of the older items. In Fig. 1 we show only those stages and zones mentioned in the text.

Stratigraphers have long considered that different faunal zones appear to represent very varied time intervals. Some estimates have employed rates of deposition of pelagic sediments (Churkin *et al.* 1977), while others (Cocks 1971; Boucot 1975; Ziegler 1978) have employed subjective rates of evolution of zonal fossils. Several early drafts of our time-scale were constructed using different combinations of these estimated durations, but it soon became clear that modifications were necessary. For example, we have reduced the relative duration of the Upper Devonian (compared with the Lower and Middle Devonian) by 10% from that suggested by Ziegler (1978) in order to accommodate Items 26 and 28 on our best fit time-scale line.

In the Ordovician, there are no published estimates for the relative duration of series or stages. Although Churkin *et al.* (1977) have made estimates for the relative duration of some Ordovician graptolite zones, we are not satisfied that the sedimentation rates of these pelagic sediments are constant through time. It is in this period that our present scale deviates most from our previous scale (McKerrow *et al.* 1980). Items 9, 10, 11 and 12 together suggest that the

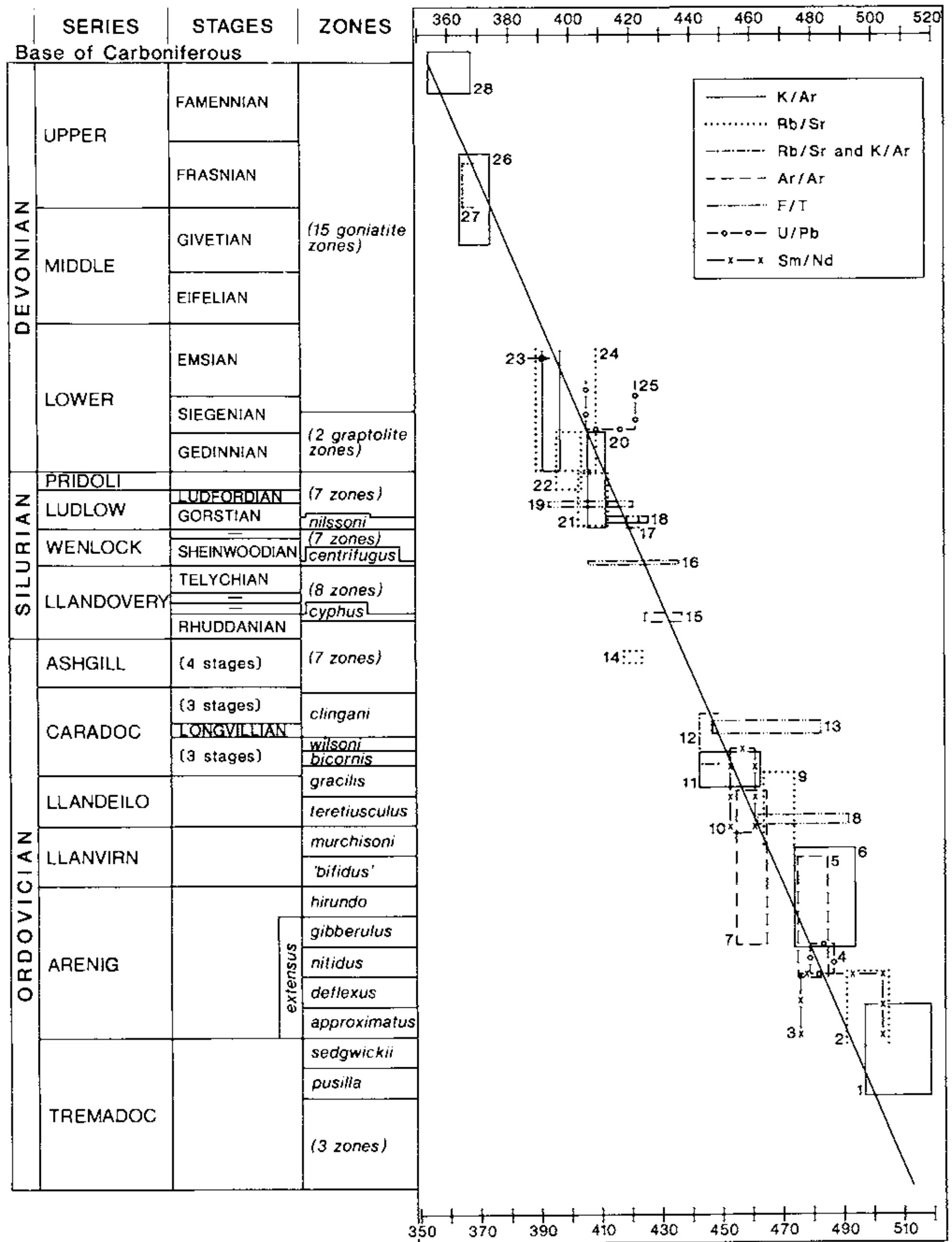

FIG. 1. Time-scale for the Ordovician, Silurian and Devonian. *Key to numbers*: 1 = Rhobell Volcanic Group; 2 = Twt Hill Granite; 3 = Southern Uplands Basalt; 4 = Byne Hill Trondhjeimite; 5 = Hare Bay Metamorphic Aureole; 6 = Colmonell Gabbro; 7 = Bay of Islands Metamorphic Aureole; 8 = Shales at Bach-y-graig; 9 = Benan Conglomerate clasts; 10 = Borrowdale Volcanic Group; 11 = Bail Hill Volcanic Group; 12 = Chasmops Limestone, Kinnekulle; 13 = Gelli-grin Calcareous Ashes Formation; 14 = Stockdale Rhyolite; 15 = Descon Formation; 16 = Buildwas Formation; 17 = Laidlaw Volcanics; 18 = Middle Elton Formation; 19 = Bringewood Formation; 20 = Gocup Granite; 21 = Wormit Bay Lava; 22 = Lorne Lavas; 23 = Skiddaw Granite; 24 = Shap Granite; 25 = Katahdin Batholith; 26 = Mount Morgan Tonalite; 27 = Cerberean Volcanics; 28 = Kelso Lavas.

bicornis and *wilsoni* Zones of the early Caradoc lie close to 450 Ma (± 7), while Item 16 gives an age of 431 ± 6 Ma for the top zone of the Rhuddanian stage (Lower Llandovery). We have decided that the Caradoc and Ashgill stages (based on trilobites and brachiopods) are more likely to have equal durations than the graptolite zones so, while we use the relative duration for the Llandovery stages suggested by Cocks (1971), in the Ashgill and Caradoc we assign each stage an equal duration. In a similar way, we have arbitrarily given equal durations to the '*bifidus*', *murchisoni* and *teretiusculus* Zones (Llanvirn and early Llandeilo). In the Arenig, we have the estimates of R. A. Fortey (pers. comm., 1983) for the relative durations of the graptolite zones, which have been drawn up on the basis of graptolite durations in Australia, Spitzbergen and Europe. This scale seems preferable to the assumption (Harland *et al.* 1982) that each Ordovician series has an equal duration, but in view of the fact that later graptolite zones have very different durations, this part of our scale may well be emended when more data become available. We have even less certainty about the pre-*nitidus* Zone part of our scale. The base of the Tremadoc will only be at 513 Ma if the trilobite zones of this series (Cowie *et al.* 1972) have similar durations to the Arenig graptolite zones. This is quite uncertain. The base of the Tremadoc could, in our opinion, lie anywhere between 495 and 523 Ma.

TABLE 1

	McKerrow *et al.* 1980	Jones *et al.* 1981	Odin 1982	Harland *et al.* 1982	This paper (for errors see text)
Carboniferous	360		360 + 5 −10	360	354
Famennian	369			367	365
Frasnian	377		375 ± 5	374	374
Givetian	383			380	383
Eifelian	390	383	385 + 8	387	391
Emsian	398	392		394	401
Siegenian	403	396		401	406
Gedinnian	411	401	400 + 10 −5	408	412
Pridoli	413	405		414	414
Ludlow	420	412		421	420
Wenlock	425	419		428	425
Llandovery	438	433	418 + 5 −10	438	435
Ashgill	445		425 ± 8	448	442
Caradoc	467		438 ± 5	458	454
Llandeilo	479		c455	468	461
Llanvirn	489		470 ± 10	478	470
Arenig	504		475 + 10 −5	488	492
Tremadoc	519		495 + 10 −5	505	513

The data

The numbers used for each item correspond to those shown in Fig. 1.

1. *Rhobell Volcanic Group*. Volcanic rocks south-east of the Harlech Dome in Wales cut late Tremadoc sediments which are certainly as young as the *S. pusilla* Zone and may be as young as the *A. sedgwickii* Zone (Kokelaar *et al.* 1982). They are overlain by the Garth Grit, which is probably of *deflexus* Zone age (R. Fortey, pers. comm. 1982) A mean conventional K/Ar age of 508 ± 11 Ma has been determined from five pargasitic amphibole separates from cumulates in a metabasalt (Kokelaar *et al.* 1982).

2. *Twt Hill Granite*. The Institute of Geological Sciences has obtained an Rb/Sr isochron of 498 ± 7 Ma from this intrusion, near Caernarvon, North Wales, which is overlain unconformably by the Fach Wen Formation of probable middle Arenig age (R. D. Beckinsale, pers. comm. 1981, 1982). There is no evidence to provide an older stratigraphical age limit, so the rectangle in Fig. 1 has not been closed at its base.

3. *Southern Uplands Basalt*. Basalts underlying Arenig cherts from the base of Sequence 2 (Leggett *et al.* 1979) near Raven Gill, south of Abington in the Southern Uplands of Scotland, have been analysed by Dr. M. F. Thirlwall (pers. comm. 1982), who has obtained a Sm/Nd age of 490 ± 14 Ma from localities 37 and 41 of Lambert *et al.* (1981, Fig. 1b and Table 3). The cherts are followed by black mudstones containing *Tetragraptus fruticosus* (Peach & Horne 1899, p. 288) and are therefore *deflexus* Zone or older in age.

4. *Byne Hill Trondhjeimite*. Bluck *et al.* (1980) obtained a zircon U/Pb age of 483 ± 4 Ma from this intrusion south of Girvan, Scotland. After some field investigations of our own, we now agree with them that it is likely that trondhjeimite clasts in the adjacent Arenig Ballantrae Volcanic Group are derived from the Byne Hill intrusion. We thus agree with tbeir conclusion that that intrusion is of middle Arenig (*nitidus* Zone) age. This zircon age was corroborated by less precise K/Ar ages of 487 ± 16 and 478 ± 8 Ma on amphiboles from a related gabbro and an amphibolite. It is considered tbat this amphibolite developed during emplacement of the Ballantrae Volcanic Group.

5. *Hare Bay Metamorphic Aureole*. $^{40}Ar/^{39}Ar$ dates on hornblendes from a metamorphic aureole of allochthonous ophiolites at Hare Bay, Newfoundland of 480 ± 5 Ma (Dallmeyer 1977) are post-early Arenig and pre-late Llanvirn. Items 5 to 9 of this paper are considered more fully in McKerrow *et al.* (1980).

6. *Colmonell Gabbro*. Harris *et al.* (1965) have obtained a K/Ar age on biotite from the Colmonell Gabbro, Scotland of 484 ± 10 Ma. This intrusion cuts the middle Arenig Ballantrae Volcanic Group, and appears to be earlier than overlying late Llanvirn sediments (Ingham 1978).

7. *Bay of Islands Metamorphic Aureole*. Hornblendes from the metamorphic aureoles of allochthonous ophiolites in western Newfoundland, the Bay of Islands ophiolite and the Little Port Complex, give a $^{40}Ar/^{39}$ Ar age of 460 ± 5 Ma (Dallmeyer & Williams 1975; Archibald & Farrar 1976). The stratigraphical time span of this date is limited by the probable middle Arenig age of the igneous rocks and the final emplacement of the thrust sheet in the *N. gracilis* Zone.

If our time-scale line is approximately right, the fact that it intersects the top of the rectangle for Item 7 (Fig. 1), suggests that the metamorphism may have occurred close to the age of the final emplacement of the thrust sheets.

8. *Shales at Bach-y-graig*. Zircons in an ash at the base of the *teretiusculus* Zone at Bach-y-graig, nearly Llandrindod Wells, Wales, have fission track ages of 476 ± 10 and 478 ± 12 Ma, with a 1σ error (Ross *et al.* 1982); combining these we calculate an age of 477 ± 15 Ma, with a 2σ error.

9. *Benan Conglomerate granite clasts*. An Rb/Sr mineral isochron age of 469 ± 5 Ma (Longman *et al.* 1979) for granite clasts in the Benan Conglomerate of Girvan, Scotland is now regarded by us as a maximum age for the late Llandeilo (see also McKerrow *et al.* 1980).

10. *Borrowdale Volcanic Group*. Four garnet host rock pairs from Llandeilo/early Caradoc volcanics in the nortb of England have a Sm/Nd age of 457 ± 4 Ma (Thirlwall & Fitton 1983). The samples came from low in the volcanic sequence, and are likely to be Llandeilo, rather than early Caradoc, in age.

11. *Bail Hill Volcanic Group*. The Bail Hill volcanic rocks near Sanquhar in the Southern Uplands overlie graptolitic shales of the *N. gracilis* Zone and yield a K/Ar age on biotite of 453 ± 10 Ma (Harris *et al.* 1965).

12. *Chasmops Limestone, Kinnekulle*. Biotite and sanidine

from bentonites in the Caradoc Chasmops Limestone of Mossen Quarry near Kinnekulle, Sweden, have been analysed on several occasions. They yielded a K/Ar date of 457 ± 8 Ma (Harland *et al.* 1964, Item 157); however, re-investigation of the original material (Williams *et al.* 1982) gives 445 ± 3 Ma from a biotite-feldspar Rb/Sr isochron, and 450 ± 6 Ma as the average of two K/Ar runs on the same biotite. As the K content of the biotite is only 5.9%, 'some alteration cannot be ruled out' (Williams *et al.* 1982). The best average age of 446 ± 3 Ma, is regarded by Williams *et al.* as a minimum estimate for these bentonites. In our view, the high initial $^{87}Sr/^{86}Sr$ ratio of 0.7123 ± 0.0001 may also indicate alteration, although Williams *et al.* (op. cit., p. 560) discuss this point and regard it as an original feature of the bentonite. Baadsgaard & Lerbekmo (1983) have obtained an Rb/Sr isochron from biotite and sanidine from the A_1 bentonite at Kinnekulle, which gives an age of 447 ± 1 Ma.

The age at this point in the early Caradoc is thus well established. The main uncertainty is now the precise graptolite correlation. Williams *et al.*, (1982, p. 563) state that the bentonites lie in the '*Diplograptus molestus* Zone' between the zones of *N. gracilis* and *D. clingani*, the stratigraphical situation is clearly presented by Thorslund in Waern *et al.* (1948): the bentonites in Mossen Quarry could be of any age from post-*gracilis* to mid-*clingani* Zones.

13. *Gelli-grin Calcareous Ashes Formation*. Longvillian beds near Bala, Wales have provided zircon for fission track ages (Ross *et al.* 1982) which average at 465 ± 18 (2σ) Ma.

14. *Stockdale Rhyolite*. This mid-Ashgill lava from northern England (Moseley 1978) has yielded an Rb/Sr isochron of 421 ± 3 Ma (Gale *et al.* 1979). Reasons why this date does not lie on our time-scale line have been discussed previously (McKerrow *et al.* 1980; Wyborn *et al.* 1982), but it is still used as a basis for other time-scales (Gale 1982). Compston *et al.* (1982) show that it is possible to re-interpret the data to give a minimum age of intrusion of 430 ± 7 Ma, and the time of hydrothermal alteration as 412 ± 7 Ma.

15. *Descon Formation*. A lower Llandovery (*M. cyphus* Zone) volcanic breccia in the Descon Formation of Esquibel Island, Alaska, yields (with a revised decay constant) a $^{40}Ar/^{39}Ar$ age on hornblende of 431 ± 6 Ma (Lanphere *et al.* 1977; Odin 1982; See also Kunk *et al.*, this volume).

16. *Buildwas Formation*. A bentonite in basal Wenlock (*centrifugus* Zone) siltstones of Shropshire, England, has a zircon fission track age of 422 ± 14 (2σ) Ma (Ross *et al.* 1982).

17. *Laidlaw Volcanics*. These volcanics near Canberra, Australia, are interbedded with fossiliferous sediments of early Ludlow (*nilssoni Zone*) age. K/Ar mineral analyses and Rb/Sr whole rock and mineral analyses together give an average age of 421 ± 2 Ma (Wyborn *et al.* 1982). This age is almost identical to the age determined for the Ashgill Stockdale Rhyolite (Item 14 above); Wyborn *et al.* discuss this inconsistency (see also discussion below).

18. *Middle Elton Formation*. K/Ar ages on biotites in a bentonite from the Ludlow Middle Elton Formation (upper *nilssoni* Zone) of Shropshire, England have an average of 419 ± 7 Ma (Ross *et al.* 1982). The zircon fission track age on the same bentonite is 407 ± 18 (2σ) Ma. See also Kunk *et al.*, this volume.

19. *Bringewood Formation*. Zircons from a bentonite in this Ludlow (uppermost Gorstian) formation have a weighted average fission track age of 407 ± 14 (2σ) Ma (Ross *et al.* 1982).

20. *Gocup Granite*. The post-Wenlock and pre-Siegenian Gocup Granite of New South Wales, Australia has a K/Ar date of 409 ± 3 Ma (Richards *et al.* 1977).

21. *Wormit Bay Lava*. A Rb/Sr isochron of 408 ± 5 Ma on biotite (M. F. Thirlwall, pers. comm. 1982) from a rhyolite in the Lower Old Red Sandstone at Wormit Bay on the north coast of Fife, Scotland occurs in a stratigraphic sequence of non-marine beds (Geikie 1902) containing spores of Gedinnian age (Richardson *et al.* 1984).

22. *Lorne Lavas*. Early Devonian volcanics from Argyll, western Scotland have yielded a Rb/Sr isochron of 400 ± 4 Ma (Pankhurst 1982, p. 580). The lavas overlie non-marine beds with fish and plant remains which are similar to those in the Gedinnian Arbuthnot Group of eastern Scotland (House *et al.* 1977). This date, like many other Rb/Sr isochrons on acid volcanics (see Item 14 above) does not intersect our time-scale line (Fig. 1), but it appears to be only slightly less than our present estimates for the Gedinnian stage.

23. *Skiddaw Granite*. The Skiddaw Granite of northern England has a K/Ar date of 399 ± 9 Ma (Shepherd *et al.* 1976). It was intruded into Lower Ordovician turbidites after they were folded in post-Pridoli times.

24. *Shap Granite*. The Shap Granite intruded Ordovician rocks of northern England after they were folded subsequent to the Pridoli; like the Skiddaw Granite, it post-dates the Caledonoid cleavage in the Lake District, but it is earlier than the widespread development of kink-bands (Moseley 1978). An Rb/Sr isochron gives an age of 394 ± 3 Ma (Wadge *et al.* 1978). We suggest that the age of the overlying Mell Fell Conglomerate could be Emsian or later, rather than Siegenian as suggested by House *et al.* (1977).

25. *Katahdin Batholith*. The Katahdin quartz-monzonite, north-central Maine, is an undeformed post-orogenic intrusion which cuts the Seboomook Formation (Rankin 1968) of Oriskany age (Boucot 1954). The Oriskany stage is now regarded as equivalent to the Siegenian (Gooday & Becker 1979). Zircon dates by Wones (pers. comm. 1982) give $^{207}Pb/^{206}Pb$, $^{207}Pb/^{235}U$ and $^{206}Pb/^{238}U$ ages of 414, 398, and 395 Ma respectively. Wones recommends 414 ± 8 (2σ) Ma as the best age.

The common Pb correction, affecting $^{206}Pb/^{204}Pb$ ratio of 4690 as measured, was taken as 400 Ma Pb on the Stacey and Kramers model. The correction on the measured ^{206}Pb is 0.4%, so it matters little whether the isotopic composition of the common Pb is correct or not within 10%. As the batholith is around 200 km south-east of the probable limit of old shield rocks (Rodgers 1968) and is emplaced in the calc-alkaline Piscataquis Volcanic Belt (Rankin 1968), the common Pb as used is a reasonable choice.

26. *Mount Morgan Tonalite*. This Queensland intrusion cuts Givetian sediments and is overlain by late Frasnian beds. It has a K/Ar age (on five samples of hornblende) of 369 ± 5 Ma (Williams 1982).

27. *Cerberean Volcanics*. These volcanics in Victoria, Australia are associated with a fish fauna which is regarded as early to middle Frasnian (Williams *et al.* 1982). Rb/Sr isochrons on biotite give a date of 367.1 ± 2.3 Ma; K/Ar on biotite gives 365.6 ± 4.8 Ma, and an Rb/Sr isochron (feldspar-whole rock) gives 368.6 ± 2.3 Ma (McDougall *et al.* 1966; Williams *et al.* 1982). Williams *et al.* recommend 367 ± 2 Ma as the average age; like Item 12, this should be regarded as a minimum age, because the results are all from minerals.

28. *Kelso Lavas.* These lavas occur on the Scottish-English Border above Upper Old Red Sandstone beds with a Famennian fish fauna (Odin 1982) and appear to lie very close to the base of the Carboniferous (W. H. C. Ramsbottom, pers. comm., 1982). They have a K/Ar age of 361 ± 7 Ma (de Souza, in Odin 1982).

Discussion

Since the review by McKerrow *et al.* (1980), 12 new ages have been obtained, which have small analytical errors and sufficiently good stratigraphical control to be of use in constructing the present time-scale. In addition, revised ages for five of our previous items have produced more precise results. These revisions have enabled us to discard many of the less reliable ages used in 1980.

The most significant new dates include Items 1, 2, and 3 (Rhobell Volcanic Group, Twt Hill and a Southern Uplands Basalt), which help to define the base of the Arenig at around 492 ± 7 Ma, and Item 4 (the Byne Hill Trondhjeimite), which, if the field deductions by Bluck *et al.* (1980) are correct, indicates that the Middle Arenig *nitidus* Zone lies at 483 ± 4 Ma. While this is a big improvement for the Arenig, the Tremadoc is still the least certain part of our time-scale, as we have no data points older than the end of that series. If the five Tremadoc trilobite zones are together of equal duration to the five Arenig graptolite zones, the base of the Tremadoc could lie around 513 ± 10 Ma, but if this is not true the error could be much larger. However, if the base of the Cambrian, variously estimated at 590 (Harland *et al.* 1982) to 530 Ma (Odin 1982; Odin *et al.* 1983), lies in the younger part of this range, a younger date might be preferred for the base of the Tremadoc (say 508 ± 5 Ma).

The revised minimum age of 446 ± 3 Ma for the Caradoc Chasmops Limestone from Kinnekulle (Item 12) entails one of the biggest changes from our 1980 review (previous dates on the same bed gave an age of 457 ± 8 Ma). When this minimum date is assessed together with Items 7, 8, 10 and 11, we conclude that the base of the Llandeilo must lie around 461 ± 7 Ma and the base of the Caradoc around 454 ± 7 Ma.

The Ar/Ar date of 431 ± 6 Ma (Item 15) for the Lower Llandovery Descon Formation suggests that the base of the Silurian lies around 435 ± 7 Ma. This is the only reliable date between the Caradoc and the Ludlow, but it does suggest that our estimates of the relative durations of the Upper Ordovician and Silurian stratigraphic units are not a great deal in error. But we stress that the bases of the Llanvirn, Ashgill, Llandovery and Wenlock series are all estimated by interpolation between the more precise data for the Arenig, Llandeilo/early Caradoc, early Llandovery and early Ludlow.

The early Ludlow Laidlaw Volcanics of Australia (Item 17) leave us in no doubt that the base of the Ludlow Series lies around 420 ± 2 Ma. This item (together with Items 15 to 25) shows that the Rb/Sr isochron of 421 ± 3 Ma for the Stockdale Rhyolite (Item 14) cannot reflect the age of extrusion of the lava. It is by rejection of this date that the present scale differs considerably from the Ordovician and Silurian portions of Odin's (1982) scale.

Although we are confident about the age of the basal Ludlow, we are much more uncertain about the base of the Devonian. The rectangles for Item 22 (Lorne Lavas) and 25 (Katahdin Batholith) do not overlap, and our line only just intersects the rectangle for Item 21 (Wormit Bay). Like other Rb/Sr isochrons on acid volcanics, Items 21 and 22 may be yielding dates that are too young, so we have constructed Fig. 1 such that our time-scale line just intersects the rectangle for Item 25. If this is correct, the base of the Devonian will lie at around 412 ± 3 Ma, but if more reliance is put on the U/Pb data from Katahdin Batholith of Maine (and less on Items 21 and 22), the basal Devonian could lie close to 415 Ma. The Skiddaw and Shap granites (Items 23 and 24) appear to have dates well after the start of the Devonian, effectively eliminating them as useful points in the construction of this scale.

Dates on the Middle/Upper Devonian Cerberean Volcanics of Victoria (Item 27) and the basal Carboniferous Kelso Lavas (Item 28) together suggest that the end of the Devonian lies between 350 and 360 Ma, but the errors could be greater if the estimated relative durations of the Devonian stages are much in error. The Devonian section of the time-scale remains poorly defined, except around the Givetian–Frasnian boundary (Items 26 and 27).

The fission track ages of Ross *et al.* (1982) and our time-scale

We agree with Gale & Beckinsale (1983) that most published fission track ages have errors which are too large to be useful in constructing a time-scale. Only the four fission track ages on zircons and/or apatites with the smaller errors in stratigraphically well-dated tuffs and bentonites (Ross *et al.* 1982) are shown in Fig. 1. In the construction of this figure, the weight attached to these has been negligible in relation to

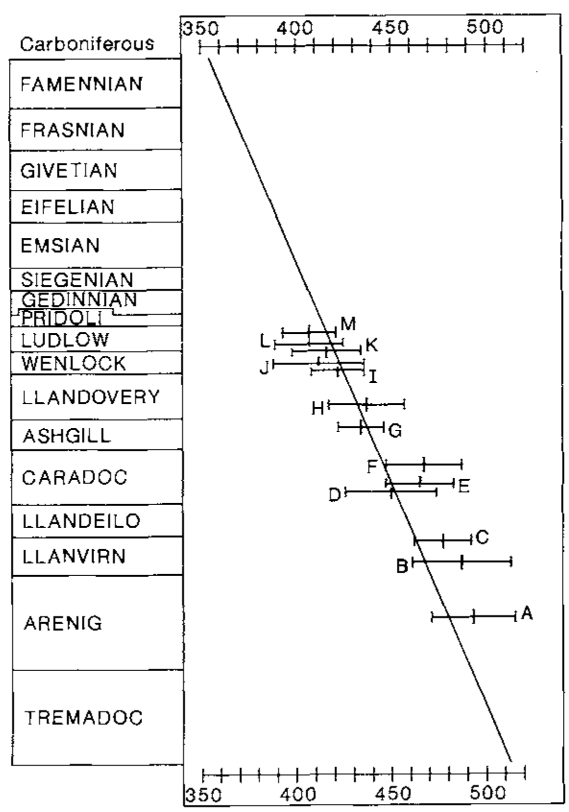

FIG. 2. The fission track ages of Ross *et al.* (1982) plotted on our time-scale. *Key to letters*: A = Llyfnant Flags; B = Stapeley Volcanics; C = Shales at Bach-y-graig (also Item 8 above); D = Fronddерw Ash; E = Gelli-grin Calcareous Ashes (Item 13 above); F = Acton Scott Formation; G = Upper Hartfell Shales; H = Birkhill Shales; I = Buildwas Formation (Item 16); J = Coalbrookdale Formation; K = Wenlock Limestone; L = Middle Elton Formation (Item 18); M = Bringewood Formation (Item 19 in this paper).

other data with smaller analytical errors. Accordingly, it is possible to consider these fission track ages independently of the other results (Fig. 2). It will be seen that almost all the fission track ages (when plotted with a 2σ error bar) intersect our recommended time-scale line. As a whole, when considered separately from our other data, the fission track ages would suggest a slightly longer duration for the Arenig to Ludlow time interval (and by implication, a shorter Devonian period). However, the ages of the Laidlaw Volcanics and the Katahdin Batholith (Items 7 and 25) are in conflict with a latest Silurian age of 400 Ma or so. We therefore consider that the Wenlock and Ludlow fission track ages, as a group, may be slightly too young (although not significantly different, individually, from our scale within a 2σ error). Similarly in the early Ordovician, the fission track ages have means which lie on the older side of our line, where, for the present, we give more weight to the Byne Hill Trondhjeimite age (Item 4).

Comparison with other time-scales

Our new scale deviates substantially from McKerrow *et al.* (1980) in the earlier parts of the Ordovician. New dates on the Rhobell Volcanic Group, the Twt Hill Granite, the Byne Hill Trondhjeimite, the Borrowdale Volcanics and the Chasmops Limestone (Items 1, 2, 4, 10 and 12) now show that the Arenig, Llanvirn and Llandeilo Series all start later than was previously thought (Table 1).

The present scale still differs markedly from those of Gale *et al.* (1980) and Odin (1982), mainly due to the weight placed on the 421 ± 3 Ma date of the Ashgill Stockdale Rhyolite (which we have included in this paper as Item 14, even though we have not used it to construct our time-scale line). We feel that the new data presented here reinforce our (1980) view that there must be some error present in that date. This matter has already been discussed (McKerrow *et al.* 1980; Ross *et al.* 1982; Wyborn *et al.* 1982), and we do not pursue it here.

A comparison of our present scale with Harland *et al.* (1982) shows identical dates both for the base of the Pridoli Series and for the Frasnian Stage; our biggest differences are at the base of the Tremadoc (8 Ma), the base of the Llanvirn (8 Ma), the base of the Ashgill (6 Ma) and the base of the Carboniferous (6 Ma). Although we have used some data not available to them, the principal differences between our scales results from different assumptions about the relative durations of series and stages. As stated earlier in this paper, the more we obtain accurate isotopic age dates, the less we shall have to make such stratigraphical assumptions on the duration of series and stages, but for the time being it is the only reasonable method of assessing the duration of stratigraphic intervals between dated items.

Jones *et al.* (1981) show consistently younger ages for the Silurian and early Devonian than other recent time-scales. Although based on 50 age determinations, they employ data which we have rejected, either due to large analytical errors or because of stratigraphical uncertainties. In order to apply a least squares regression line to the data, Jones *et al.* have ignored the errors on their data points, so that the same weight is given to vague and to accurate data. They have also used the Boucot (1975) and Ziegler (1978) durations of the Silurian series and Devonian stages without any modifications.

Summary

The last few years have seen a rapid increase in the number of accurate radiometric ages available from rocks which have adequate to highly precise stratigraphic assignments. A new scale has been constructed using 18 ages which were not available two years ago. The most accurate points on our scale are:

Givetian/Frasnian boundary:	368 ± 8 Ma
Early Ludlow:	421 ± 2 Ma
End of Early Llandovery:	431 ± 6 Ma
Early Caradoc:	450 ± 7 Ma
Middle Arenig:	482 ± 5 Ma

Stratigraphical interpolation between these points provides our summary scale (Table 1, column 5). The greatest uncertainties lie outside the range of these five points: in the latest Devonian and earliest Ordovician.

We hope this summary may encourage further efforts to fill in the gaps in the time-scale. We have still some way to go before we can discuss precise rates of evolution, rates of sedimentation, and rates of other geological processes with some confidence.

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Numerical calibration of the Palaeozoic time-scale; Ordovician, Silurian and Devonian periods

N. H. Gale

SUMMARY: Thirty-five reliable dates for rock units of well-defined biostratigraphy within the Ordovician, Silurian and Devonian periods are listed. Two alternative graphical presentations of the data show that all the data, including Rb-Sr whole rock ages of acid rocks, are internally consistent. Less precise fission track dates are consistent with the other data, but too imprecise sufficiently to constrain the time-scale. Estimates are given for the numerical calibration of the base of the Carboniferous, Frasnian, Eifelian, Gedinnian, Llandovery, Caradoc, Llandeilo and Arenig, by considering the sets of reliable ages for rock units near each boundary.

Much effort has recently been devoted to dating rock units of well-defined stratigraphy but reviews of such data (e.g. Cohee *et al.* 1978; McKerrow *et al.* 1980; Ross *et al.* 1982), whilst giving a more or less critical account of the stratigraphy, have often been rather less critical of the geochronology and have in particular sometimes misunderstood or paid scant attention to the errors associated with the numerical ages. In comparing and reviewing a set of stratigraphically controlled geochronological ages it must first be accepted that those ages can only be validly compared if errors at the same confidence level are attached to all of them. Further, in constructing a numerical time-scale it is essential to use errors at about the 95% confidence level or roughly ± two standard deviations of the mean. Moreover, when comparing geochronological data from different laboratories and using different radioactive geochronometers, it is not sufficient to use for error estimates merely the laboratory estimates for analytical precision. Some attempt must, for instance, be made to assess also systematic errors, for example in the measurement of the ratios of parent/daughter element concentrations and in the half-lives of the different geochronometers.

These ground rules are perhaps rather elementary, but even in very recent publications they have been ignored to the point where erroneous conclusions have been drawn. To take an example from the recent paper by Ross *et al.* (1982), it is claimed that the mean zircon fission track age of three Caradocian (Rocklandian) zircons from North America, at 453 ± 3.3 ($1\sigma_m$) Ma, is about 10 Ma younger than supposedly correlative units in the British Caradocian type sections, which averaged 462 ± 3.4 ($1\sigma_m$) Ma. [σ_m = standard error of the mean.] In fact these two dates are statistically indistinguishable within the two sigma errors of about ± 7 Ma common to each of them, quite apart from the fact that small sample statistics should be applied to test the difference of tbe means. Faced with such a test the means are indeed indistinguishable, and the alleged geological problem is thus wholly illusory. The recent review (McKerrow *et al.* 1980) of the Ordovician, Silurian and Devonian time-scales was similarly bedevilled with difficulties due to inadequate assessments of error, though the similar review in this volume (McKerrow *et al.* 1984) is admirably free of such problems.

Recent advances in geochronology have resulted in the production of a number of isotopic ages for rocks with good stratigraphic control with quoted precisions of less than 3%. Indeed there are now about a dozen radiometric ages for Ordovician-Silurian-Devonian times with quoted precisions of about 1%. Since these ages are based on a variety of methods (K-Ar, $^{40}Ar/^{39}Ar$, Rb-Sr, U-Pb and Sm-Nd), in their intercomparison in constructing a time-scale we must bear in mind that it is doubtful whether we know the respective half-lives to this degree of accuracy, and that there is some difficulty in avoiding interlaboratory analytical bias at this level.

Though a point has now been reached where it is possible to list 35 apparently reliable and precise radiometric dates for this part of the time-scale (see Table 2), even so they are insufficient to calibrate the bases even of all the series, far less the stages. In that situation a choice has to be made whether one rests content with the numerical calibration of just those stratigraphic boundaries for which there are a sufficient number of nearby, reliable and corroborative radiometric ages, or whether one resorts to some interpolative procedure to attempt to calibrate also those boundaries which are at present not directly documented by isotopic ages.

One temptation which ought firmly to be resisted is to use radiometric ages for one formation only, however good both the geochronology and the stratigraphy may appear, as an anchor point in the time-scale. Corroboration must be sought from equally good ages for other rock units of nearby stratigraphic location.

Interpolation, though previously much used (Boucot 1975; Gale *et al.* 1979a, 1980; McKerrow 1980) presents hazards of subjectivity, and calibration of series or stage bases so obtained must be regarded as provisional until relevant direct radiometric ages are available. Interpolation requires estimates of the relative durations of the stratigraphic series and stages. Such estimates when based on rates of sedimentation are now known to be unreliable and have rightly been rejected by McKerrow *et al.* (this volume); in their place they use estimates based on subjective rates of evolution of zonal fossils. Palaeontologists are indeed to be congratulated for their skill in deriving so well the general features of the relative durations of the stratigraphic series, but to use these relative durations for quantitative interpolation invites error and distracts attention from the primary goal of eventually establishing at least the series durations directly from radiometric ages.

That the subjective estimates of the relative durations of the series and stages are merely approximate guides was recognized by McKerrow *et al.* (this volume) who, in their graphical presentation of numerical radiometric ages against a stratigraphic time-scale (periods, series, stages and zones), successively adjusted the relative durations of the stratigraphic scale until a straight line passed through almost all the rectangles representing the analytical errors (2σ) and the stratigraphical uncertainties in the data used. Table 1 demonstrates briefly some of the major changes in relative

TABLE 1. Approximate relative durations of some stratigraphic intervals used in the graphical interpolation method

	McKerrow *et al.* 1980	Gale *et al.* 1980	Gale & Beckinsale 1983	McKerrow *et al.* 1982	McKerrow *et al.* this volume
Devonian/Silurian	1.75	1.64	1.69	1.75	2.45
Silurian/Ordovician	0.35	0.37	0.34	0.38	0.30
Devonian/Ordovician	0.61	0.61	0.57	0.67	0.74
Caradoc/Silurian	0.84	0.80	0.85	0.56	0.53
Caradoc/Ashgill	4.0	3.6	4.0	2.25	1.8
Caradoc/Arenig	1.5	1.54	1.35	0.97	0.58

stratigraphic durations introduced at various times by different authors who have used this graphical approach.

One conclusion reached by McKerrow *et al.* (this volume) is that some Rb-Sr isochron dates for acid volcanic rocks lie off their preferred interpolated time-scale line. It will be shown in this paper that the graphical interpolation method can be used in a way, not at variance with palaeontological guidelines, which results in a time-scale line which passes through these dates for acid volcanic rocks. Two such alternative graphical plots are presented which are constructed to pass through 34 out of the 35 available data points. Together with the solution proposed by McKerrow *et al.* they serve to demonstrate the subjectivity of this approach. The data is used also to suggest age intervals for the bases of the Tremadoc, Arenig, Llanvirn, Caradoc, Ashgill, Llandovery, Ludlow, Gedinninian, Famennian and Tournaisian stages.

The data

Of the 35 items listed briefly in Table 2, 23 are listed also, and the stratigraphy discussed, by McKerrow *et al.* (this volume). We have not used the four fission-track ages listed by McKerrow *et al.* (this volume) for reasons to be discussed later. We have rejected also the unpublished discordant U-Pb zircon dates for the Katahdin Batholith; neither can the data be assessed nor the possibility that inherited zircons are reflecting too old an age — note that in their Fig. 1 McKerrow *et al.* have great difficulty in forcing their time-scale line through this datum.

We list also 12 reliable items not used by McKerrow *et al.* (this volume); of these, 11 are fully discussed under the NDS numbers quoted in Table 2 in Odin (1982), whilst the stratigraphy of the Oliverian Syenite has been discussed by Foland & Loiselle (1981).

TABLE 2. Critical radiometric dates for the Palaeozoic time-scale

Item	*Locality*	*Geochronological Scheme*	*Ma* ± 2σ	*Stratigraphy*	*Reference*
1.	Kelso Lavas, Scottish Borders	K-Ar basalt	361 ± 7	pre early Tournaisian post late Devonian	NDS 165
2.	Post-cauldron intrusions, SE Australia	K-Ar biotite, hornblende, muscovite	362 ± 3*	Famennian	NDS 235
3.	Cerberean Volcanics, SE Australia	K-Ar biotite	367 ± 4*	Frasnian	NDS 234
4.	Mt. Buller granodiorites and Woods) Pt. dyke swarms, SE Australia)	K-Ar, biotite, hornblende	381 ± 4*	Eifelian–Givetian	NDS 233
5.	Shap Granite, English Lake District	Rb-Sr, WR, biotite	394 ± 4*	Gedinnian–Givetian	NDS 241
6.	Skiddaw Granite, English Lake District	K-Ar, biotite	398 ± 8	post Pridoli	NDS 192
7.	Pembroke Volcanics, Appalachians, USA	Rb-Sr, WR	393 ± 7	Early Gedinnian–Lower Emsian	NDS 237
8.	Lorne Lavas	Rb-Sr	399 ± 5	Gedinnian	A.
9.	Hedgehog Volcanics, Appalachians, USA	Rb-Sr, WR	400 ± 10^{+}	Late Gedinnian	NDS 223
10.	Gocup Granite, Australia	K-Ar, muscovite	409 ± 5$^{\neg}$	Post Wenlock pre Siegen	NDS 210
11.	Laidlaw Volcanics, Australia	(K-Ar, sanidine (K-Ar, biotite (Rb-Sr, WR (Rb-Sr, biotite	409 ± 5) 420 ± 5) 425 + 17) 421 ± 5)	Ludlow, *nilssoni*) Zone)	B.
12.	Middle Elton Formation, Shropshire	K-Ar, biotite	419 ± 10	Gorstian, Ludlow	C.
13.	Quoddy Volcanics, Appalachians, USA	Rb-Sr, WR	408 ± 8	Late Llandovery	NDS 238
14.	Descon Formation, SE Alaska	^{40}Ar-^{39}Ar, hornblende	431 ± 7	Llandovery, *M. cyphus* Zone	NDS 128
15.	Stockdale Rhyolite, English Lake District	Rb-Sr, WR	421 ± 5*	Ashgillian	NDS 243

16.	Eskdale Granite, English Lake District	Rb-Sr, WR	429 ± 4	Late Caradoc	NDS 189
17.	Oliverian Syenite	Rb-Sr, WR	441 ± 5	Caradoc, post lower *clingani*?	D.
18.	Carters Limestone, Alabama, USA	K-Ar, biotite, sanidine	455 ± 10	Early to Middle Caradoc	NDS 129
19.	Tyrone Limestone, Kentucky, USA	K-Ar, biotite	443 ± 10	Middle Caradoc	NDS 161
20.	Tyrone Limestone, Kentucky, USA	^{39}Ar-^{40}Ar	433 ± 5	Middle Caradoc	E.
21.	Kinnekulle, Sweden	(Rb-Sr, biotite (K-Ar, biotite	445 ± 4 450 ± 6	post *gracilis*,) pre *clingani*)	F.
22.	Borrowdale Volcanics, English Lake District	Sm-Nd	457 ± 5	Llandeilo to early Caradoc	G.
(23.)	Bail Hill Volcanics, Scotland	K-Ar, biotite	455 ± 15$^{\square}$	*gracilis* Zone	H.
24.	Benan Conglomerate, Scotland	Rb-Sr	470 ± 5	pre *gracilis* Zone	NDS 135
(25.)	Colmonell Gabbro, Scotland	K-Ar	484 ± 10	(post mid-Arenig) (pre late Llanvirn)	H.
26.	Gt. Cockup Picrite, English Lake District	K-Ar, biotite, hornblende	468 ± 10	Lower Llanvirn	NDS 191
27.	Hare Bay Ophiolite, Newfoundland	^{40}Ar-^{39}Ar	480 ± 5	Mid Arenig to Mid Llanvirn	I.
28.	Byne Hill, Scotland	U-Pb, zircons	484 ± 4	Mid Arenig, *nitidus* Zone	NDS 134
29.	Bay of Islands gabbro, Newfoundland	^{40}Ar-^{39}Ar	460 ± 5	Mid Arenig *or* *gracilis* Zone	J.
(30.)	Twt Hill Granite	Rb-Sr	498 ± 7	pre mid-Arenig	K.
(31.)	Rhobell Fawr, Wales	K-Ar	508 ± 17$^{\square}$	Late Tremadoc	L.
32.	Krivoklat-Rokycany, Bohemia	Rb-Sr, WR	491 ± 14	Late Cambrian to pre early Tremadoc	NDS 130
33.	Southern Uplands Basalt	Sm/Nd	490 ± 14	*deflexus* Zone or older	M.
34.	Mount Morgan Tonalite	K-Ar, hornblende	369 ± 5	Givetian-Frasnian	F.
35.	Wormit Bay Lava	Rb-Sr, biotite	408 ± 5	Ludlow-Pridoli	M.

o Stratigraphy from W. S. McKerrow

NDS: Abstracts *In*: Odin, G. S. 1982. *Numerical Dating in Stratigraphy*, Part Two, John Wiley, Chichester.
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E. Quoted in Gale, N. H., Beckinsale, R. D. & Wadge, A. J. 1979. *J. geol. Soc. Lond.*, **136**, 235–42.
F. Williams, I. S., Tetley, W., Compston, W. & McDougall, I. 1982. *J. geol. Soc. Lond.*, **139**, 557–68.
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M. Thirlwell, M. F. 1982. (pers. comm.)

In Table 2 four items, whose numbers are enclosed in parenthesis, are less reliable than the others. The data for the Twt Hill Granite is not yet published and cannot be assessed, moreover the stratigraphy is not yet certain. For Rhobell Fawr, Item 31 in Table 2, the Cambridge data consists of five K-Ar dates on pargasites of 506, 505, 491, 530 and 508 Ma; the mean is 508 Ma, $\sigma = 14.02$, $\sigma_m = 6.27$, Students t at the 95% level is 2.78, so the 95% confidence level is ± 17 Ma (the authors quote 508 ± 11 Ma). Confidence in the present data for Rhobell Fawr is weakened by the age of 475 ± 12 Ma quoted as NDS 122 in Odin (1982). The old data for the Colmonell Gabbro (Item 25) and the Bail Hill Volcanics (Item 23) are not very reliable. For the Bail Hill volcanics four K-Ar dates on biotites were quoted of 463.4, 463.4, 445.3 and 448.4 Ma (corrected to ICC); the mean is 455 Ma, $\sigma = 9.64$, $\sigma_m = 4.82$, Students t is 3.18 at the 95% confidence limit, so the error is ± 15 Ma. An objection may be made in that one should not take the mean of a group of variable K-Ar ages on altered biotites (though McKerrow *et al.* (this volume) quote 453 ± 10 Ma for Bail Hill), since the younger ages probably represent argon loss, but the Bail Hill biotites contain potassium at a level where Obradovich & Cobban

(1976) have shown that biotites can occasionally record a K-Ar age higher than the true age. Probably the existing data for Bail Hill and the Colmonell Gabbro should be given low weight for time-scale purposes.

The errors quoted for some items in Table 2 are slightly higher than the laboratory precision quoted by the author, in an attempt to allow for possible interlaboratory systematic bias. In particular, note that the age for the Stockdale Rhyolite was originally quoted for comparative purposes (Gale *et al.* 1979a) as 421 ± 5 Ma, but has been persistently misquoted as 421 ± 3 Ma.

The new data (Item 11 in Table 2) for the acid Laidlow Volcanics leave McKerrow *et al.* (this volume) 'in no doubt that the base of the Ludlow Series lies around 420 ± 2 Ma'. The stratigraphy of the volcanic formations in the Canberra area has been uncertain for a long time. It seems that the stratigraphy of the dated Canberra volcanics has now been established by lithological correlation with volcanic formations in the Yass area where the stratigraphic relations are not in doubt (Wyborn *et al.* 1982); it may be that the dated volcanics are not quite so closely constrained as is at present claimed. Leaving that aside, the rocks include inherited zircon (Gulson & Rankin 1977) and Wyborn *et al.* may possibly too lightly have dismissed the idea that the biotite may also in part be inherited, though the concordance of the Rb-Sr and K-Ar ages for the biotite would normally cause one to accept them as dating the igneous event. The mean sanidine K-Ar age is lower than the mean biotite age, yet sanidine is generally more resistant to secondary alteration than biotite, contains less initial Ar and is generally accepted as a reliable mineral for K-Ar dating (Baadsgaard & Lerbekmo 1982; Hess *et al.* 1983). Again, the Rb-Sr whole rock system in the Laidlaw Volcanics is clearly disturbed and the dated biotites are chloritized, with lowered potassium content. The study of Obradovich & Cobban (1976) showed that in these circumstances biotite ages can be both too low or too high; though it is true that this study of late Cretaceous biotites revealed difficulty only for biotites containing five or less per cent potassium (the dated biotites for the Laidlaw Volcanics contained 5.9 and 6.5% potassium), it might be that the effect could occur at higher potassium concentrations for older biotites; could it occur also for the Rb-Sr system? Faced with such possibilities, however remote, it seems unwise to use alone the data for the Laidlaw Volcanics as an anchor point for the time-scale; we must await corroborating evidence.

The new precise data for the Kinnekulle Bentonite (Williams *et al.* 1982) is welcome. The Al bentonite which was dated comes from a quarry at Mossen (Byström-Asklund *et al.* 1961), but the stratigraphy was established directly only for the core taken at Kullatorp, 1.2 km. from Mossen. The Mossen section was not established by Thorslund (1948), but given to him by Rudberg; the correlation between the bentonite beds at Kullatorp and Mossen is probably good, but was said by Thorslund (1948, p. 352 and Fig. 3) to be no better than tentative. If the correlation is correct, the Al bentonite lies anywhere between the post-*gracilis* to mid-*clingani* Zones.

Graphical presentations of the data

We reiterate that, in a preliminary discussion of the data, we use the graphical method chiefly as a convenient method for visual presentation and to emphasize that more than one graphical plot can be made of the existing data. In common with McKerrow *et al.* (1984) we have been guided in our choice of relative stratigraphic subdivisions by the publications of Cocks *et al.* (1971), Williams *et al.* (1972), Ziegler *et al.* (1974, 1978), Boucot (1975), House *et al.* (1977), and House *et al.* (1979). Like them, we too have modified the stratigraphic axis until a scale was obtained which allowed a straight line to pass through the rectangles representing the uncertainties in the data used, but unlike them we have sought to use *all* the reliable data now available and have not accepted a final solution which does not allow the line to pass through all the data, including that for acid rocks.

McKerrow *et al.* (this volume) stress that their time-scale line does not pass through the Stockdale Rhyolite datum and state incorrectly that their preferred time-scale 'differs markedly from those of Gale *et al.* (1980) and Odin (1982), mainly due to the weight placed on the 421 ± 3 (sic) Ma date of the Ashgill Stockdale Rhyolite'. The time-scale suggested in Odin (1982) (properly Gale (1982)) was based on a discussion of three dates for the Silurian and 19 for the Ordovician; moreover Gale & Beckinsale (1983) stressed that undue weight should not be given to the Stockdale Rhyolite age and showed that all the data then available (including that for five acid volcanic formations) could be accommodated on a straight line by suitable choice of the stratigraphic axis — the line was not artificially constrained to pass through the Stockdale Rhyolite datum.

Examination of Fig. 1 of McKerrow *et al.* (this volume) shows that their preferred line is chiefly constrained by having to make it *just* pass through the corner of the rectangles for the Katahdin Batholith, the Wormit Bay Lava and the Laidlaw Volcanics biotite data — even so, the Laidlaw Volcanics biotite datum can only just be accommodated. In contrast Fig. 1 of this paper shows how, with a different vertical axis, all the 35 data points (using the sanidine data for the Laidlaw Volcanics) can be made to lie on a straight line.

It should perhaps especially be noticed that Rb-Sr isochron ages for acid volcanics (Items 7, 8, 9, 13, 15 and 32) fit this line quite well. These acid volcanics are all either rhyolites or ignimbrites and the fact that they fit the line, defined essentially by dates on other types of rocks, should at last dispel the widely held prejudice against dates for such acid rocks. All geochronological systems may, in some circumstances, be reset, but it has been argued before (Gale *et al.* 1979b, 1980, 1983) that certain types of acid rocks are not always reset. Direct proof of this comes from examples of Rb-Sr isochron ages for acid volcanics which are concordant with ages derived from other decay schemes. For instance, Cleverley (1977) has dated Karoo dolerites and rhyolites from the Lebombo of Swaziland. An Rb-Sr whole rock isochron for the rhyolites yields an age of 189 ± 7 Ma with an MSWD of 1.1; the dolerites not only fall on the same isochron but also yield an independent mean K-Ar age of 188 ± 5 Ma. Moreover, recent work on the Cerberean Volcanics in Australia (Williams *et al.* 1982) gives a mean Rb-Sr biotite-whole rock age of 367 ± 3 Ma and a mean K-Ar biotite age of 366 ± 5 Ma for the biotite rhyodacite, whilst for the contemporaneous basal rhyolite the Rb-Sr isochron age for whole rocks and feldspars yields an age of 369 ± 3 Ma. Further, Auvray (NDS 249, in Odin 1982) quotes Rb-Sr whole rock dates for granites and rhyolite ignimbrites in the Trégor area, Northern Massif Armoricain; the granites yield 554 ± 19 Ma whilst the rhyolitic ignimbrites gave 547 ± 12 Ma.

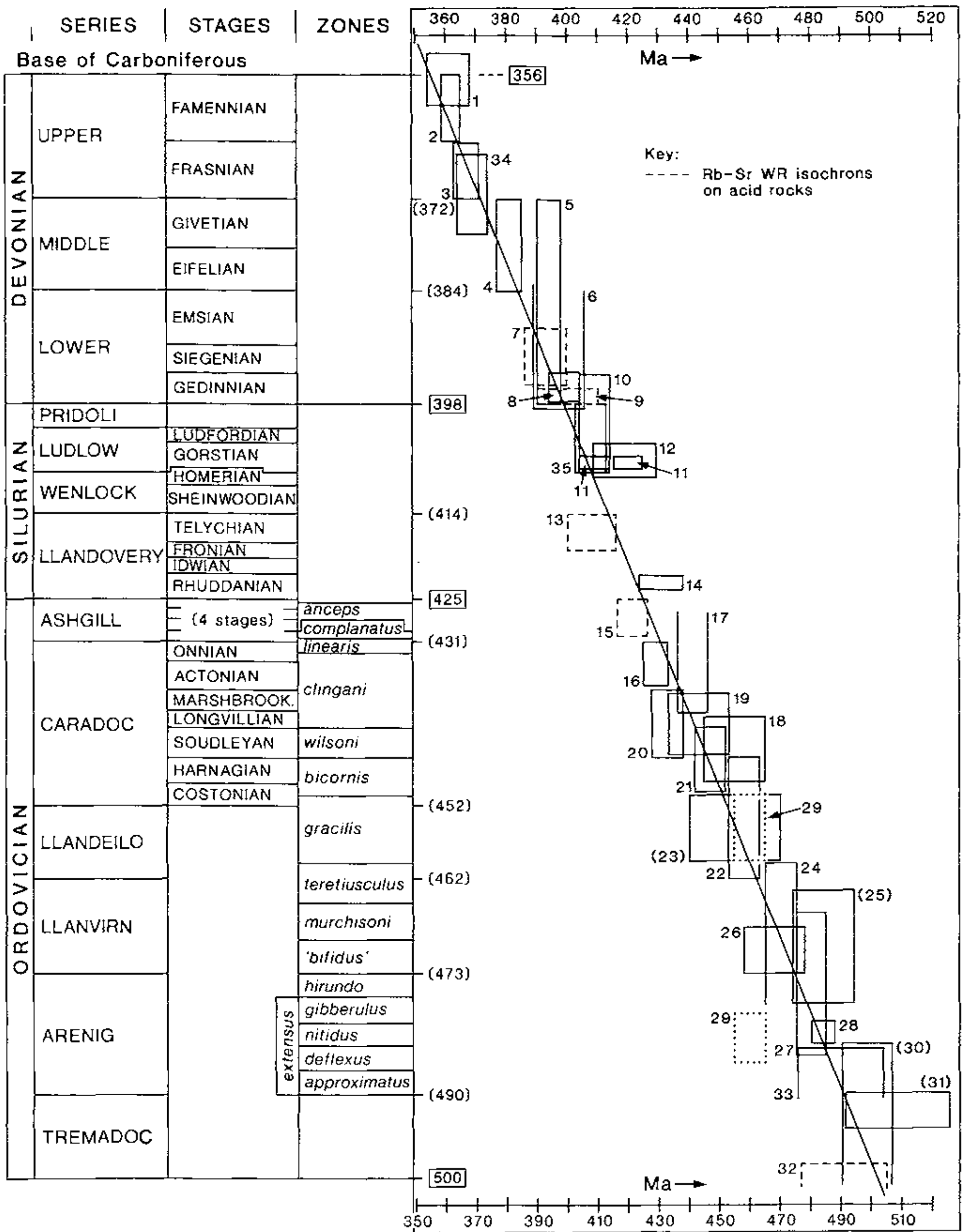

FIG. 1. Plot of critical radiometric ages against a stratigraphic axis where the relative lengths of the stages and zones have been chosen to be generally consistent with the rather subjective estimates made by palaeontologists and stratigraphers. The numbers of the items correspond with the list given in Table 1. Two sigma errors for the ages are used throughout. Rb-Sr isochron ages on acid volcanics are distinguished by a broken line. The stratigraphic evidence for the Bay of Islands gabbro, Item 29, places it either in the Mid Arenig *or* in the *gracilis* zone; the correlation with other data shows that it must be placed unequivocally in the *gracilis* zone. The dates shown for the bases of Series and stages are merely those given by the intersections with the straight line passing through the data; final recommendations for these numbers are given in the text.

In Fig. 1 a line cannot be forced through the biotite data for the Laidlaw Volcanics without causing it also not to pass through some other reliable data. In Fig. 2 the duration of the Silurian has been contracted in relation to the Ordovician and Devonian Periods; in this way the Laidlaw Volcanics biotite data can just be accommodated with all the other data with a relatively small effect on the resulting time-scale except in the Silurian period. Other variations of this approach can be envisaged, but are of little value in the absence of more closely defined data.

Fission track data

The fission track ages of Ross *et al.* (1982) have not been used here because the errors are too large to he useful in constructing a time-scale. The authors believe that the errors to be attached to these fission track ages have been properly assessed by Gale & Beckinsale (1983). However, in that paper the authors relied too much on criticisms made by Storzer & Wagner (1982) of certain technical features of the fission track measurements reported by Ross *et al.* (1982). The authors are now convinced that the technical quality of these measurements was of a very high standard, that possible track annealing was properly considered and that the zircons, dated by the external detector method, were oriented in such a way that the geometry factor was known with certainty.

McKerrow *et al.* (this volume) have shown in their Fig. 2 that these fission track ages, when plotted with a 2σ error bar, all intersect their recommended time-scale line, but place little constraint on it. The same is true if the fission track ages are plotted on Figs 1 and 2 of this paper.

Constraints imposed on series and stage bases by the data

In this section we investigate how the set of 35 reliable data may be used to estimate probable intervals within which the bases of some of the Series and Stages fall.

For the base of the Carboniferous we have the 361 ± 7 Ma age for the Kelso lavas (Item 1), placed by De Souza (1982) as post late Devonian, pre early Tournaisian, and probably very

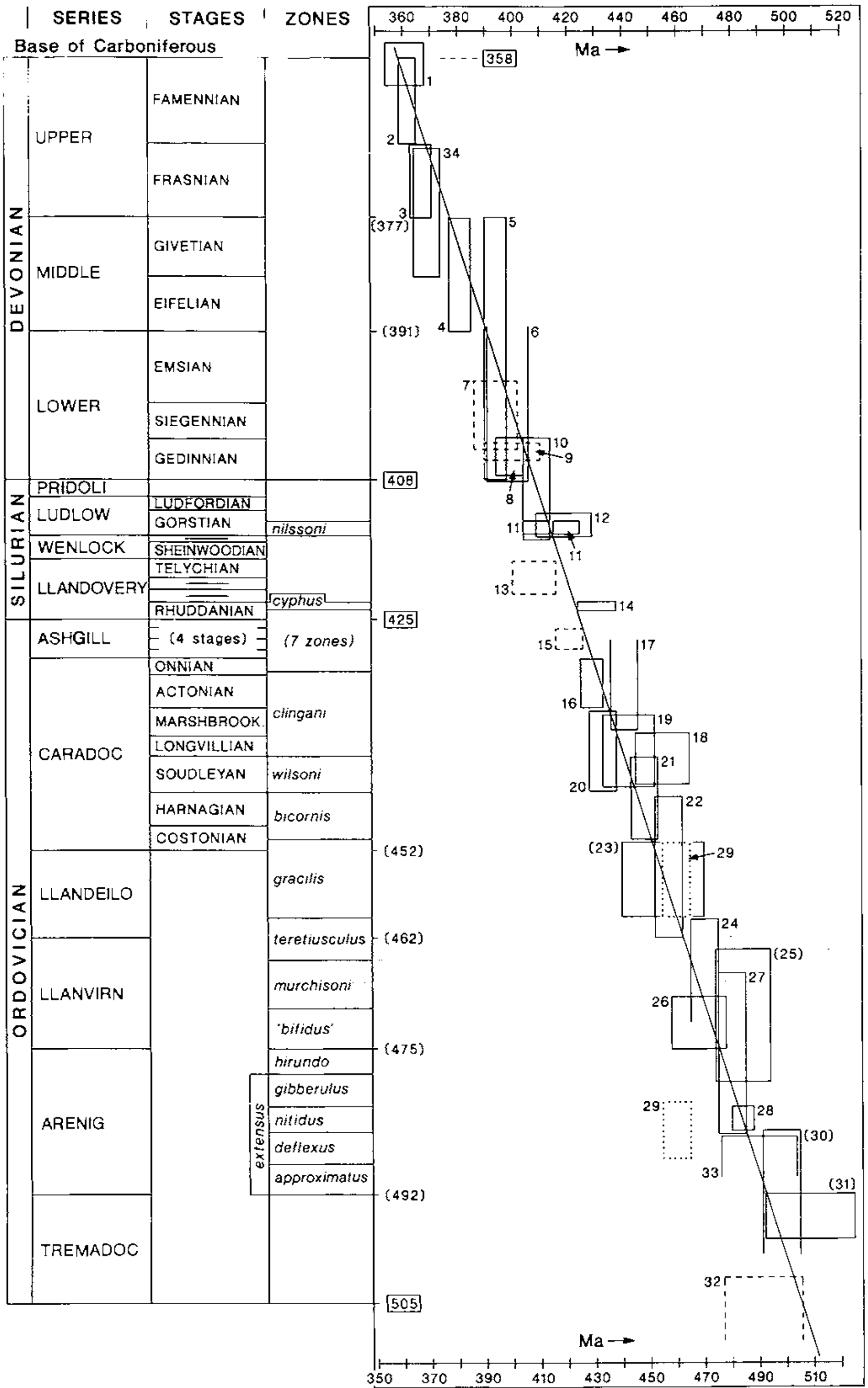

FIG. 2. Alternative plot with a different stratigraphic axis, chosen to reconcile the biotite age for the Laidlaw Volcanics.

close to the base of the Carboniferous. Further information comes from the minimum age of 362 ± 3 Ma for the Famennian post-cauldron intrusions (Item 2) and the 367 ± 4 Ma minimum age for the Frasnian Cerberean Volcanics (Item 3), whilst the Mount Morgan tonalite (Item 34) gives a minimum age of 369 ± 5 Ma for the Givetian-Frasnian and in the Early Visean a minimum age of 353 ± 7 Ma (NDS 166, in Odin 1982) comes from the dating of the Garleton Hill Lavas. Taken together a bracket between 355 to 365 Ma seems reasonable for the Devonian–Carboniferous boundary.

Little can be added to the discussion (Gale 1982) of the Upper/Middle and Middle/Lower Devonian boundaries, since the only new datum is Item 34 for the Mount Morgan Tonalite. There seems no reason to depart from the suggestions of 375 ± 5 Ma and 385 ± 8 Ma respectively for these boundaries. For the base of the Devonian Items 7, 8, 9 and 35 are relevant, together with other less reliable data discussed by Gale (1982), and suggest that the boundary lies in the bracket 400^{+10}_{-5} Ma.

For the Silurian new data are the minimum age of 419 ± 10 Ma for the Gorstian (Item 12) and the dates for the

Laidlaw Volcanics (Item 11). Near the base of the Llandovery the dates for the Llandoverian Descon Formation (431 ± 7 Ma, Item 14) and the Ashgillian Stockdale Rhyolite (421 ± 5 Ma, Item 15) overlap and are not inconsistent either with a date of 425 Ma for the base of the Silurian or with a late Llandoverian date of 408 ± 8 Ma (Item 13) or a late Caradoc date of 429 ± 4 Ma (Item 16), or even 419 ± 10 Ma for the Gorstian. It is difficult to accommodate within this series of dates the biotite date of 421 ± 5 Ma for the Laidlaw Volcanics, though the sanidine date of 409 ± 5 Ma fits well. However, as discussed by Wyborn *et al.* (1982), the maximum difference between the estimates of age for the Stockdale Rhyolite and the Laidlaw Volcanics biotite is ~ 8–10 Ma, allowing both for experimental precision and possible interlaboratory bias. It is conceivable that 10 Ma might ultimately prove to be the true age difference between the Ludlow and Ashgill, and this is essentially the situation graphically represented in Fig. 2. Since the preparation of Table 2 and Figs. 1 and 2 a late contribution to this volume by Kunk *et al.* (p. 89) has been brought to the author's attention. They report new $^{40}Ar/^{39}Ar$ dates for the Descon Formation; a total gas age of 433 Ma and a preferred 1200°C step age quoted as 436.2 ± 5.0 Ma (2σ), both of which agree within errors with the original total gas age of 431 ± 7 Ma (2σ) (Item 14). It seems that these ages represent a good minimum age for the *Monograptus Cyphus* Zone in Alaska, and the doubts whether the original data might reflect too old an age due to excess argon are now removed. The same authors report $^{40}Ar/^{39}Ar$ plateau ages of 423.7 ± 1.7 and 422.8 ± 5.8 Ma for biotites from the Gorstian Middle Elton Formation, confirming the data of Item 12. All this data can be accommodated in Fig. 2, though slight changes in the relative lengths of the series are required to effect it comfortably. The position is not completely clear at present, but it can provisionally be suggested that the base of the Silurian lies in the bracket 425^{+10}_{-5} Ma, and that the duration of the Silurian is between 10 and 40 Ma.

Since the discussion by Gale (1982) there have been several new dates reported within the Ordovician; viz. Items 17, 21, 22, 28 (revised stratigraphy), 30, 31 and 33. The Late Tremadoc continues to be poorly dated by imprecise dating of Rhobell Faur at 508 ± 17 Ma (cf. earlier data of 475 ± 12 Ma, NDS122 in Odin 1982). It now appears that the data for Byne Hill (Item 28) suggests that the Didymograptus nitidus Zone of the Arenig is to be dated at 484 ± 4 Ma; together with Items 30 and 33 and basal Arenig dates on glaucony from Estonia (NDS125) and Sweden (NDS132) the modern data is consistent with basal Arenig lying in the interval 490 ± 6 Ma. The base of the Tremadoc remains uncertain.

Items 18, 19, 20, 21, 23, 24 and 29 together with data previously considered by Gale (1982) suggest that the base of the Llandeilo must be near 462 ± 8 Ma and the base of the Caradoc near 452 ± 8 Ma. This large change for the base of the Caradoc, compared with Gale (1982), is necessitated by the new data. The situation is not changed by the revision reported in this volume by Kunk *et al.* of the date of the Tyrone Limestone. The date of 433 Ma (Item 20) originally reported to me by Sutter is apparently incorrect due to erroneous neutron flux monitoring, and has now been corrected to 454.1 ± 2.1 Ma. This change provides a salutary lesson that too much reliance should not be placed on a single date. In fact little weight was given to dates for the Tyrone Limestone in constructing a time-scale by Gale (1982) or Gale & Beckinsale (1983), since the stratigraphy is known only as roughly equivalent to mid-Caradocian.

Comparison with previous time-scales

Table 3 compares these estimates, based on the best available modern data, with some other recently proposed time-scales. In particular it will be seen that new data has largely reconciled the time-scale proposed in this paper and that proposed by McKerrow *et al.* The remaining differences, in the Silurian Period, are *not* to be attributed to undue weight being given, in this paper, to the Stockdale Rhyolite datum; rather they are to be attributed to the attempt to take into account here all data not yet proved to be unreliable. It remains clear that further refinement of the time-scale is needed by the addition of more reliable data, especially in the latest Devonian, earliest Ordovician and in the Silurian Periods; and that the suggestions made in this paper can at best provide an interim solution.

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TABLE 3. Comparison of estimates of the ages of the base of some chronostratigraphic intervals

	McKerrow *et al.* 1980	Odin 1982	Gale & Beckinsale 1983	McKerrow *et al.* this volume	This paper
Carboniferous	360	360^{+5}_{-10}	356 ± 5	354^{+10}_{-5}	360 ± 5
Frasnian	377	375 ± 5	372 ± 5	374	375 ± 5
Eifelian	390	385 ± 8	384 ± 5	391	385 ± 8
Gedinnian	411	400^{+10}_{-5}	398 ± 5	412 ± 5	400 ± 8
Llandovery	438	418^{+5}_{-10}	425 ± 5	435 ± 7	425^{+10}_{-5}
Caradoc	467	438 ± 5	452 ± 5	454 ± 7	452 ± 8
Llandeilo	479	~ 455	462 ± 5	461 ± 7	462 ± 8
Arenig	504	475^{+10}_{-5}	490 ± 5	492 ± 7	490 ± 6

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Age of biostratigraphic horizons within the Ordovician and Silurian systems

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SUMMARY: Three samples that have a bearing on the age of horizons within the Ordovician and Silurian systems, two previously dated by the conventional K-Ar method and one by the $^{40}Ar/^{39}Ar$ total-fusion method, have been reanalysed using the $^{40}Ar/^{39}Ar$ age-spectrum method. Conventional K-Ar and total-fusion $^{40}Ar/^{39}Ar$ ages can always be questioned because of the relative ease with which the K-Ar system can be disturbed, either thermally or chemically (i.e. Dalrymple & Lanphere 1969; Clauer *et al.* 1982). The $^{40}Ar/^{39}Ar$ age-spectrum method has the potential for identifying disturbed K-Ar systems (i.e. Berger 1975; Harrison & McDougall 1980). The authors feel that the age-spectrum data from these samples are significant because the previous results for these samples have been questioned in recently proposed Palaeozoic time-scales because of a possible disturbance of the K-Ar isotopic system (i.e. Gale *et al.* 1979, 1980; Gale 1982).

Analytical method

The $^{40}Ar/^{39}Ar$ age-spectrum method of dating involves the use of mineral standards for monitoring the neutron dose during irradiation (Dalrymple *et al.* 1981). We have chosen to use the standard hornblende, MMhb-1 (Alexander *et al.* 1978) and to accept its age as 519.4 Ma (Dalrymple *et al.* 1981). Conventional K-Ar ages of MMhb-1 measured by both Lanphere and Obradovich do not differ from 519.4 Ma at the 95% confidence level. During this study we also analysed GA-1550 biotite, an intralaboratory standard used at Australian National University by Ian McDougall and co-workers (Joplin 1971; Williams *et al.* 1982). This biotite yields concordant K-Ar and Rb/Sr ages, and the mean K-Ar age of 97.9 Ma was used in this study. The J value (Dalrymple & Lanphere 1971) calculated from the hornblende MMhb-1 and the biotite GA-1550 agree within 0.5% the J values based on GA-1550 but are systematically lower than those based on MMhb-1. The J values quoted in Table 1 of this paper are based solely on MMhb-1, but obviously the data presented here are also comparable with data based on GA-1550.

The term 'weight-average plateau age' is used as described by Fleck *et al.* (1977), except that the test for concordancy on the 'plateau' has been made by comparing the $^{40}Ar_R/^{39}Ar_K$ ratios and their errors using the critical value test described by Dalrymple & Lanphere (1969, p. 120). The error assigned to the weight-average plateau age is simply two standard deviations (2σ) of the mean of all apparent ages making up the 'plateau'. The error assigned to the age of an individual temperature step was estimated using the equation given by Dalrymple & Lanphere (1971) and represents a 1σ estimate of precision. Any significant change in the age assigned to the monitor mineral would change the $^{40}Ar/^{39}Ar$ age of the unknown accordingly.

Description of samples and discussion of results

Esquibel Island, Alaska, USA

Sample 71AE225D hornblende, originally collected and analysed by Lanphere *et al.* (1977), is from a sedimentary breccia in the Descon Formation. The breccia is composed of broken euhedral crystals of hornblende, pyroxene, and plagioclase, as well as volcanic rock fragments, in a graywacke matrix. The crystals and rock fragments are autobrecciated juvenile volcanic material and are not considered to be detrital material from an older volcanic terrain. The breccia lies immediately above a shale sequence containing graptolites of late Ordovician and early Silurian age. The base of this breccia is four metres above shale containing a graptolite fauna characteristic of the *Monograptus cyphus* Zone of Lower Llandovery age (Churkin *et al.* 1971). The age of the breccia thus provides a younger limit for the age of both the *Monograptus cyphus* Zone and the Ordovician–Silurian boundary.

The original age determination of 71AE225D hornblende, performed by Lanphere at the US Geological Survey in Menlo Park, California, used the $^{40}Ar/^{39}Ar$ total-fusion method, which cannot detect a disturbance of the K-Ar isotopic system by either loss or gain of ^{40}Ar (excess ^{40}Ar). Accordingly, this sample was reanalysed by Kunk & Sutter at the US Geological Survey in Reston, Virginia. The analytical results (Table 1) show that this hornblende contains no detectable excess ^{40}Ar and that the hornblende may have lost a very small amount of ^{40}Ar. The total-gas age presented in Table 1 is identical with Lanphere's total-fusion age. The 1200°C step (Table 1), which contains 96% of the argon released from the hornblende, yields an apparent age (preferred age) of 436.2 ± 5.0 Ma (2σ), which agrees within the limits of analytical precision with the original $^{40}Ar/^{39}Ar$ total-fusion results of Lanphere *et al.* (1977). We believe, on this basis, that 436.2 ± 5.0 Ma (2σ) truly represents a minimum age for the *Monograptus cyphus* Zone and the Ordovician–Silurian boundary. By means of the sedimentation-rate model of Lanphere *et al.* (1977), the age of the Ordovician–Silurian boundary can be estimated as at least 439 ± 6 Ma (combining probable errors in the sedimentation-rate model and analytical uncertainty in the age of the 1200°C step of 71AE225D (hornblende)).

Hopedale, Shropshire, England

A bentonite in the Middle Elton Formation, Gorstian Stage, early Ludlow was collected by Ross *et al.* (1982). Obradovich, using the conventional K-Ar method at the US Geological Survey in Denver, Colorado, measured the age of two biotite separates, 76Sh25 and 76Sh25(2), and calculated a mean age of 419 ± 7 (2σ)* for this bentonite. Zircon from this same bentonite was dated by C. W. Naeser at the US Geological Survey in Denver, Colorado. He calculated a fission-track

*Obradovich (letter to Dr E. Calvin Alexander 1981) reported a mean K-Ar age of 516 ± 6 Ma (2σ) for standard hornblende MMhB-1. This differs from the accepted age of 519.4 Ma (Dalrymple *et al.* 1981) by about 0.7%.

TABLE 1. Analytical data

TEMP °C	40-Ar / 39-Ar	37-Ar^{+} / 39-Ar	36-Ar / 39-Ar	$39\text{-}Ar_K$ (% of total)	40-Ar % radiogenic	$39\text{-}Ar_K^{++}$ (moles $\times 10^{-12}$)	APPARENT^{+++} K/Ca (mole/mole)	APPARENT^{++++} AGE (Ma)
	71AE225D hornblende; SE Alaska; Llandoverian, *Monograptus cyphus* Zone							
	J = 0.006627			Sample wt. = 0.0756 g				
550	303	14.1	0.9703	0.3	5.7	0.004	0.04	194.6 ± 146.6
800	199	28.6	0.5793	0.9	15.0	0.0012	0.02	324.4 ± 26.0
1200	50.5	4.51	0.0324	96.1	81.7	1.29	0.12	436.2 ± 2.5
FUSE	426	9.52	1.322	2.7	8.5	0.036	0.05	386.7 ± 114.0
							Total-gas age =	433
							Preferred age =	436.2 ± 5.0(2σ)
	76Sh25 biotite; Shropshire; Ludlovian, *Neodiversograptus nilssoni* to *Lobograptus scanicus* Zone							
	J = 0.006575			Sample wt. = 0.0720 g				
350	117.4	0.095	0.2746	1.1	30.9	0.097	5.5	385.8 ± 4.4
450	76.82	1.92	0.1374	0.5	47.3	0.048	0.27	386.6 ± 5.7
650	47.56	0.155	0.02425	3.3	84.9	0.298	3.4	425.0 ± 2.0
850	42.89	0.011	0.00784	11.9	94.6	1.06	49	426.6 ± 2.0
950	45.29	0.005	0.01583	11.2	89.7	1.00	103	426.9 ± 2.0
1000	45.63	0.009	0.01704	10.7	89.0	0.954	55	426.8 ± 1.9
1050	44.96	0.021	0.01464	10.4	90.4	0.935	24	427.1 ± 2.2
1100	47.47	0.031	0.02392	10.2	85.1	0.915	17	424.9 ± 2.0
1150	50.49	0.036	0.03496	6.6	79.5	0.593	14	422.7 ± 2.0
1200	50.53	0.045	0.03494	5.8	79.6	0.522	12	423.1 ± 2.0
1250	47.00	0.058	0.02272	9.6	85.7	0.862	9.0	423.9 ± 2.1
FUSE	53.30	0.036	0.04421	18.6	75.5	1.67	15	423.4 ± 2.1
							Total-gas age =	424.5
							Weight-average plateau (1100°C-Fuse) age =	423.7 ± 1.7(2σ)
	76Sh25(2) biotite; Shropshire; Ludlovian, *Neodiversograptus nilssoni* to *Lobograptus scanicus* Zone							
	J = 0.006610			Sample wt. = 0.1094 g				
350	39.42	0.016	0.0342	1.3	74.3	0.16	33	319.4 ± 2.2
450	49.75	0.002	0.0295	1.0	82.5	0.13	300	432.8 ± 2.2
650	42.08	0.002	0.0043	6.8	97.0	0.84	300	430.8 ± 2.2
800	41.82	0.001	0.0042	8.5	97.0	1.04	500	428.6 ± 2.2
900	44.68	0.001	0.0139	3.5	90.8	0.43	500	428.4 ± 2.3
1000	43.25	0.001	0.0088	8.0	94.0	0.98	500	429.2 ± 1.9
1050	44.37	0.001	0.0129	10.3	91.4	1.29	500	428.4 ± 2.3
1100	43.78	0.001	0.0098	9.2	93.4	1.13	500	431.4 ± 2.0
1350	41.55	0.002	0.0052	48.3	96.3	5.96	200	423.4 ± 2.0
FUSE	169.4	0.001	0.4385	3.1	23.5	0.38	500	422.0 ± 3.8
							Total-gas age =	425.0
							Weight-average plateau (1350°C-Fuse) age =	422.8 ± 5.8(2σ)

$^{+}$ ^{37}Ar corrected values were determined using a decay constant of 8.25×10^{-4} disintegrations/hour for ^{37}Ar.

$^{++}$ $^{39}Ar_K$ concentrations were calculated using the measured sensitivity of the mass spectrometer and thus have a precision of 10%–20%.

$^{+++}$ Apparent K/Ca ratios were calculated using the equation given in Fleck *et al.* (1977).

$^{++++}$ The isotopic composition of argon was measured with a V. G. Micromass MM1200B mass spectrometer at the US Geological Survey in Reston, Virginia. Samples were irradiated in the Central Thimhle facility of the US Geological TRIGA Reactor in Denver, Colorado and $(^{36}Ar/^{37}Ar)_{Ca}$, $(^{39}Ar/^{37}Ar)_{Ca}$ and $(^{40}Ar/^{39}Ar)_K$ ratios used were those reported by Dalrymple *et al.* (1981). The monitor mineral used in this study was MMhb-1, which has been described by Alexander *et al.* (1978). Constants used in the age calculations are those recommended by Steiger & Jäger (1977).

age of 407 ± 18 Ma (2σ) (Ross *et al.* 1982). Sutter & Kunk have reanalysed the two biotite separates from this bentonite by the $^{40}Ar/^{39}Ar$ age-spectrum method. Although the age spectra of both biotites are somewhat discordant (Table 1 and Fig. 1), their total-gas $^{40}Ar/^{39}Ar$ ages (424.5 and 425.0 Ma) are the same, within analytical uncertainty, as Obradovich's conventional K-Ar mean age of 419 ± 7 Ma (2σ). The $^{40}Ar/^{39}Ar$ weight-average plateau age of 76Sh25 biotite is 423.7 ± 1.7 Ma (2σ), and we believe that this is the best age estimate for this bentonite. Thus, a minimum age of 423.7 ± 1.7 Ma (2σ) for the Gorstian Stage (early Ludlow) is indicated. This is in excellent agreement with the suggested age of 421 ± 2 Ma (2σ) for the Laidlaw Volcanics (Wyborn *et al.* 1982) in Australia which occur at a similiar biostratigraphic level (*Monograptus nilssoni* Zone).

Northern Kentucky, USA

Biotite CM-10, from a bentonite in the uppermost part of the Tyrone Limestone of northern Kentucky of Blackriveran to early Kirkfieldian age (thought to be roughly equivalent to mid-Caradocian), was originally analysed by Sutter by the

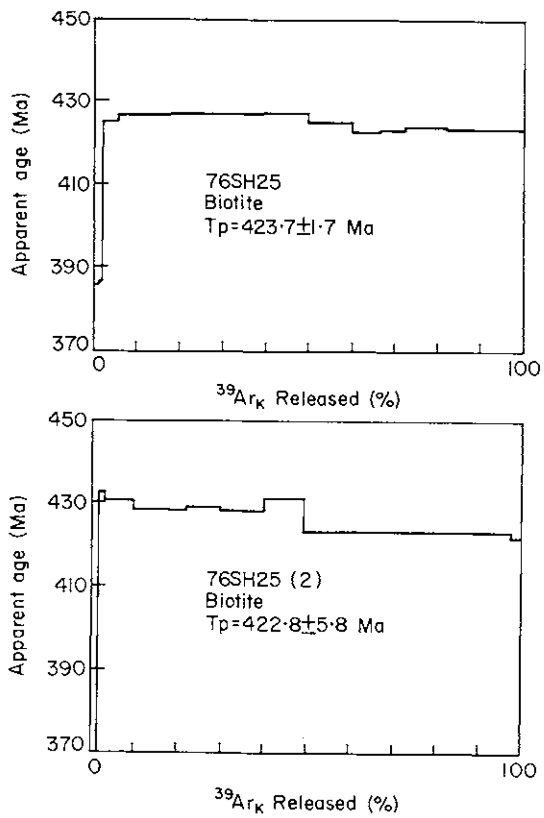

FIG. 1. ^{40}Ar-^{39}Ar Age spectra for Samples No. 76SH25 and 76SH25(2)

^{40}Ar/^{39}Ar age-spectrum method. Although this biotite yielded a weight-average plateau age of 433 ± 10 Ma (2σ), the age was later found to be in error because of faulty neutron-flux monitoring. The biotite has subsequently been reanalysed along with other biotites from bentonites in the midcontinent region of eastern North America at the same stratigraphic level (Kunk & Sutter, 1984). Kunk & Sutter have shown that this stratigraphic level has an age of 454.1 ± 2.1 Ma (2σ) from Alabama to Ohio. However, the original age result of sample CM-10 biotite has been used rather arbitrarily by Gale *et al.* (1979, 1980) and Gale (1982, this volume) in constructing a time-scale for the Palaeozoic.

Conclusions

The analytical results presented here provide additional documentation for the ages of one horizon in the Ordovician and two horizons in the Silurian systems. These ages differ significantly from those suggested by Gale (this volume) for the same stratigraphic levels. In particular, our results suggest a significantly different age for the Ordovician–Silurian boundary. We prefer an age of 439 ± 6 Ma (2σ) for this point, whereas Gale suggests 425 ± 5 Ma.

In addition, we estimate the early Ludlow to be a minimum 423.7 ± 1.7 Ma (2σ) on the basis of biotite from sample 76Sh25. This age, however, is the same as the whole-rock Rb-Sr isochron age obtained by Gale *et al.* (1979) for the Stockdale Rhyolite of Ashgillian age. The results presented here, those for the Laidlaw Volcanics (Wyborn *et al.* 1982), and those for the Stockdale Rhyolite cannot all be correct. Because our results and those of Wyborn *et al.* on samples from the same biozone, widely separated geographically, and analysed by separate laboratories by different methods, yield the same age, we can only conclude that, however precise and analytically accurate, the result for the Stockdale Rhyolite is geologically inaccurate. Compston *et al.* (1982) suggested that the Rb-Sr data of Gale *et al.* (1979) for the Stockdale Rhyolite could be explained by means of a two-stage ^{87}Sr evolution model. This two-stage model suggests that the Stockdale Rhyolite is at least 430 ± 7 Ma (2σ) old and that it probably had a net loss of ^{87}Sr about 412 ± 7 Ma ago.

Finally, we want to set the record straight on the use of the original ^{40}Ar/^{39}Ar data of Sutter's on CM-10 biotite from the Middle Ordovician of North America. Kunk & Sutter (1984) have shown that the bentonites, (including the one from which CM-10 was collected) from the midcontinent region of the USA, which are of Blackriveran to early Kirkfieldian age are 454.1 ± 2.1 Ma (2σ) old. The 433 ± 10 Ma (2σ) age of CM-10 biotite that has appeared in the literature (Gale *et al.* 1979, 1980; Gale 1982, this volume) should be disregarded in the construction of a Palaeozoic time-scale.

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Remarks on the numerical scale of Ordovician to Devonian times

G. S. Odin

SUMMARY: The Early Palaeozoic is the most difficult portion of the time-scale to calibrate due to the low number of precise calibration points although knowledge has greatly increased during the last few years. Thus, two-thirds of the data selected by McKerrow *et al.* (this volume) had not been published in 1980. These new data permit a reassessment of the preceding proposals. The location of some boundaries is still somewhat subjective; this explains differences in recommended or preferred numbers proposed by different authors. However, the correct assessment of the available data by the different authors allow one to define *intervals of time* (better than just numbers), and some sort of 'non disagreement' is now obtained which was not the case earlier. The present comment intends to show the points of agreement, to underline some differences of interpretation, and discuss some data not considered by McKerrow *et al.*

Selection of reliable calibration points

McKerrow *et al.* (this volume) have considered 28 items (the author will note them MLC-00 in the following text) but they seem to use only 23 of them.

MLC-14 is rejected because it was obtained from acid volcanics, MLC-8-13-16 and 19 do not influence the scale because they are not precise enough (fission-track counting). In fact they have fully used less than 23 items. In the discussion they 'effectively eliminate .... the Skiddaw and Shap granites, MLC 23-24, as useful points in the construction of ... their ... scale'. They also nearly eliminate MLC-21-22 (Rb-Sr ages on acid volcanics) in order to be able to use MLC 25 to which they give a very high weight; they thus virtually eliminate four items just to be able to use this age of the Katahdin batholith. As a result, 19 items are considered for a time span of about 150 Ma. Compared to the 45 items used by Forster & Warrington (this volume) for the nearly similar duration of the late Palaeozoic and Triassic the work of elaborating an early Palaeozoic time-scale is not so easy. Moreover, although analytically precise, the item MLC-7 from the Bay of Island is for an intrusion located somewhere in the Arenig, Llanvirn or Llandeilo series, while the four items MLC-2-3-5 and 6 are not precisely located in the stratigraphic sequence. The author therefore fully agrees with McKerrow *et al.* on the important point that, unfortunately, a *definitive time-scale cannot yet be obtained from the presently available data* for the early Palaeozoic times, including the Devonian.

Given this preliminary point of agreement, the author would like to propose to the reader some points of discussion.

The number of items selected is a question of personal judgement. Jones *et al.* in 1981, quoted nearly 50 items of different usefulness to elaborate their 'Silurian and Early Devonian geochronology'. Their final estimates were not very different from those recommended in 'Numerical Dating in Stratigraphy', written nearly at the same time, and discussing nearly 50 dates from the basal Tremadoc to the top Devonian (see Gale 1982a, p. 477; Odin & Gale 1982, p. 492; and relevant items). Those items, taken from this work, will be quoted below under the form NDS-000.

In the present volume, Gale quotes 32 items to elaborate the Tremadoc to Devonian portion of the time-scale, excluding only the 4 items MLC-8-13-16 and 19 based on fission track-counting; this is one-third more than McKerrow *et al.*

The author agrees with McKerrow *et al.* that all these points are not of similar reliability. However, the author suggests that some data, complementary to their limited list, would have helped in choosing numbers without increasing the complexity of the picture. The most interesting of these are;

1. Data from volcanics from top Cambrian-base Tremadoc dated 491 ± 14 Ma (see Vidal & Charlot 1982, NDS 130) in Bohemia.
2. Data from the Rhobell volcanics published before those quoted here and *analytically* more representative: 475 ± 12 Ma (see Rundle *et al.* 1982, NDS 122).
3. Data from the probably mid-Caradoc Tyrone and Carters limestones from USA obtained in several laboratories at different times including very recently (see Kunk *et al.*, this volume) but not available to the authors at the time of writing.
4. Data from the paper by Richards & Singleton (1981). These are from numerous Devonian rocks including mid Devonian and Famennian plutonics for which no data are shown in the Fig. 1 of McKerrow *et al.* (summarized data by Richards, 1982, NDS 233 to 235); several apparent ages, mean at 381 ± 4 Ma, are available there for mid-Devonian formations and numerous other data around 360 Ma, with a low analytical uncertainty, for Famennian plutonics.
5. Data from the post Givetian Hoy lavas by Halliday *et al.* (1982, NDS 244) which conclude to an age of 379 ± 10 Ma for at least part of the Givetian sequence.
6. Data from the basal Carboniferous sequence from France abstracted by Charlot & Vidal (1982, NDS 133) and Peucat & Charlot (1982, NDS 229) which are also discussed in the author's comment of the time-scale of the Carboniferous to Triassic times by Forster & Warrington (this volume).

Quoting these last three groups of data would have lead McKerrow *et al.* to avoid their remark about the very low number of data for the late Devonian times. Indeed, McKerrow *et al.* have possibly rejected these six groups of results with good reasons; however, given the low number of remaining calibration points, a better view of the state of the art would have been obtained in reviewing more data even to show *why* they were not to be used. This would have resulted in a more informative synthesis and possibly lead to slightly different conclusions.

Discussion on the aims of time-scale calibration

McKerrow *et al.* have considered some criteria independent of the radiometric data to evaluate their boundary ages. This needs a comment. If geochronologists try to build a numerical time-scale, it is, mostly, because they hope that this considerable work will have some usefulness to the other geologists. The author's opinion is that the systematic use of hypotheses of extrapolation remove most of the interest of the exercise.

Excluding the simple interest of knowing the numerical age of the units defined by the stratigraphers which, in itself, is gratuitous, the usefulness of a numerized stratigraphical time-scale has two aspects. The first one is the possibility of establishing the contemporaneity of plutonic events with regard to the stratigraphical sequence and, therefore, to be able to have a good picture of the general Earth history. This possibility will not be attained if the ages proposed use uncertain extrapolation systems or if the uncertainties linked with these extrapolation systems — which are very large — are not fully estimated and considered. The second one is the possibility of using the numerical time-scale to estimate the rate (or duration) of geological events: tectonic, biologic or sedimentologic. The use of hypothetic extrapolation systems: biozone duration, sediment thickness, oceanic spreading rate, to deduce the age of different points of the stratigraphic sequence take out half of the interest of the resulting estimates and sometimes lead to circular arguments.

It seems to be better founded, even if the result is still incomplete, to consider first the calibration of the time-scale without using extrapolation systems. When extrapolations are used, the resulting estimates must be *distinctly* considered and, if possible, differently written. The author has chosen to show in parentheses the extrapolated numbers and to recommend intervals of time for boundary estimates obtained from geochronological results. The next comments will just consider geochronological results.

Discussion on the age proposed for the Ordovician to Devonian stratigraphic boundaries

Base of the Tremadoc

McKerrow *et al.* propose for the Cambrian–Ordovician boundary the number of 513 Ma in their Table 1, and the figure of 513 ± 10 Ma in the summary; elsewhere, they note that it could be at 508 ± 5 Ma (if the basal Camhrian is about 530 Ma old as was suggested in Odin *et al.* (1983)); they finally remark that it 'could lie anywhere between 495 and 523 Ma'.

The author must emphasize the very low prohability of the highest number proposed for the top Cambrian because, in this case, the base and the top of the system would in the author's view nearly coincide. Since the first modern data obtained in Morocco by Ducrot & Lancelot (1977) the data are accumulating each year for an age near 530 Ma for somewhere in the Tommotian (basal Cambrian stage). The author's discussion with Gale & Doré (this volume) gathers the most recent evidence in this, including a new series of results by Peucat, previously not considered, and rejects the probability of an age older than 540 Ma. On the other hand, according to the new geological data obtained from the Wrekin area (Ercall granophyre intrusive in the quartzite and *not* covered by them as previously accepted: Beckinsale (1983)) the young age of 520 Ma, for the base of the Cambrian, presently seems too young.

The results from the late Cambrian in Bohemia, quoted above (NDS 130), are probably insufficient alone to define the age of the Cambrian–Ordovician boundary (Ph. Vidal himself has now collected new volcanic rocks from Bohemia to try to obtain more data) but it is compatible with a basal Cambrian at around 530 Ma and the few other available results except item MLC-1 discussed and rejected below. The interval of time 490–505 Ma here preferred (Gale 1982) does not conflict with the 495–523 interval of McKerrow and appears to the author to include the highest probability for location of the Cambrian–Ordovician boundary.

Base of the Arenig

The evaluation of the age of the base of the Arenig series cannot be precise. Items MLC-2 and 3 are only maximum ages; the lower Arenig sequence is younger than 498 ± 7 Ma and 490 ± 14 Ma; the author would settle for younger than 'about 495'. The author disagrees with McKerrow *et al.* who consider their item MLC 1 as 'one of the most significant new dates'. The Rhobell volcanic rocks are certainly stratigraphically favourable: just *at* the boundary, more or less one biozone. Unfortunately, the analytical results accepted in MLC-1 are analytically questionable. Gale (this volume) recalculated an age of 508 ± 17 Ma, instead of 508 ± 11, due to the scattering of the five potassium and argon analyses obtained from a *single* hornblende separated in Cambridge (Kokelaar *et al.* 1982). Beckinsale & Rundle (1980, results summarized in NDS 122 by Rundle *et al.*) suggested a *minimum* age of 475 ± 12 Ma according to *five* hornblendes from similar meta-basalts of the same formation which have admittedly suffered possible argon loss. The large scattering of the results does not allow one to give full confidence to the apparent ages obtained from this formation. In such a situation, it is usual to choose the maximum age obtained (justified by geochemical uncertainties). This leads one to choose the result obtained in the Cambridge laboratory. However, it is appropriate to remember the tendency of this laboratory to produce dates on the high side, as shown in the past (see examples in NDS 18; NDS 165 to 169; de Souza 1982; Patchett *et al.* 1980, p. 653). These interlaboratory direct or indirect comparisons show that K-Ar ages from Cambridge tends to be high compared with results from Berne, Strasbourg, East Kilbride or Leeds, although natural excess argon cannot be rejected in all the apparent ages here questioned. For analytical and geochemical reasons the actual Rhobell volcanic radiometric age remains to be measured more precisely. New results obtained using another dating method would be very interesting.

The author agrees with McKerrow *et al.* that a key point was obtained from the probably Arenig rocks from the Ballantrae area (MLC-4 and NDS 134 by Bluck *et al.* (1982)).

From the above data and other less constraining results (NDS 125-131-132), the author considers it necessary to modify the time interval favoured in 1982 (470 to 485) to the one of 475–495 Ma, which both considers the new results and underlines a somewhat larger absolute uncertainty.

Llanvirn boundaries

There is no new data nor definitive ages to precisely locate the base of the Llanvirn series. The new estimate of McKerrow *et al.*, younger than in 1980, is equivalent to the time interval 460–480 Ma suggested in 1982 by Gale, which may remain unchanged. For the top, the number (455) was suggested in 1982 without the possibility of estimating the uncertainty. The item MLC-10 seems to confirm this number. However, partly due to nearly similar ages obtained from rocks located somewhere in the top Llandeilo to mid Caradoc (MLC-11 and 12) and the author's discussion of the Llandeilo–Caradoc boundary (this volume p. 45), the author recommends the location of the age of the Llanvirn–Llandeilo boundary between 450 and 470 Ma with a preferred number of 460.

Base of the Caradoc

The author has discussed this problem in detail in this volume (p. 44–5). The interval of time 442–458 Ma with a preferred number of 450 seems more appropriate compared to the estimate of 433–443 proposed in 1982 due to the availability of new results.

Base of the Ashgill

There is only one result in the literature obtained from an Ashgillian formation; McKerrow *et al.* quote its apparent age as 421 ± 3 Ma (MLC-14) although the original authors recommended 421 ± 5 Ma (Gale *et al.* 1979; and Gale, this volume). Concerning this item, KcKerrow *et al.* write that 'It is by rejection of this data that the present (MLC) scale differs considerably from the Ordovician and Silurian portions of Odin's scale'. This point, more moderately repeated later in the same discussion needs comments.

First, it is probably a good place here to remember that, although the author agreed to take the responsibility of the scale shown in the author's book, this scale was elaborated in common with several specialists, to whom the author is very indebted and of whom Noël Gale was not the least active, being the author of the Ordovician–Silurian portions of the time-scale. Since that time, Gale has proposed not to weight too heavily the quoted item (Gale & Beckinsale 1983): we should take into account first, the last opinion written by this author.

Second, a rapid examination of the scale proposed by Gale (1982) for the 'Ordovician and Silurian' times shows it to be based on more than one date; two for the Silurian and 18 others for the Ordovician. Consequently, it is not correct to note that the whole 'Ordovician and Silurian portions' of the scale depend on one date though clearly located at a key point.

From this datum (NDS 243 = MLC-14) compared with the less constraining age of the Arisaig volcanics (NDS 239) and the well documented age of the late Caradoc Eskdale granite (429 ± 8 Ma: NDS 189 by Rundle 1982) the interval of time of 425 ± 8 Ma was published in 1982. Looking at the somewhat diverging figures presently available, the author thinks it more prudent not to try to formalize the extreme possible ages of this boundary, which could be very far from each other. The author suggests a rounded number of 435 Ma even though this age fully disagrees both with the apparent ages proposed in the abstracts NDS 189, being too old by five Ma, and with the apparent ages of the Silurian of Alaska (MLC-15 = NDS 128 at 431 ± 6 Ma increased at 436 ± 5 by Kunk *et al.* in this volume), being too young by at least five Ma. This uncomfortable situation remains to be improved by future researches.

Base of the Silurian

The compromise proposed in Gale *et al.* (1980) for the Ordovician–Silurian boundary around 425 Ma seems to be an interesting possibility which tries to reject as few of the available results as possible. It seems, however, too young by possibly 10 Ma looking at the age of the Alaskan hornblendes revised by Kunk *et al.* (this volume). Also Wyborn (pers. comm. 1983 and Wyborn *et al.* 1982) considers that the basal Silurian must be older than 425 Ma due to the nearly similar apparent age of the Australian Laidlaw volcanics of Ludlow age.

These acid volcanics from Australia, give apparent ages at 419 ± 3.8 Ma (two biotite K-Ar ages), 421.1 ± 2.6 Ma (five biotite Rb-Sr ages) and 424.5 ± 15.6 Ma (a 14 points whole-rock Rb-Sr isochron age, the large ± of which indicates a certain inhomogeneity in these ignimbrite flows), but also 408.9 ± 7.2 Ma (three sanidine K-Ar ages). In these effusive rocks, the sanidines give apparent ages considered by the authors to reflect alteration, although their potassium contents are usual (9.8–10.2% K), compared to the biotite apparent ages, considered correct in spite of a clear alteration; potassium content of 6.5 and 5.9% K. The situation is usually the reverse with biotite more rejuvenated than sanidine (Baadsgaard & Lerbekmo 1982). The minimum age of about 421 Ma is accepted by Wyborn *et al.* as the age of these early Ludlow volcanics. The above quoted Ashgill Stockdale rhyolite age of 421 ± 5 Ma was discussed by Compston *et al.* (1982) who suggested another possible interpretation of the data leading to 'an estimate of 430 ± 7 Ma for the age of extrusion'. They consider this age as a minimum but it does not conflict with an estimate at about 425 Ma for the Ordovician–Silurian boundary. The Alaskan data would suggest an older age, as quoted above while the results obtained from the late Llandovery Quoddy formation of USA (NDS 238, apparent age: 408 ± 8 Ma) and the late Ordovician Arisaig volcanics of USA (NDS 239, apparent age at 408 ± 12 Ma) seem to deviate in the other way.

The present data suggest therefore that the Ordovician–Silurian boundary may be around a number such as 425 Ma but could equally well be around 435 Ma or 420 Ma. These numbers are suggested while waiting for the results of new researches which are now being undertaken, under the aegis of the IGCP Project 196, on late Ordovician and Silurian bentonites from several basins. Because of these new researches, and the low number of precise radiometric data (Gale & Beckinsale 1983) available from inside the Silurian system, the author thinks that it is too early to propose estimates for the age boundaries of the stratigraphical units in this system.

The Silurian–Devonian boundary

The Silurian–Devonian boundary was located between 410 and 395 Ma in the author's book with a preference for the number 400 Ma; McKerrow *et al.* propose the more precise interval of time 407–417 Ma, slightly older as a whole, with a preference at 412 Ma. The author's opinion is that stronger arguments are available for a boundary near 400 Ma particularly after looking at several new items quoted by the authors colleagues. The Fig. 1, of McKerrow *et al.*, clearly shows that most of the present results are located on the young side of their line, except MLC-25, which they prefer to four or five others (see above: selection of reliable calibration points). The results which could lead one to prefer the date of 400 Ma, over 412, are as follows:

1. The Lorne lavas from Scotland (MLC-22) gave an age of 400 ± 4 Ma for a Gedinnian formation just above the Silurian–Devonian boundary.
2. The Wormit Bay lavas from Scotland (MLC-21) gave an age of 408 ± 5 Ma for a formation located within the Ludlow or Pridoli series.
3. Concerning the post-Wenlock–pre-Siegenian Gocup granite from Australia, the authors (MLC-20) quote the K-Ar age of 409 ± 3 Ma but not the Rb-Sr age of 402 ± 3 Ma for similar rocks (see NDS 210). It is correct, however, to note that Richards *et al.* (1977) stress that the age of 409 ± 3 Ma was for the Pridoli *or* Gedinnian times 'most probably to the end of the latter'; consequently these results could agree with both numbers 400 or 412.
4. McKerrow *et al.* do not quote the new dates by Gale on the early Gedinnian (although possibly slightly younger) Pembroke formation from the USA (see NDS 237: 397 ± 7 Ma). The correlative Eastport formation from USA gave a confirmatory result of 401 ± 12 Ma (NDS 222, by Fullagar 1982). These formations, or equivalent ones, are intruded by plutonics dated at 390–405 Ma (NDS 236) in New Brunswick, Canada, and Maine, USA. These last dates do not fully eliminate the possibility of an age as old as 412 Ma for the earliest Devonian; but the previous ones suggest a younger age.
5. The possibly Gedinnian Shap granite (MLC-24 see also abstract NDS 241 by Wadge *et al.* 1982) and Skiddaw granite (MLC-23, see also abstract NDS 192 by Rundle 1982) from N. England gave ages of 394 and 399 ± 9 Ma respectively. These two results could reinforce a Silurian–Devonian boundary at about 400 Ma although the author agrees that the stratigraphical constraints are large, i.e. early Gedinnian or younger and that we cannot reject an age as old as 412 Ma. However, the rejection of these data because they do not fit in the extrapolation system of McKerrow *et al.* is a circular argument.
6. Finally, the new results from the late Llandovery Quoddy formation at 408 ± 8 Ma (see NDS 238 by Gale 1982, which refines previous results at 408 ± 40 Ma: NDS 220, by Fullagar 1982) lead the author also to think that the Silurian–Devonian boundary cannot be as old as 417 Ma, the maximum age proposed by McKerrow *et al.* This conclusion is common to many of the items quoted above.

On the other hand, there are three radiometric studies which could agree with a Silurian–Devonian boundary clearly older than 400 Ma.

1. The first concerns the admittedly post-Siegenian Katahdin batholith quoted by the McKerrow *et al.* (MLC-25). The Pb-Pb age of 414 ± 8 Ma and the U-Pb ages of 398 and 395 Ma are quoted but the Pb-Pb age is recommended. This date, from well above the boundary, suggests an age older than 406 Ma, the minimum analytical apparent age recommended, for the Silurian–Devonian boundary. The same batholith was considered by Fullagar (NDS 225) and Armstrong (1978, his item 448). A significantly younger Rb-Sr whole-rock apparent age of 387 ± 16 Ma was obtained, and could have been usefully quoted by McKerrow *et al.*
2. The bentonite of the Elton formation (early Ludlow, from England, MLC-18) was dated at 419 ± 7 Ma (two biotite K-Ar ages). Further confirmatory results are proposed in this volume by Kunk *et al.* which obtained a 40/39 Ar plateau age of 423.7 ± 1.7 Ma (the analytical precision quoted must be increased, by about one percent, to take into account the calibration process). This age, analytically a minimum at about 418 Ma, supports a Silurian–Devonian boundary older than 400 Ma.
3. The Laidlaw volcanics of similar stratigraphic age lead Wyborn *et al.* 1982 (see above) to accept an age of 421 Ma or more for the Ludlow.

As a whole, these results suggest that the Silurian–Devonian boundary could lie at about 410 Ma but that the age of 400 Ma or possibly less is also well supported in several areas. An age as old as 417 Ma, the maximum age accepted by McKerrow *et al.* would reject many data and is considered improbable here. The author therefore proposes to leave unchanged the interval of time 395–410 Ma published in 1982. It is the authors great hope that further studies now undertaken on Ludlow bentonites from Great Britain will contribute to the solution of the remaining uncertainties emphasized above.

Devonian boundaries

Some dates not quoted in McKerrow *et al.* have already been discussed. Looking at the rather similar estimates proposed in this work and the one previously proposed in the author's own book, (see Fig. 1), it is not necessary to comment any further detail here.

However, two general remarks may be made. Due to the use of their system of extrapolation McKerrow *et al.* are able to propose ages for stage boundaries inside subsystems. The author considers this exercise unadvisable in absence of sufficient radiometric data.

The accepted age of 354 Ma, for the top Devonian does not consider some analytical results obtained in the Armoricain Massif and elsewhere, which could support this number; it appears too precise, although possible, and more probable than the age of 365 Ma proposed by Forster & Warrington. (See the relevant comment in this volume p. 116).

Conclusions

The author would like to propose some remarks and emphasize points of agreement with McKerrow *et al.* to conclude this review.

It is useful to recognize the radiometrically undocumented boundaries in order to encourage laboratories and stratigraphers to search for new information. Until such information is available age estimates should be clearly indicated as open.

The error bar quoted in many studies concerns the analytical *reproducibility* which can be very good and less than one percent with the most sophisticated systems of measurement. However, the time-scale calibration needs to compare results from different laboratories and different methods of analysis. The experience shows that the measurement system calibration can hardly be better than ± one percent simply because the interlaboratory reference materials themselves are not known at better than this precision and, furthermore are often not even homogeneous enough (see Odin 1982, and the remarks concerning the standard hornblende MM Hb in the paper by Kunk *et al.*, in this volume). These presently incompressible uncertainties must be added to the analytical precision when comparing different single laboratory age estimates. The result is that an age of about 400 Ma cannot be known better than ± 4 Ma. The rectangles used in a figure such as the figure proposed by

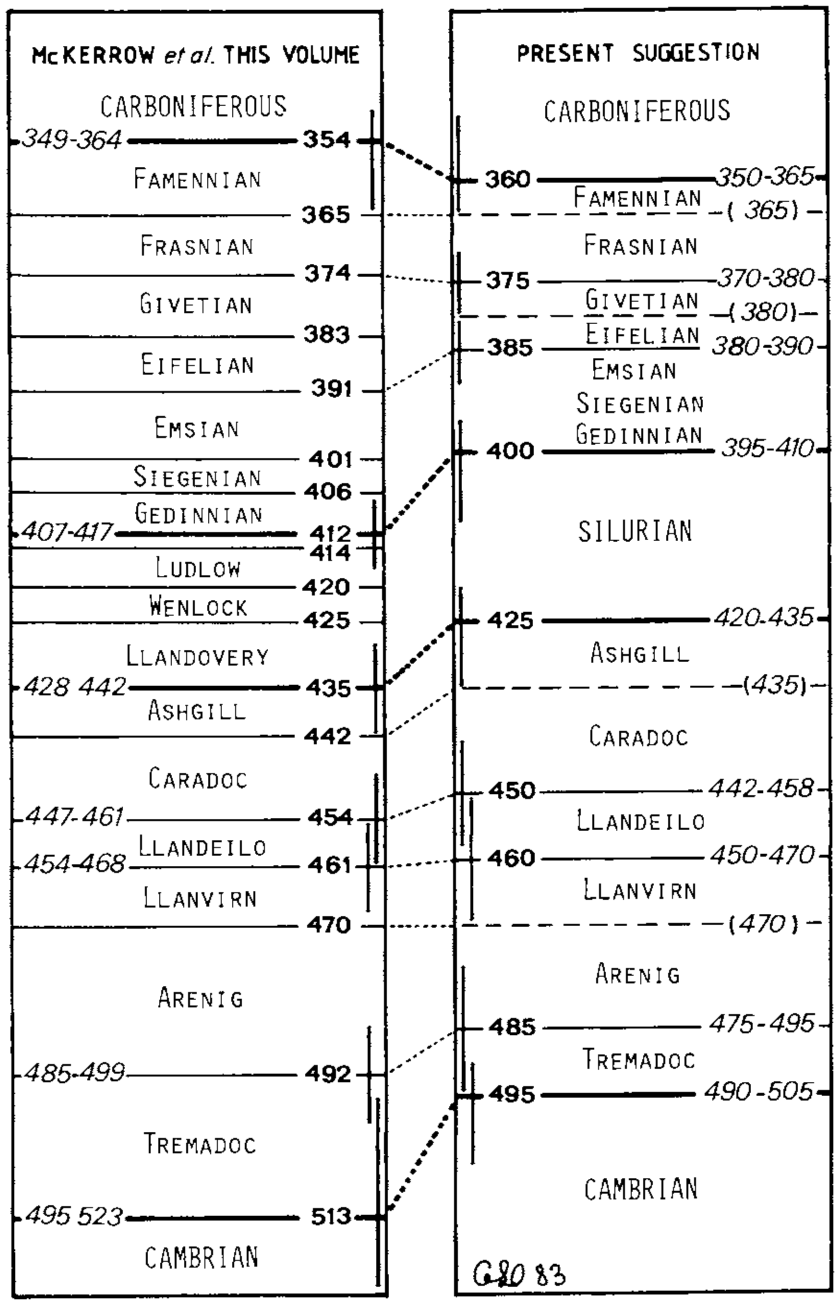

FIG. 1. Numerical time-scale of the Palaeozoic. The scale at the right is recommended. Boundaries are suggested in an interval of time defined according to radiometric dates alone; for the boundaries presently not documented, numbers are suggested in parentheses; the actual tool useful for geological application is proposed in italics: right hand numbers.

McKerrow *et al.* must therefore be increased in some cases so giving a higher degree of freedom when drawing the suggested line (for example MLC-12, 14, 17, 20, 27).

It is essential not to weight too heavily one date, when several others indicate contradictory results, even if the single date appears nearly perfect in the present state of the knowledge.

Other points of agreement include; the short duration of the Silurian times about 25 Ma, or less, and the probability of a short duration for the Cambrian times, much shorter than thought before.

Several series (and even system) boundary age estimates can hardly be considered today as definitive because they are very insufficiently documented.

There is, consequently, a large area of research still open for coordinated studies of stratigraphy and geochronology. Although often taken as a somewhat secondary aspect of geochronological research, the dating of stratigraphically well known formations must become a major goal of research if the geologists actually want to obtain a correct scale for the geological events that they observe.

ACKNOWLEDGEMENTS: I am grateful to Norman Snelling who was kind enough to improve my English.

Reply by McKerrow, Lambert and Cocks

It is clear that more exact data are still needed to produce a time-scale which will be reliable enough to discuss detailed rates of geological and biological processes during the Ordovician, Silurian and Devonian periods.

There has already been considerable published discussion on many of the points raised by Professor Odin (Gale *et al.* 1980; McKerrow *et al.* 1980; Williams *et al.* 1982; Gale & Beckinsale 1983; Odin *et al.* 1983), and we have little to add now. Several of Professor Odin's criticisms refer to data which we used in our 1980 paper, and which we have now rejected either on stratigraphical or chemical grounds. We see no great point in cluttering up our present discussion with such data.

We do not claim to have produced more than an interim time-scale based on the data available at present. The new work that has been done since our 1980 paper was published illustrates the merits of an interim time-scale; workers can see where the gaps are, and try to fill them. We are certain that, with the continued interest in these problems, more data will be forthcoming.

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Geochronology of the Carboniferous, Permian and Triassic

S. C. Forster and G. Warrington

SUMMARY: Criteria concerning sample and analytical data and stratigraphic control have been used in a critical assessment of the suitability, for use in construction of a Phanerozoic time-scale, of radiometric data relevant to the Carboniferous, Permian and Triassic periods. Few of the age determinations available in 1982 satisfy these criteria and many of those used previously as a basis for time-scales for this part of the Phanerozoic are considered unacceptable by present standards. On the basis of this review, ages of 365 ± 5 Ma, 290 ± 5 Ma and 250 ± 5 Ma, respectively, are proposed for the beginning of the Carboniferous, Permian and Triassic periods, and 205 ± 5 Ma for the end of the Triassic. Radiometric ages are related, where possible, to the principal chronostratigraphic divisions of the rock successions representing those periods.

A prerequisite for the establishment of a reliable geochronological time-scale for any part of the Phanerozoic is the availability of accurate radiometric age data from rocks whose stratigraphical position is clearly defined and whose ages have not been re-set by subsequent events.

The suitability of the available sets of data for use in the construction of a radiometric time-scale for the Carboniferous, Permian and Triassic periods has been assessed by application of the following criteria:

1. *The stratigraphic position of the dated rocks or minerals must be known.* Ideally, samples for radiometric dating should be from extrusive igneous rocks intercalated with fossiliferous sediments which are precisely dated by independent biostratigraphic means. Such rocks are, however, often altered and may have suffered partial loss of radiogenic daughter products. Intrusive igneous rocks, as well as reworked volcanics, metamorphic rocks and most detrital sediments, give only maximum or minimum ages but tend to yield more reliable radiometric data because they are more likely to have formed stable, closed isotopic systems.
2. *The radioisotopic data should be as precise and unambiguous as possible.* Ages should be derived from completely undisturbed systems and be confirmed by isochron techniques or by a number of separate conventional K-Ar, Rb-Sr or fission track determinations. If the chemistry of the rock allows, several different dating methods should be used on rocks and minerals from the same geological environment. The data should be internally consistent and should not be in conflict with the known geological sequence.
3. *Complete analytical details should be published.* It is important that the quality of any results can be assessed independently and that the original data can, if necessary, be recalculated using new techniques or regressions. Details of the decay constants used and the manner in which the errors have been assessed are essential.
4. *The petrography of samples used for whole-rock determinations should be described so that the suitability of the samples can be independently assessed.* In the case of determinations from minerals, only those from species normally regarded as reliable age indicators (Dalrymple & Lanphere 1969; Faure 1977) should be used.

A large number of age determinations on rocks of Carboniferous to Triassic age have been published. In this review, the radiometric data available in nearly 500 separate articles have been examined by the senior author (S. C. Forster) and, following application of the above criteria, only 45 dated items (Fig. 1) have been accepted from this voluminous literature as suitable for time-scale purposes. The majority of the published ages have been discarded because of lack of analytical information or lack of stratigraphical constraint, or because of the obviously discrepant nature of the data presented. Single, unconfirmed, conventional age determinations have frequently been used to provide key marker points on a radiometric time-scale; such data are useful in terms of geochronological reconnaissance but cannot, until confirmed by further research, be used to define points on a time-scale. Much of the data used in the 1964 Geological Society Phanerozoic Time-scale (Harland *et al.* 1964) fails to meet the criteria applied here and is, by modern standards, inaccurate, imprecise and internally inconsistent. Its continued inclusion in reviews such as that by Armstrong (1978) is unjustified and misleading. Many of these dates have been superceded or refined by modern techniques and, rather than continuing to discuss such data, attention is directed here to newer, well-correlated radiometric data which satisfy as many of the above criteria as possible, and to the identification, for future work, of parts of the time-scale where similar reliable data are currently lacking.

Lambert, reviewing the Phanerozoic Time-scale in 1971, commented that the radiometric chronology of Carboniferous rocks was better known than that of any other pre-Tertiary period. However, for substantial parts of the Carboniferous, and also of the Permian and Triassic periods, there are no reliable radiometric marker points. Where good radiometric data are available, the stratigraphic constraints are often too wide to make the dates significant, and in other instances the experimental error on dated rocks which are stratigraphically well-localized may be wide and preclude dating of many of the stage boundaries within each system.

The ages quoted in this paper have been recalculated where necessary using the standard decay constants recommended by the IUGS Subcommission on Geochronology (Steiger & Jäger 1977). In general, this has resulted in a better correspondence between the ages obtained by different dating methods and has resolved some of the difficulties encountered in previous reviews of Carboniferous to Triassic geochronology in which various ages were reported and often alternative time-scales erected depending on which decay constant was used for ^{87}Rb (see, for example, Banks 1973 and Waterhouse 1978).

It is important to compare all analytical results at the same, 2 sigma, level of precision and, wherever possible, the ages quoted here show the error at that level (95% confidence limits). Exceptions have been made where no errors were quoted in the original publication or where no details were

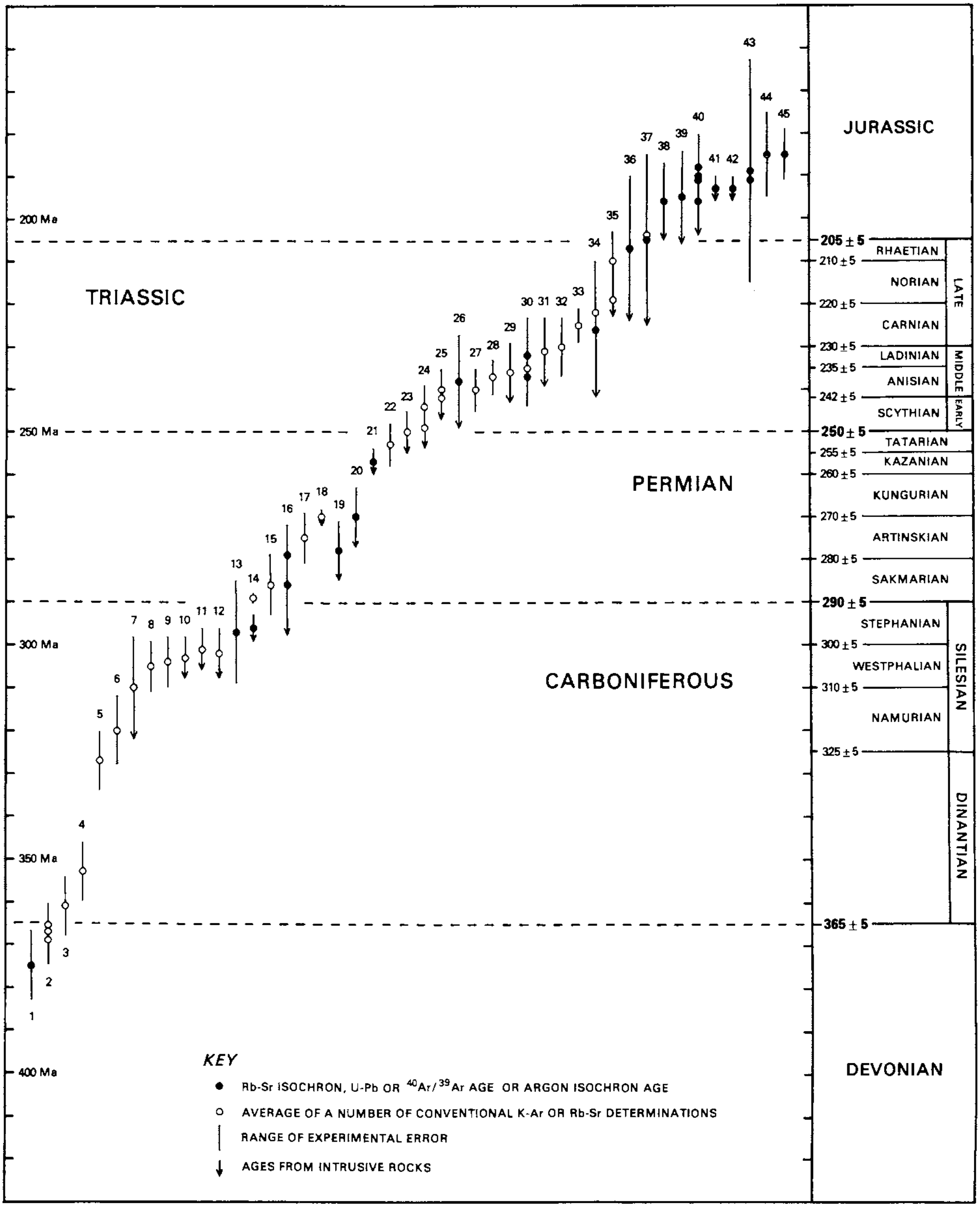
200 Ma
250 Ma
300 Ma
350 Ma
400 Ma
JURASSIC
TRIASSIC
PERMIAN
CARBONIFEROUS
DEVONIAN
205 ± 5
RHAETIAN
210 ± 5
NORIAN
220 ± 5
CARNIAN
230 ± 5
LADINIAN
235 ± 5
ANISIAN
242 ± 5
SCYTHIAN
LATE
MIDDLE
EARLY
250 ± 5
TATARIAN
255 ± 5
KAZANIAN
260 ± 5
KUNGURIAN
270 ± 5
ARTINSKIAN
280 ± 5
SAKMARIAN
290 ± 5
STEPHANIAN
300 ± 5
WESTPHALIAN
310 ± 5
NAMURIAN
SILESIAN
325 ± 5
DINANTIAN
365 ± 5
KEY
Rb-Sr ISOCHRON, U-Pb OR $^{40}Ar/^{39}Ar$ AGE OR ARGON ISOCHRON AGE
AVERAGE OF A NUMBER OF CONVENTIONAL K-Ar OR Rb-Sr DETERMINATIONS
RANGE OF EXPERIMENTAL ERROR
AGES FROM INTRUSIVE ROCKS

FIG. 1.

given of the way in which the experimental error was calculated. Since data in these categories should probably be excluded from consideration by strict application of the criteria listed above, less significance is attached, in the following discussion, to ages from such sources. The MSWD is quoted, where available, for all Rb-Sr isochron ages; it is not available for some of the older published isochrons. The commonest and most important source of uncertainty associated with many of the available data is, however, geological and is unquantifiable in terms of a precise error. The effects of analytical uncertainties are commonly less important than the accurate interpretation of the geological event that is recorded in each individual situation.

The Carboniferous

The age data used here for the Carboniferous period are largely from the British Isles and the stratigraphic definitions and nomenclature used follow those proposed by George *et al.* (1976) for the Lower Carboniferous (Dinantian subsystem) and by Ramsbottom *et al.* (1978) for the Upper Carboniferous (Silesian subsystem).

The Devonian–Carboniferous boundary

Radiometric data are available from rocks above and below the Devonian–Carboniferous boundary in Scotland and Australia but the stratigraphic relationships in the two areas are not easily reconciled with the reported age data. Since the age proposed herein for the boundary is slightly older than that advocated by McKerrow *et al.* (1980 and this volume), Gale *et al.* (1980) and Odin & Gale (1982), the radiometric and stratigraphic evidence will be considered in some detail. In our view, the value of McKerrow *et al*'s (1980 and this volume) and Gale *et al*'s (1980) correlation lines for the Lower Palaeozoic is debatable, particularly when extrapolated to the Devonian–Carboniferous boundary, because their technique, which involves plotting apparent relative stage lengths against the radiometric data, allows a number of alternative correlation lines to be derived from the same data. The age proposed in this contribution is based on a direct assessment of the radiometric data and the stratigraphic constraints on those data.

In both the Scottish Borders and south-east Australia the biostratigraphic boundary between the Devonian and Carboniferous is placed within unfossiliferous red-bed sequences.

A good marker point for the Upper Devonian is provided by the Hoy lavas of Orkney (Item 1) which have been dated by Halliday *et al.* (1977, 1979) using the $^{40}Ar/^{39}Ar$ method. The lavas rest unconformably on folded and faulted Middle Old Red Sandstone sediments which, in the Orcadian Basin, range up to Givetian in age; they are overlain by the Hoy Sandstone which is of probable Upper Old Red Sandstone age. Halliday *et al.* (op. cit.) suggested an age of approximately 368 Ma for the Givetian–Frasnian boundary, this being the average value obtained from two $^{40}Ar/^{39}Ar$ stepwise degassing analyses. However, one of these shows evidence of some disturbance of the argon isotopes and only data from the Hellia basalt, which gave a $^{40}Ar/^{39}Ar$ plateau age of 375 ± 8 Ma, are considered acceptable. This should be regarded as a minimum age for the basalt since both analyses showed evidence of slight argon loss.

The Cerberean Cauldron Volcanics (Item 2) of central Victoria, Australia, have frequently been used to provide data close to the Devonian–Carboniferous boundary (McDougall *et al.* 1966). Recent work by Richards & Singleton (1981) and Williams *et al.* (1982) has resulted in a refinement of the age reported by McDougall *et al.* (1966), and in a revision of the biostratigraphical position of the dated horizon. Richards & Singleton (1981) refined some of the earlier K-Ar determinations on the biotite rhyodacite

FIG. 1. Carboniferous –Triassic time-scale. Age data used in Fig. 1.
(1) Hoy lavas, Orkney. Hellia basalt, $^{40}Ar/^{39}Ar$ plateau age of 375 ± 8 Ma. (2) Cerberean Cauldron Volcanics. Biotite rhyodacite, K-Ar average ages of 367 ± 1, 369 ± 6 and 365.6 ± 4.8 Ma; Rb-Sr mean age of 367.1 ± 2.3 Ma. (3) Birrenswark and Kelso lavas, Scottish Borders. Average K-Ar age of 361 ± 7 Ma. (4) Garleton Hills lavas, East Lothian. Average K-Ar age of 353 ± 7 Ma. (5) Upper Clyde Plateau lavas, Strathaven. Average K-Ar age of 327 ± 7 Ma. (6) Machrihanish lavas, Kintyre. Average K-Ar age of 320 ± 8 Ma. (7) East Fife Sill Complex. Average K-Ar age of 310 ± 12 Ma. (8) Passage Group lavas, Ayrshire. Average K-Ar age of 305 ± 6 Ma. (9) Ruhr tonstein. Average K-Ar age of 304 ± 6Ma. (10) Midland Valley Sill Complex, Scotland. Average K-Ar age of 303 ± 5 Ma. (11) Whin Sill, northern England. Average K-Ar age of 301 ± 5 Ma. (12) Brittany granites. Average K-Ar age of 302 ± 6 Ma (biotite). (13) Bulgonunna Volcanics, Australia. Rb-Sr isochron age of 297 ± 12 Ma. (14) Bulgonunna granites, Australia. Rb-Sr isochron age of 296 ± 3 Ma; K-Ar average mineral age of 289 ± 1 Ma. (15) Mauchline Volcanics, Ayrshire. Average K-Ar age of 286 ± 7 Ma. (16) Great Serpentine Belt, New South Wales. $^{40}Ar/^{39}Ar$ total fusion ages of 279 ± 12 and 286 ± 12 Ma. (17) Lizzie Creek Volcanics, Australia. Average K-Ar age of 275 ± 6 Ma. (18) Lizzie Creek granites, Australia. Average K-Ar age of 270 ± 2 Ma. (19) Nordagutu granite, Oslo Rift. Rb-Sr isochron age of 278 ± 7 Ma. (20) Oslo Suite plutons. Rb-Sr isochrons of 270 ± 7 and 270 ± 1.7 Ma. (21) Quillabamba granite, South America. U-Pb age of 257 ± 3 Ma. (22) Sydney Basin latites, Australia. Average K-Ar age of 253 ± 5 Ma. (23) Marlborough gabbros and diorites, Australia. Average K-Ar age of 250 ± 5 Ma. (24) South-east Queensland granites. Average K-Ar ages of 244 ± 5 Ma (hornblende) and 249 ± 5 Ma (biotite). (25) Granodiorite/adamellite, Marlborough, Australia. Average K-Ar ages of 242 ± 5 Ma (hornblende) and 240 ± 5 Ma (biotite). (26) Coasa granite, South America. U-Pb age of 238 ± 11 Ma. (27) Neara Volcanics, Australia. Average K-Ar age of 240 ± 5 Ma. (28) Puesto Viejo Formation, Argentina. Mean K-Ar age of 237 ± 4 Ma. (29) Goomboorian diorite, Australia. Average K-Ar age of 236 ± 7 Ma. (30) Grenzbitumenzone bentonites, Switzerland. High sanidine: average K-Ar age of 232 ± 9 Ma, $^{40}Ar/^{39}Ar$ plateau age of 232 ± 8 Ma. High sanidine + overgrowth: average K-Ar age of 235 ± 6 Ma, $^{40}Ar/^{39}Ar$ plateau age of 237 ± 7 Ma. (31) Monzoni intrusive complex, Italy. Average K-Ar age of 231 ± 8 Ma. (32) Sugars basalt, Australia. Average K-Ar age of 230 ± 7Ma. (33) Val Serrata tuff, Switzerland. Mean K-Ar age of 225 ± 4 Ma (sanidine). (34) Maryborough Basin granites, Australia. Rb-Sr isochron age of 226 ± 16 Ma; average K-Ar age of 222 ± 2.5 Ma. (35) Indonesian tin granites. Average K-Ar mineral age of 219 ± 4 Ma. Average Rb-Sr age of 210 ± 7 Ma. (36) Kulim granite, Malaysia. Rb-Sr isochron age of 207 ± 17 Ma. (37) Guichon Creek batholith, British Columbia. Average K-Ar age of 204 ± 5 Ma; Rb-Sr isochron age of 205 ± 20 Ma. (38) Palisades Sill, New Jersey. $^{40}Ar/^{39}Ar$ plateau age of 196 ± 9 Ma. (39) Mt. Carmel Sill, Newark Basin, USA $^{40}Ar/^{39}Ar$ plateau age of 195 ± 4.2 Ma, argon isochron age of 195 ± 11 Ma. (40) Liberian dykes, West Africa. $^{40}Ar/^{39}Ar$ plateau age of 191 ± 6 Ma, argon isochron age of 196 ± 8 Ma (plagioclase); $^{40}Ar/^{39}Ar$ plateau age of 188 ± 8 Ma., argon isochron age of 190 ± 5 Ma (whole rock). (41) Freetown Igneous Complex, Sierra Leone. Rb-Sr isochron age of 193 ± 3 Ma. (42) Initial Karroo volcanism (Pronksberg andesite). K-Ar isochron age of 193 ± 3 Ma. (43) Bowser Basin volcanics, British Columbia. Rb-Sr isochron ages of 191 ± 18 and 189 ± 26 Ma. (44) Toodoggone Volcanics, British Columbia. Rb-Sr isochron age of 185 ± 10 Ma. (45) Hazleton Group Volcanics, British Columbia. Rb-Sr isochron age of 185 ± 6 Ma.

which occurs at the top of the > 900 m thick sequence of felsic lavas comprising the Cerberean Volcanics; partial re-analysis of the biotite sampled by McDougall *et al.* (1966) gave a mean K-Ar age of 367 ± 1 Ma, close to the K-Ar average age of 369 ± 6 Ma and the Rb-Sr biotite ages of 358 ± 7 Ma and 365 ± 7 Ma obtained by those workers. Analyses of the same biotite by Williams *et al.* (1982) have given a mean K-Ar age of 365.6 ± 4.8 Ma and a Rb-Sr mean age of 367.1 ± 2.3 Ma. The dated rhyodacite horizon overlies sediments which contain *Bothriolepis* and *Phyllolepis*, a fish fauna of Frasnian (late Devonian) age (Marsden 1976; Richards 1978). Long (1982) has suggested that this is the oldest of a number of Frasnian assemblages known from Victoria and is possibly of early Frasnian age.

The volcanics of the Cerberean Cauldron occupy only one of a number of cauldron subsidence areas in central Victoria where extrusive volcanic activity was widespread and voluminous during the late Devonian, and was followed by the intrusion of numerous high-level plutons. Red-beds of the late Devonian to early Carboniferous Mansfield Group were deposited in a chain of fault-bounded basins along the eastern margin of the area. Although many of the igneous rocks in the area have been dated (Richards & Singleton 1981) those of the Cerberean Cauldron still appear to have the best stratigraphic correlation, since most of the other cauldrons of similar radiometric age contain few fossiliferous sediments.

The late granites which cut the cauldron volcanics include the Lysterfield granodiorite with an average K-Ar age of 364 ± 7 Ma and many other granites with average K-Ar ages of about 360 Ma. These ages are close to the Rb-Sr age of 365 Ma (muscovite; no errors quoted) from the Strathbogie granite, which has an average K-Ar age of 362 ± 4 Ma, and to a Rb-Sr isochron age of 368 ± 3 Ma (MSWD = 0.48) from a granite in the Terricks Range which has an average K-Ar age of 361 ± 3 Ma (Richards & Singleton 1981). The only granite which appears to both intrude the volcanics and to be demonstrably overlain by the red-beds of the Mansfield Group is the Barjarg granite, thought to be one of the last to be emplaced. It is the youngest pluton in the Tolmie Igneous Complex and occurs some 40 km north-east of the Cerberean Cauldron (Marsden 1976). A Rb-Sr isochron age of 382 ± 11 Ma was reported from this granite by McDougall *et al.* (1966). Richards & Singleton (1981) reported a Rb-Sr mineral isochron age of 366 ± 21 Ma from the same samples; weathered biotite gave a Rb-Sr age of 360 Ma. The Barjarg granite is thought to have been exposed fairly rapidly and before the deposition of the Mansfield Group, which overlies Upper Devonian sediments and volcanics and contains an early Carboniferous non-marine fauna and flora in its upper part; the lower part of the group may be of late Devonian age. However, it is thought that most of the post-orogenic deposition followed the intrusion of the granites relatively quickly (Marsden 1976). Richards & Singleton (1981) concluded that the Frasnian is no younger than 367 ± 1 Ma, and that the Devonian–Carboniferous boundary must be younger than the 360 to 365 Ma ages obtained from the granites in Victoria. Williams *et al.* (1982) reached a similar conclusion but emphasized that the ages obtained from the Cerberean Volcanics must be regarded as minima which set a lower limit to the age of the early Frasnian, a view with which we concur.

In the Scottish Border region of Great Britain, the Devonian–Carboniferous boundary occurs within the unfossiliferous red-bed sequence of the Old Red Sandstone. The Birrenswark and Kelso lavas (Item 3), with a minimum K-Ar average age of 361 ± 7 Ma (De Souza, 1982), underlie rocks of Courceyan age (the Lower Border Group and Cementstones) in the Northumberland basin (George *et al.* 1976). They overlie fluviatile Upper Old Red Sandstone sediments of Famennian age (House *et al.* 1977) and follow a period of non-deposition which may have lasted for some 10^4 to 10^6 years (Leeder 1976, 1982 Fig. 8) and during which calcrete palaeosols formed at the top of the Old Red Sandstone. No diagnostic fossils have been found in the northern part of the Northumberland Trough in sediments contiguous with the volcanics but Courceyan miospores and ostracod faunas occur in the Cementstones elsewhere in the Northumberland Basin (Neves *et al.* 1972; George *et al.* 1976).

The Birrenswark and Kelso lavas are taken locally to mark the base of the Carboniferous but it seems likely that, in this area, the red-bed facies of the Old Red Sandstone persisted into Carboniferous times (George *et al.* 1976). Thus, the lavas are regarded as almost certainly early Carboniferous in age and the minimum age of 361 ± 7 Ma may relate to an horizon some distance above the Devonian–Carboniferous system boundary.

The age of the Devonian–Carboniferous boundary must be interpolated between an upper limit of at least 367 ± 1 Ma, based upon data from the early Frasnian Cerberean Cauldron Volcanics, and a lower limit of at least 361 ± 7 Ma based upon the data from the Birrenswark and Kelso lavas at an undefined level in the Courceyan. On this evidence, the Frasnian and Famennian appear to have occupied a relatively brief period of time but the data available provide only minimum ages; an age of 365 ± 5 Ma is proposed for the Devonian–Carboniferous system boundary.

The Dinantian

Radiometric ages for British Lower Carboniferous successions were reviewed by George *et al.* (1976); satisfactory data are scarce and it is impossible, at present, to date the boundaries of stages within the Dinantian with any degree of confidence. However, the start of the major alkali-basaltic volcanism in the Midland Valley of Scotland in mid-Dinantian (Arundian) time can be dated fairly accurately using data from the Garleton Hills lavas of East Lothian (Item 4) which are supported by dates from other volcanics at a similar stratigraphic level. Sanidine from trachytes in the upper part of the Garleton Hills succession has an average K-Ar age of 353 ± 7 Ma (De Souza 1982); these lavas occur between sequences assigned to the Pu and TC miospore zones of Neves *et al.* (1973) and are of Holkerian to Asbian age (George *et al.* 1976). The Arthur's Seat volcanics of Edinburgh occur at a similar stratigraphic level and a K-Ar age of 354 ± 7 Ma has been reported for a lava at the base of that succession (Fitch *et al.* 1970); tuffs amongst these volcanics also contain Pu zone (i.e. late Courceyan to Holkerian) miospores (Neves *et al.* 1973). These ages suggest that the Tournaisian–Visean series boundary is no younger than 355 Ma. Further support for the age of the mid-Dinantian volcanic activity comes from the Sm/Nd age of 356 ± 10 Ma (Van Breemen & Hawkesworth 1980) reported from a granulite-facies gneiss xenolith in the Partan Craig vent which cuts tuffs of the East Lothian volcanics, including the Garleton Hills lavas referred to above. Van Breemen & Hawkesworth (1980) interpreted the age as reflecting the onset of the volcanic activity which brought it to the surface in early Carboniferous times.

The Clyde Plateau lavas form an extensive lava pile covering most of the western part of the Midland Valley of Scotland, and rest unconformably on sediments ranging in age from Devonian to mid-Dinantian (? Chadian); they are overlain by sediments of latest Dinantian age. Over 35 mineral and whole-rock conventional K-Ar age determinations on Clyde Plateau lavas and associated intrusions by De Souza (1982) gave ages ranging from 332 Ma to 326 Ma on the least altered samples; De Souza suggested that this span of ages corresponds with the time taken for the lava pile to accumulate. Lavas at the top of the succession near Strathaven (Item 5) have an average K-Ar age of 327 ± 7 Ma (anorthoclase). Lavas and tuffs of similar age nearby are overlain by a Brigantian limestone, so that this date provides a minimum age for the late Dinantian.

A minimum age of 320 ± 8 Ma (K-Ar) has been obtained from an outlying section of the Clyde Plateau lavas on the Kintyre peninsula (De Souza 1982); there, the Machrihanish lavas (Item 6) rest unconformably on Devonian and older rocks and are overlain by sediments containing faunas of possible Dinantian or Namurian age (George *et al.* 1976). On the basis of the age of 327 ± 7 Ma from the upper Clyde Plateau lavas, which are overlain by uppermost Dinantian sediments, an age of 325 ± 5 Ma is proposed for the Dinantian–Silesian boundary.

The Silesian

Very few reliable radiometric data are available for rocks of Namurian age. A K-Ar average age of 305 ± 6 Ma has been reported from the Passage Group lavas of Ayrshire (Item 8) which are post-Namurian A and pre-Westphalian A (De Souza 1982); this can only be regarded as a low minimum age since the lavas analysed showed alteration.

The East Fife Sill complex (Item 7), in the north-east of the Midland Valley of Scotland, has an average K-Ar whole-rock age of 310 ± 12 Ma (Forsyth & Rundle 1978); the sills, which are of fairly limited lateral extent, cut Carboniferous strata ranging in age from late Dinantian to Namurian (E_2). Farther west in Fife, similar sills are also found to cut only pre-Westphalian strata and, therefore, a Namurian or early Westphalian age seems possible. Francis (1967, 1968) has shown that the sills have a close genetic association with tuffs ranging in age from uppermost Dinantian to mid-Namurian. The average age of 310 ± 12 Ma therefore provides a minimum age for the Namurian; Forsyth & Rundle (1978) concluded that the emplacement of the East Fife Sill complex occurred at the older end of this age range and may well have been 322 Ma or older.

The most reliable data available for the Westphalian are from the Ruhr in Germany (Damon & Teichmüller 1971); primary sanidine from a tonstein horizon in the coal seam Hagen 2, of Westphalian C age, gave an average K-Ar age of 304 ± 6 Ma (Item 9). This provides a good minimum age for Westphalian C deposits.

The Whin Sill of northern England (Item 11) occupies an historic position as a marker point on the geological time-scale (Dubey & Holmes 1929; Holmes 1931); this intrusion is bracketed stratigraphically between Coal Measures of Westphalian B to C age and Upper Brockram sediments of probable early Permian age. The average K-Ar whole-rock age of 301 ± 5 Ma obtained by Fitch & Miller (1967) and Tarling *et al.* (1973) therefore provides a minimum age for the Westphalian B to C rocks of northern England. Confirmation of this radiometric data comes from the co-magmatic Midland Valley Sill Complex of Scotland (Item 10) which has an average K-Ar whole-rock age of 303 ± 5 Ma (Fitch *et al.* 1970). Francis (1978) has correlated the tholeiitic Whin Sill and Midland Valley Sill Complex with a north-west trending dyke swarm in Skåne, southern Sweden, which has similar petrographic features and from which Klingspor (1976) reported a K-Ar isochron age of 300 ± 4 Ma based on data from 21 samples.

From the limited amount of data available, an age of 310 ± 5 Ma is proposed for the Namurian–Westphalian boundary, since this shows the best agreement with the data from the Westphalian.

There are few stratigraphically-controlled dates from post-Westphalian Carboniferous rocks. The two-mica granites of southern Brittany (Item 12) have given average K-Ar biotite and muscovite ages of 302 ± 6 Ma and 321 ± 6 Ma respectively, and a Rb-Sr isochron age of 331 ± 9 Ma (Vidal 1973; Ries 1979). Deformed pebbles of an identical granite occur in Stephanian molasse sediments in western Brittany and the biotite age of 302 ± 6 Ma is therefore regarded as a maximum age for the Stephanian. A large number of age determinations is available for the granites of south-west England (Miller & Mohr 1964; Dodson & Rex 1971; Harding & Hawkes 1971) and for the post-orogenic volcanics in the New Red Sandstone in Devon (Miller & Mohr 1964; Hawkes 1981), but the poor stratigraphic control on these intrusions and volcanics makes them of little use for the construction of a time-scale.

From the limited amount of data available it is difficult to define an age for the Westphalian-Stephanian boundary with any precision but an age of about 300 ± 5 Ma seems likely and is not in conflict with any of the radiometric data available.

The Permian

The majority of Permian stages are based upon Tethyan and low-latitude marine successions which occur in the southern United States and in an east-west trending central Asian belt linked, northwards, with the Urals region. Several chronostratigraphic schemes are currently in use and reflect the importance attached in different areas to various fossil groups, notably ammonoids (Furnish 1973), fusilinid foraminifera (Stepanov 1973), brachiopods (Waterhouse 1978) and conodonts (Kozur 1981). A three-fold division of the sequence into early, middle and late Permian is commonly adopted though divisions of the succession into two and four series or subsystems have also been proposed. There is some agreement that the base of the system should be defined at the appearance of the fusilinid *Schwagerina* in the succession in the Urals.

The radiometric data used here for the Permian period are from Australia and, to a lesser extent, from South America and north-west Europe, areas remote from those in which the various Permian stages are defined. The Russian stage sequence from the Urals was, however, adopted by Dickins (1976) in a recent review of Australian Permian successions and, because of the preponderance of Australian age data used here, is followed in this account.

The Carboniferous–Permian boundary

The Mauchline Volcanics of Ayrshire, Scotland (Item 15), and an associated vent intrusion at Carskeoch, have a minimum average K-Ar age of 286 ± 7 Ma (De Souza 1982). The volcanics overlie Westphalian D sediments unconformably. The lower lava flows are intercalated with sediments which contain a sparse flora. The plants were originally regarded as Westphalian D in age (Mykura 1965) but were later assigned Stephanian (Wagner 1966) or early Permian (Smith *et al.* 1974) ages. Wagner (1983) now considers these remains to comprise a Rotleigende flora of probable Autunian (early Permian) age. The stratigraphic position of the Mauchline Volcanics is therefore comparatively well-defined and, on the basis that those rocks are probably early Permian rather than latest Carboniferous (Stephanian), an age of 290 ± 5 Ma is proposed for the Carboniferous–Permian system boundary. There are few other radiometric data relevant to the age of this boundary and these are generally of little value because of poor stratigraphic control.

The Bulgonunna Volcanics in the Drummond Basin, east Queensland, Australia (Item 13), rest unconformably on Lower Carboniferous Drummond Group sediments. In the Bowen Basin, these volcanics underlie the Lizzie Creek Volcanics (Item 17), the upper part of which is of Sakmarian age (see below). The Bulgonunna Volcanics have given a Rb-Sr isochron age of 297 ± 12 Ma (Webb & McDougall 1968). Granites which intrude these volcanics, and are closely related to them (Item 14), are overlain by sediments of Permian age in the Bowen Basin; they have given a combined mineral/whole rock Rb-Sr isochron age of 296 ± 3 Ma, and an average K-Ar mineral age of 289 ± 1 Ma (Webb & McDougall 1968).

Data from the Nychum Volcanics of north Queensland, Australia, which were used by Waterhouse (1978) to support an age of 294 Ma or older for the Carboniferous–Permian boundary, must be viewed with caution since the whole-rock Rb-Sr and Th-Pb analyses reported from these rocks by Black *et al.* (1972) gave a range of anomalous results which were attributed to the mixing of two isotopically distinct magma types. A tuff near the base of the volcanics contains a mixed *Cardiopteris–Glossopteris* flora, suggesting a position close to the Carboniferous–Permian boundary. The nearby Featherbed Volcanics are stratigraphically slightly younger than the Nychum Volcanics and have given Rb-Sr biotite and whole-rock isochron ages in the range 270 to 310 Ma (Black & Richards 1972; Black 1980). Associated with the Featherbed Volcanics is a young phase of the Elizaheth Creek Granite, dated by Rb-Sr isochron at 299 ± 6 Ma and by Rb-Sr (biotite) at 300 ± 3 Ma (Black *et al.* 1978). These dates provide a lower geochronological limit of about 300 Ma for the Nychum Volcanics but there is no palaeontological evidence of the stratigraphic position of the Featherbed Volcanics.

In the Oslo Graben, Norway, radiometric data have been obtained from intrusive igneous rocks and from volcanics which rest upon sediments containing a Rotliegende flora and an early Permian fresh-water fauna (Henningsmoen 1978). The most reliable radiometric ages from this area have been obtained from the intrusives. The results of Rb-Sr isochron studies on granite plutons show a reasonable concordance of ages at about 270 Ma (Item 20); Heier & Compston (1969) obtained a Rb-Sr isochron of 270 ± 7 Ma (MSWD = 1.37) for rocks in the main Oslo (larvikite–nordmarkite–ekerite) Series. An isochron for these rocks plus four samples of biotite-granites gave a result of 270 ± 1.7 Ma (MSWD = 1.86) but Heier & Compston (1969) noted that the relationship of the biotite-granites to the main Oslo Series is debatable. A Rb-Sr isochron age of 272 ± 9 Ma was reported by Sundvoll (1975) from larvikite from the main Oslo Series but full analytical details were not published. The Nordagutu granite (Item 19), which intrudes Precambrian basement immediately to the south-west of the Oslo Graben, is thought to be co-genetic with the Oslo Series plutons and has given a Rb-Sr isochron age of 278 ± 7 Ma (MSWD = 1.46) (Jacobsen & Raade 1975). Though apparently somewhat older, this result is within the range of experimental error of the Oslo Series results.

Radiometric dates from the volcanics of the Oslo Graben have been published without full analytical details. A Rb-Sr whole-rock age of about 289 Ma was obtained from lavas and intrusives in the graben (Sundvoll 1976) and Rb-Sr whole-rock ages of 286 ± 8 Ma and 288 ± 5 Ma were obtained from two of the earliest lavas in the Krokskogen area (Sundvoll 1978). It is impossible, in the absence of full analytical details, to substantiate the contention of Sundvoll (1978) that the igneous rocks of the Oslo Graben were produced over the relatively long period of at least 10 Ma; other authorities (Oftedahl 1960, 1967; Heier & Compston 1969) consider the period of igneous activity to have been much shorter. A Rb-Sr age of about 270 Ma is taken here as a minimum age for the early Permian sediments that are cut by the Oslo Series intrusions.

Minor intrusions in north Argyll (Speight & Mitchell 1979), the Ross of Mull (Beckinsale & Obradovich 1973) and southern Norway (Halvorsen 1970; Faerseth *et al.* 1976) have yielded average K-Ar whole-rock ages of 291 ± 5 to 275 ± 8 Ma. These results may reflect a widespread phase of dyke intrusion in areas around the North Sea during the early Permian. Volcanics occur in Rotliegende sequences in the North Sea on the flanks of the Rynkobing–Fyn High (Dixon *et al.* 1981) but no radiometric ages have been published for those rocks.

The remaining radiometric data relevant to the Permian time-scale have been obtained from Australia and, to a lesser extent, South America.

In the north-west shelf region of the Bowen Basin, Queensland, a thick sequence of lavas, the Lizzie Creek Volcanics (Item 17), contains, near its top, a marine fauna of probable Sakmarian (Tastubian) age (Dickins *et al.* 1964). The remainder of this volcanic sequence is unfossiliferous and its older age limit is, therefore, uncertain, but a maximum geochronological age can be interpolated from the stratigraphically lower Bulgonunna Volcanics (Item 13) which are dated at about 296 Ma (see above). The only radiometric data available from the Lizzie Creek Volcanics are two conventional K-Ar determinations on plagioclase from a basalt; these gave an average age of 275 ± 6 ma (Webb & McDougall 1968) but this is of limited significance as the stratigraphic position of the sample within the *c.*3000 m thick volcanic sequence is not known, and the uncertainty on the analyses is said to be large (Webb 1969). Granites associated with the volcanics between Bowen and Collinsville have an average K-Ar age of 270 ± 2 Ma (Webb & McDougall 1968) (Item 18).

Nephrite from two localities in the Great Serpentine Belt of New South Wales (Item 16) has given $^{40}Ar/^{39}Ar$ total fusion ages of 279 ± 12 Ma and 286 ± 12 Ma (Lanphere & Hockley 1976). These radiometric data are reliable but the stratigraphic control is poor since the ultramafic bodies of the Great Serpentine Belt have been tectonically emplaced along, or to the east of, the Peel Fault System. The nephrite in the area sampled, to the south-east of Tamworth, is thought to be a reaction product formed during the emplacement of the serpentinite. A younger age limit is provided by Triassic sediments which overlie the serpentinite in the Lorne Basin, and an older age limit is set in the Manning River district where emplacement is thought to post-date the early Permian sediments of the Dalwood Group (Lower Marine Group).

The stratigraphic age of emplacement of the ultramafics has been a matter of dispute but emplacement is generally regarded as associated with the late Permian Hunter — Bowen orogeny. However, the radiometric ages from nephrite of the Great Serpentine Belt only provide a minimum age of *c*.282 Ma for a post-early Permian (post-Dalwood Group) event.

A number of volcanic horizons in middle to late Permian sequences in the Sydney Basin, Australia (Item 22), have been dated by conventional K-Ar methods (Evernden & Richards 1962; Facer & Carr 1979). Permian and Triassic marine and fresh-water sediments rest unconformably on Lower and Middle Palaeozoic rocks in the south-western part of the Basin. The Shoalhaven Group, at the base of the Permo-Triassic, comprises clastic marine sediments which contain Kazanian brachiopods (Waterhouse 1976) and is overlain by the Illawara Coal Measures, of Tatarian age; these are succeeded by fluviatile and lacustrine sediments of Triassic age. The Gerrigong Volcanics, comprising four latite flows, are intercalated with sediments in the upper part of the Shoalhaven Group. The Berkeley latite, data from which was used in the construction of the Geological Society Phanerozoic Timescale of 1964, and the Minnamurra latite are intercalated in the Illawara Coal Measures succession. The radiometric data from the latites and a number of associated intrusions have been summarized by Facer & Carr (1979) but must be used with caution because no experimental errors were quoted for the older data of Evernden & Richards (1962). However, using the most reliable ages from the Berkeley latite, and from the Dapto-Saddleback and Bombo latites in the Shoalhaven Group, Facer & Carr (1979) proposed an age of about 253 ± 5 Ma for these lavas. They point out, however, that all the K-Ar data indicate minimum ages only, as alteration is widespread in the igneous rocks. For this reason an age of 255 ± 5 Ma is proposed here for the Kazanian–Tatarian boundary, rather than the minimum age of 253 Ma suggested by Facer & Carr (1979).

A tuff from the top of the Gyranda Formation in the Bowen Basin, Queensland, has given a duplicate K-Ar age from biotite of 244 Ma (Webb & McDougall 1967; no errors quoted). The Gyranda Formation overlies a marine sequence containing faunas of Artinskian to Kazanian age and is overlain by the Baralaba Coal Measures which contain Chidruan (Tatarian) microfossils (Webb 1981). The age from the tuff appears low, especially in comparison with other, better substantiated ages; as a single date from one sample it can only be regarded as indicating a minimum age for the late Permian. A similar age was reported for an intrusion cutting the Illawara Coal Measures, of Tatarian age, in the Sydney Basin (Carr & Facer 1980). This intrusion had a conventional K-Ar age of 243 ± 10 Ma.

A large number of K-Ar determinations have been published from the volcanics of Mendoza Province, Argentina (Creer *et al.* 1971; Valencio & Mitchell 1972; Valencio *et al.* 1975; Toubes & Spikermann 1976, 1979; Linares *et al.* 1978; Rocha-Campos *et al.* 1971). Volcanics in the Carrizalito Group, which is regarded as middle or late Permian, are dated by K-Ar isochron at 266 ± 4 Ma (average conventional K-Ar age of 265 ± 3 Ma) and 249 ± 6 Ma (average conventional K-Ar age of 251 ± 4 Ma) (Linares *et al.* 1978). These rocks are overlain by the early Anisian Puesto Viejo Formation, and rest unconformably on volcanics of the Cochico Group, which are regarded as early Permian and are dated by K-Ar isochron at 273 ± 8 Ma (average conventional K-Ar age of 273 ± 5 Ma) (Linares *et al.* 1978). However, stratigraphic control of the dated volcanics is poor since they occur in unfossiliferous continental deposits which can only be assigned a late Palaeozoic to early Mesozoic age (Valencio *et al.* 1977) and these ages are not, therefore, suitable for use in the construction of a radiometric time-scale.

In Peru, the 4000 m of unfossiliferous continental sediments and volcanics of the Mitu Group, of inferred Kungurian age (Rocha-Campos 1973), rest unconformably on Copocabana Group limestones, which contain an early Permian fusiline fauna, and are overlain by Triassic (early Ladinian or Norian) limestones (Capdevila *et al.* 1977). In the eastern Cordillera of Peru the Quillabamba granite (Item 21) intrudes red molasse sediments of the Mitu Group and has given a U-Pb age of 257 ± 3 Ma. The Coasa granite (Item 26), which cuts early Permian limestones, has given a U-Pb age of 238 ± 11 Ma (Lancelot *et al.* 1978). The Villa Azul granite, cutting Mitu Group sediments some 200 km to the north-west of the Quillabamba granite, has been dated at 256 Ma (K-Ar) in a reconnaissance study by Stewart *et al.* (1974), but full analytical details and errors were not published. A single conventional K-Ar age of 265 ± 25 Ma has been reported by Rocha-Campos & Amaral (1971) from a coeval stock associated with the Mitu Group volcanics. The San Ramon granite of central Peru, also thought to be coeval with the Mitu Group volcanics, has given a Rb-Sr isochron age of 246 ± 10 Ma. (Capdevila *et al.* 1977).

Radiometric evidence for the age of upper Permian sequences is inconclusive; in Australia, two phases of granite emplacement occurred in southern Queensland during the latter part of the period. These have their maximum geochronological expression at about 250 Ma (granites near Marlborough and Cookstown; Items 23, 24) and between 240 and 235 Ma (granites between Marlborough and New England; Item 25) (Webb & McDougall 1968). In the Marlborough district the 250 Ma intrusions are post-early Permian and border on the Gogango Overfolded Zone, where sediments as young as late Permian have been folded; the Marlborough gabbros and diorites show no evidence of being affected by this folding and, therefore, are probably late Permian or younger. The 240 to 235 Ma episode of granitic intrusion was the most widespread in eastern Queensland, occurring over a wide area from Marlborough southwards towards Tamworth in the region of late Permian uplift. These younger granites intrude fossiliferous early Permian sediments and are overlain by mid-Triassic to Jurassic sediments; originally all were thought to be of Permian age but some, at least, may be of Early Triassic age (Webb, A. W., 1969; Olgers *et al.*, 1974; Webb, J. A., 1981).

The Triassic

The age data used here for the Triassic period are from sources in Australasia, the Americas, Africa and Europe and are related to the succession of stages defined in Tethyan and North American Triassic sequences on the basis of ammonoid faunas (Silberling & Tozer 1968; Tozer 1974, 1978; Wiedmann *et al.* 1979).

The Permian–Triassic boundary and the Early Triassic

No radiometric data are available from stratigraphic levels close to the biostratigraphically defined Permian–Triassic system boundary. The significance of dates from granites in Queensland (Items 23–25, above) is diminished by lack of stratigraphic control; the late Permian Gerrigong Volcanics have good stratigraphic control but the radiometric data from that source have internal inconsistencies. Therefore, in the absence of reliable radiometric data from Permian sequences, any estimate of the age of the Permian–Triassic system boundary must be based on extrapolation from available dated horizons in Triassic successions.

The oldest dates reported from rocks of undoubted Triassic age are those from volcanics in the Toogoolawah Group in Australia and the Puesto Viejo Formation of Argentina. The Neara Volcanics (Item 27) of the Esk Rift Valley in south-east Queensland are a series of andesites and trachytes within the 5000 m thick Early to Middle Triassic Toogoolawah Group. Two conventional K-Ar ages of 242 ± 5 Ma (whole rock) and 239 ± 5 Ma (hornblende) have been reported by Irwin (1976) from two different horizons within these volcanics but analytical details and sample descriptions were not published. On palynological evidence the Toogoolawah Group is thought to be Anisian to Ladinian in age (De Jersey 1972). The radiometric ages obtained from the Neara Volcanics only allow the assignment of an age of about 240 Ma to Anisian–Ladinian successions in Queensland.

The Kin Kin Beds of the Gympie Basin, Queensland, which are intruded by the Goomboorian diorite (Item 29) contain Early Triassic (Smithian) ammonites. The diorite has an average K-Ar mineral age of 236 ± 7 Ma (Green & Webb 1974; Murphy *et al.* 1976) which is, therefore, a minimum age for the Smithian Stage.

In Argentina, volcanics within the Puesto Viejo Formation (Item 28) have been assigned a mean K-Ar age of 237 ± 4 Ma on the basis of five determinations from different volcanic horizons sampled at or near the base of the formation at two different localities (Valencio *et al.* 1975). The formation comprises 300 m of continental sediments with intercalated acidic and basic extrusive rocks. Fossil reptile assemblages in the upper part of the sequence are indicative of an Early to Middle Triassic, probably late Scythian, age (Anderson & Cruickshank 1978). The Puesto Viejo Formation rests unconformably upon Permian volcanics of the Carrizalito Group, dated at about 265 ± 5 Ma (Linares *et al.* 1978). The petrography of the dated samples is not described and the relative freshness of the lavas cannot, therefore, be ascertained. When compared with the average age of about 240 Ma from the Anisian/Ladinian Neara Volcanics, the Puesto Viejo results appear rather low. It is possible that the ages at the older end of the range (about 241 ± 10 Ma) are closer to the true age of the rocks. The mean age of 237 ± 4 Ma is, therefore, probably best regarded as a minimum for the Scythian.

The Permian–Triassic system boundary is considered to be at least 5 to 10 Ma older than the dates obtained from Early and Middle Triassic successions in Australia and Argentina but, with so little information available from late Permian successions, it is difficult to propose an objective figure. The system boundary may well be older than 250 Ma and, on the basis of the Triassic data, an age of 250 ± 5 Ma appears reasonable. This is considerably older than the 230 to 235 Ma value assigned in many previous time-scales but accords with recent estimates of 245 ± 5 Ma by J. A. Webb (1981) and Odin & Kennedy (1982), 247 Ma by Armstrong (1978) and 250 Ma by Hellmann & Lippolt (1981).

Middle Triassic

A detailed geochronological study was made of a number of feldspar populations preserved in volcanic bentonite horizons in the Grenzbitumenzone (Item 30) of the Monte San Giorgio area, Ticino, Switzerland (Hellmann & Lippolt 1981). The dated feldspars included both primary volcanic sanidines and secondary authigenic K-feldspars. The stratigraphic position of the bentonites is precisely known; they are interbedded with sediments containing mollusc faunas indicative of a position close to the Anisian–Ladinian boundary. The high sanidine phenocrysts, of primary volcanic origin, had an average K-Ar age of 232 ± 9 Ma and phenocrysts composed of a high sanidine core with a secondary overgrowth of K-feldspar had an average K-Ar age of 235 ± 6 Ma; these ages were supported by $^{40}Ar/^{39}Ar$ plateau ages of 232 ± 8 Ma and 237 ± 7 Ma from the primary high sanidines and sanidine plus overgrowth, respectively. Argon isochron plots of the primary high sanidines had an intercept of 290 ± 90 on the $^{40}Ar/^{36}Ar$ axis, suggesting that the argon isotopes have been virtually undisturbed; the sanidines plus overgrowth had an intercept of 190 ± 192, the high error in this case being caused by the highly radiogenic nature of some of the data points used on the isochron plot. Wholly authigenic feldspars yielded an average K-Ar age of 226 ± 8 Ma, with a $^{40}Ar/^{39}Ar$ plateau age of 225 ± 9 Ma, which was interpreted as representing the age of alteration of the original volcanic tuffs into bentonites (Hellmann & Lippolt 1981).

The dates from the primary volcanic sanidines from the Grenzbitumenzone are important for the establishment of a time-scale for the Triassic period. From a detailed study of twenty-six different feldspar samples, representing three separate bentonite horizons sampled at seven sites, Hellmann & Lippolt (1981) concluded that there were no significant differences in the age data from different grain size fractions or from different levels in the sequence; the lowest and highest bentonites sampled were, however, only two metres apart. The 12 m thick Grenzbitumenzone was laid down during an interval represented by only one or perhaps two ammonoid zones (Rieber 1969, 1973); thus, the stratigraphical error is ± one zone and is well within the range of experimental error of the radiometric data. These results, which combine analytical precision and internal consistency with a well-defined stratigraphical position, provide the most precise data available for any level within the Triassic. On the basis of these dates, an age of 235 ± 5 Ma is proposed for the Anisian–Ladinian boundary.

High sanidine from the Val Serrata Tuff (Item 33), which occurs some 110 m above the Grenzbitumenzone, gave a mean K-Ar age of 225 ± 4 Ma (Hellmann & Lippolt 1981). This provides a minimum age for the early Ladinian but appears rather low when compared with other data from higher levels in the Triassic.

In the Maryborough Basin of Queensland, Webb & McDougall (1967) established a minimum age of 225 Ma for the Middle Triassic. A number of small granite plutons gave a Rb-Sr isochron age of 226 ± 16 Ma, and an average K-Ar mineral age of 222 ± 2.5 Ma (Item 34). On the assumption that these plutons are all of the same age, the stratigraphic limits on these radiometric ages are post-Brooweena Formation (Early to Middle Triassic) and pre-Landsborough Sandstone (early Jurassic). To the west of the Maryborough Basin, Webb & McDougall (1968) described a number of intrusions that cut the Neara Volcanics, dated at about 240 Ma (Item 27), and the Esk Formation (Anisian to Ladinian in age) in the Esk Rift. The Somerset Dam Gabbro, which cuts the Neara Volcanics, had average K-Ar ages of 219 Ma (hornblende) and 211 Ma (plagioclase) (no errors quoted), and the Brisbane Valley Porphyrites, a dyke swarm intruding the Esk Formation, had an average K-Ar age of 223 Ma (hornblende; no errors quoted). These ages from intrusions that are post-Anisian–Ladinian indicate that the Middle Triassic is older than 220 to 225 Ma.

Late Triassic

The Predazzo and Monzoni igneous complexes in northern Italy intrude Ladinian and early Carnian limestones. Volcanics occur within this sedimentary sequence and, in the Predazzo area, are post-dated by the granite, the emplacement of which is thought to have followed shortly after the extrusive activity; a tentative upper limit of late Carnian is placed upon the intrusive complexes. A range of dates has been reported from these complexes by Borsi & Ferrara (1967), Borsi *et al.* (1968) and Ferrara & Innocenti (1974) but several discrepancies are apparent between the values obtained from Rb-Sr and K-Ar analyses.

Most of the Rb-Sr data concurred at about 238 Ma but the K-Ar data suggested an age of between 221 and 233 Ma, with an average of about 227 Ma. Investigations undertaken to resolve some of these anomalies suggest that the discrepancies lie in the Rb-Sr data (Ferrara, pers. comm. 1982) and that more reliance should be placed upon the K-Ar data. The K-Ar ages from phlogopite in veins in the aureole of the Monzoni granite (Item 31) probably provide the best estimate of the age of this complex at 231 ± 8 Ma; those from the Predazzo Complex show a wide range of values and are less satisfactory.

The Sugars Basalt (Item 32) occurs at the base of the Ipswich Coal Measures in the Ipswich Basin in south-east Queensland; its basal contact is not seen, but it is thought to overlie Palaeozoic basement. The Ipswich Coal Measures have been assigned a Carnian age on palynological evidence (De Jersey 1971). The basalt has yielded an average K-Ar age of 230 ± 7 Ma (Webb & McNaughton 1978) but, in the absence of a good upper age limit, this date can only be used as a general guide to the age of the overlying Carnian sediments and, since the samples are described as relatively unaltered (i.e. not completely fresh), is regarded as a minimum age.

A provisional K-Ar mean age of 229 ± 5 Ma has been reported by Gonzalez & Toselli (1975) from volcanics overlying continental sediments which contain Ladinian to Carnian fossil reptile assemblages in the Ischigualasto-Ischichuca Basin of north-west Argentina. Full analytical details were not published with this result.

The Indonesian tin granites (Item 35) have yielded Rb-Sr and K-Ar ages with a good correlation at about 216 Ma (Edwards & McLaughlin 1965; Priem *et al.* 1975; Priem & Bon 1982). K-Ar ages from biotite and hornblende from the plutons of Belitung, Bangka and the Tuju Islands range from 215 ± 10 Ma to 229 ± 20 Ma, with an average age of 219 ± 4 Ma; Rb-Sr mineral ages of 208 ± 6 Ma (biotite) and 212 ± 7 Ma (feldspar) have been reported from the Tanjung Pandang massif on Belitung. A Rb-Sr whole-rock plus biotite isochron of 213 ± 5 Ma has been obtained for the Belitung, Bangka and Tuju Islands plutons, but this has a high MSWD of 7.9, indicating that geological heterogeneities are present in the granites dated. Priem & Bon (1982) concluded that the best estimate of the age is provided by the average Rb-Sr and K-Ar values obtained from biotite/whole-rock pairs and biotite/hornblende pairs; this gives an average age of 216 ± 2 Ma. On Bangka Island the granite intrudes a folded sequence of flysch deposits containing fossils of late Carboniferous to Triassic age; the youngest fossils occur in a limestone lens and are of Norian age. The folded sequence is overlain discordantly by molasse sediments of the Bintan Formation, which contain plant fragments thought to be of Rhaetian (Jongmans 1951) or Neocomian (Kon'no 1972) age. The average age of 216 Ma, therefore, provides a good minimum age for the Norian and possibly indicates a maximum age for the Rhaetian. This is supported by results from the Kulim granite pluton (Item 36) in western peninsula Malaysia (Bignell & Snelling 1977). This cuts the Semanggol Formation, which contains faunas of late Ladinian to early Norian age (Burton 1973), and has yielded a Rb-Sr isochron age of 207 ± 17 Ma (MSWD = 0.5) and K-Ar ages of 205 ± 6 and 200 ± 8 Ma (biotite).

In British Columbia, large acidic plutons intrude fossiliferous Late Triassic and early Jurassic sediments and intercalated volcanics in the Intermontane Belt. The plutons have been the subject of extensive geochronological investigation by workers at the University of British Columbia and by the Geological Survey of Canada in Ottowa. The large number of radiometric ages available from British Columbia has been summarized and assessed by Armstrong (1982); only some of the major items will be considered here. The Hotailuh batholith of the Cassiar Mountains intrudes volcanics intercalated with sediments containing a Carnian fauna. Average K-Ar mineral ages fall into two groups, one at about 220 Ma and one at about 166 Ma (Wanless *et al.* 1972); this grouping suggests that the batholith is polyphase and, since no accurate younger age limit can be placed on the pluton, the older ages can only be used to provide a minimum age of about 220 Ma for the Carnian sediments cut by the granite.

In the southern part of British Columbia, other batholiths cut Nicola Group sediments and intercalated volcanics and have given K-Ar ages ranging from 234 ± 9 Ma and 215 ± 4 Ma (hornblende) from the Coldwater Stock to 202 ± 6 Ma (average of four biotite determinations) for the Iron Mask batholith (Preto *et al.* 1979), and 195 ± 6 Ma (mean of three hornblende determinations) and 198 ± 10 Ma (biotite) for the Thuya batholith (Campbell & Tipper 1971; Preto *et al.* 1979). Some of the older dates may represent early magmatism associated with the oldest (Carnian) volcanic horizons in the Nicola Group.

The Guichon Creek batholith (Item 37) intrudes the Nicola Group and has been used as a marker on many previous Triassic time-scales. Numerous K-Ar age determinations from this batholith have yielded consistent results with an average value of 204 ± 5 Ma (White *et al.* 1967; Northcote 1969; Jones *et al.* 1973) which is supported by a Rb-Sr whole-rock isochron of 205 ± 20 Ma (Preto *et al.* 1979). This isochron was, however, based partly upon data from mineralized samples of the granite and the age of the intrusion may, therefore, be slightly older. The batholith is post-Norian and is overlain by fossiliferous marine sediments of Hettangian–Sinemurian age (Frebold & Tipper 1969). The Guichon Creek data, therefore, provide a good minimum age for post-Norian–pre-Jurassic horizons.

Volcanics regarded as associated with the British Columbia plutons have better stratigraphic control but have yielded few reliable radiometric ages.

Attempts to date volcanics in the Nicola Group, which is of Carnian to late Norian age, have been unsuccessful because of the ubiquitous low-grade epidote-albite-chlorite metamorphism affecting those rocks (Preto *et al.* 1979).

In the northern Bowser Basin of central British Columbia, sediments intercalated in a thick sequence of andesitic volcanics contain early Toarcian ammonites. Rb-Sr whole-rock isochron ages of 191 ± 18 Ma and 189 ± 26 Ma have been reported from these volcanics by Gabrielse *et al.* (1980), (Item 43). The Toodoggone Volcanics (Item 44), of presumed early Jurassic age, have given a Rb-Sr isochron of 185 ± 10 Ma and the Hazleton Group volcanics (Item 45), tentatively correlated with the Sinemurian to early Pliensbachian Telkwa Formation, have given a Rb-Sr whole-rock isochron of 185 ± 6 Ma. Several plutons regarded as associated or coeval with these volcanics have been dated; the Tachek batholith, which cuts Hazleton Group volcanics, has given K-Ar hornblende ages of 199 ± 8 Ma and 209 ± 9 Ma and the Howson batholith at Talkwa Pass has given K-Ar hornblende ages of 193 ± 8 Ma and 210 ± 9 Ma (Wanless *et al.* 1974). K-Ar mineral ages of 190 ± 8 Ma, 204 ± 9 Ma (hornblende) and 193 ± 7 Ma (biotite) have been reported by Wanless *et al.* (1978) from the Black Lake batholith which is thought to be coeval with the Telkwa and Hazleton volcanics.

Although the stratigraphic constraints on many of the plutons in British Columbia are rather wide, the large number of compatible radiometric ages available from this area, spanning Late Triassic to early Jurassic time, provide a good basis for estimates of the age of the Triassic–Jurassic system boundary; an age of 205 ± 5 Ma is proposed for this boundary.

Intrusive and extrusive igneous rocks associated with the early stages of continental rifting prior to the break-up and dispersal of Pangea and Gondwanaland provide radiometric ages for Late Triassic–early Jurassic successions elsewhere. May (1971) described the widespread intrusion of dykes at this time in areas bordering the North and South Atlantic; these were the precursors of widespread magmatism which gave rise to the flood basalts of the Karroo in southern Africa, the Newark Volcanics on the eastern seaboard of North America, and major volcanism in many other areas bordering the South Atlantic and Indian Oceans. From the limited amount of radiometric data available in 1971, May concluded that all the circum-Atlantic dykes were of similar age and, on a pre-drift configuration of the Atlantic, were arranged in a radial pattern related to the stress field present in the area of initial rifting prior to continental separation. Radiometric and palaeomagnetic results obtained over the last decade have shown that though the majority of these dykes were probably intruded at about 195 Ma, there were early manifestations of the igneous activity at about 205 Ma, and also a later period of dyke intrusion at about 180 Ma (Sutter & Smith 1979).

Evidence for an early phase of magmatic activity comes from southern Africa, where isolated alkaline intrusive complexes and minor intrusions are dated at about 204 ± 5 Ma (Nicholaysen *et al.* 1962; Allsopp *et al.*, in press). Fitch & Miller (in press) report a $^{40}Ar/^{39}Ar$ plateau age of 204 ± 2 Ma from a sill cutting Ecca Group sediments in the Kruger National Park, South Africa. The onset of the major Karroo volcanism is dated at about 193 ± 5 Ma by Fitch & Miller (in press). A K-Ar isochron age of 193 ± 3 Ma is reported for the Pronksberg andesite (Item 42), the oldest member of the volcanic succession in the central Karroo area; this is supported by similar ages from K-Ar isochrons and conventional K-Ar determinations on dykes intruding the Lesotho basalts, and on horizons at the base of the Karroo volcanics in the Barkley East district, South Africa (Fitch & Miller, in press).

At a comparable time, dyke swarms were being emplaced along the west coast of Africa from Namibia to Sierra Leone and Liberia, and major basic intrusions were emplaced from Cape Province to Lesotho and Nuanetsi (Manton 1968). The Freetown igneous complex of Sierra Leone (Item 41) has been dated by Rb-Sr isochron at 193 ± 3 Ma (Beckinsale *et al.* 1977) and K-Ar ages ranging from 198 ± 8 Ma to 169 ± 10 Ma were reported by Briden *et al.* (1971). The coastal dyke swarm of Liberia (Item 40) has been dated by conventional K-Ar and $^{40}Ar/^{39}Ar$ techniques by Dalrymple *et al.* (1975); plagioclase and whole-rock analyses from a sill intruding the early to mid-Palaeozoic Paynesville Sandstone gave well-defined $^{40}Ar/^{39}Ar$ plateau ages of 191 ± 6 Ma and 188 ± 8 Ma, respectively. Argon isochron plots gave respective ages of 196 ± 8 Ma and 190 ± 5 Ma, and had intercepts close to 296 on the $^{40}Ar/^{36}Ar$ axis, showing that there had been little subsequent disturbance of the argon isotopes. Dalrymple *et al.* (1975) concluded that the average value of the isochron ages, 193 Ma, gave the most likely age of crystallization. A number of highly discrepant dates have been reported from both the Freetown igneous complex and the Liberian dyke swarm; Dalrymple *et al.* (1975) resolved these difficulties and demonstrated that the excessively high ages obtained previously from some of the dykes were caused by excess argon in those intrusions cutting crystalline pre-Cambrian basement.

On the eastern seaboard of North America the dykes and sills of the Newark Basin have been dated by Sutter & Smith (1979); intrusions in Connecticut and Newfoundland have been dated by Hodych & Hayatsu (1980), and in Nova Scotia by Hayatsu (1979), and the Palisades Sill (Item 38) has been dated by Dallmeyer (1975). The Palisades Sill in New Jersey has, for many years, been a key reference point for the Triassic–Jurassic boundary and has been the subject of numerous dating studies (Holmes 1937; Erickson & Kulp 1961; Armstrong & Besancon 1970) which gave ages ranging from 145 to 206 Ma. Recent $^{40}Ar/^{39}Ar$ whole-rock analyses by Dallmeyer (1975) have given a plateau age of 196 ± 9 Ma from the upper chill zone of the sill; this result shows good agreement with earlier conventional K-Ar analyses on biotite, which gave an average age of 195 Ma (Armstrong & Besancon 1970). Although traditionally used to date the end of the Triassic period, the sill is now thought to be of early Jurassic age; it is correlated with the Newark Group volcanics of the Hartford Basin which, on palynological evidence, are regarded as Rhaetian to Hettangian in age (Cornet & Traverse 1975). As such, its age correlates well with those of other early Jurassic sills and dykes discussed below.

The Mount Carmel Sill (Item 39), an intrusion in the Hartford Basin, has given a $^{40}Ar/^{39}Ar$ plateau age of 195 ± 4.2 Ma and an argon isochron age of 195 ± 11 Ma, with initial $^{40}Ar/^{36}Ar$ close to 296 (Sutter & Smith 1979). These ages are supported by conventional K-Ar determinations which gave an average age of 196 Ma (Armstrong & Besancon 1970). Other intrusions belonging to the Newark suite in Connecticut have been dated by K-Ar isochron (Hodych & Hayatsu 1980), but the isochron in this case shows a low intercept on the $^{40}Ar/^{36}Ar$ axis, suggesting that there has been subsequent disturbance of the argon isotopes and probable argon loss. Similarly, K-Ar isochron ages reported from the North Mountain basalt, Nova Scotia (Hayatsu 1979) and from the trans-Avalon aeromagnetic lineament (Hodych & Hayatsu 1980) have low intercepts on the $^{40}Ar/^{36}Ar$ axis suggesting partial argon loss.

Scattered representatives of the circum-Atlantic dykes occur on the eastern margin of the North Atlantic. Schermerhorn *et al.* (1978) reported average K-Ar ages of 188 ± 5 Ma and 172 ± 5 Ma from the Messejana Dyke of southern Portugal and Spain; this large dolerite dyke, of post-Hercynian, pre-Miocene age, was emplaced along a fault system which trends north-east from the coast of south-west Portugal for some 530 km until masked by Miocene sediments. The Foum Zguid Dyke in south Morocco, which trends north-east and is some 200 km in length, has given K-Ar ages of 185 ± 3 Ma and 191 ± 4 Ma (Hailwood & Mitchell 1971).

Conclusions

It is clear from the foregoing discussion that the data available for each of the periods under consideration show immense variation in both quantity and quality. Much of the useful information for levels within the Carboniferous comes from Britain, principally from the volcanic rocks in the Midland Valley of Scotland. The Permian is very poorly represented, with very few reliable dates available from areas where satisfactory stratigraphical control exists. The Triassic emerges as perhaps the best-documented of the periods discussed here, with a relatively large number of reliable dates published; a reasonable number of these dates also have reliable stratigraphic control.

The value of detailed, well-integrated geochronological studies such as that of Hellmann & Lippolt (1981) on Triassic bentonites cannot be over-emphasized. Some appreciation of the range in quality of the data reviewed for this account can be gained by comparing this work with, for example, that of Calver & Castleden (1981) on Middle Triassic basalts from Tasmania or that of Retallack *et al.* (1977) in which a single conventional K-Ar age of 216 ± 5 Ma from plagioclase from the Dalmelly basalt, of Anisian–Ladinian age, was stated to be 'one of the more reliable points for the (Triassic) scale'. Reliance on single, unconfirmed ages for the definition of points on the geological time-scale has been a major fault of many previous time-scales. Any attempt to achieve a 'best fit' solution by including data of such widely varying quality can only result in a general lowering of the ages assigned to all the major boundaries. In order to avoid this problem Armstrong (1978) proposed that the scale should be constructed at the maximum values consistent with minimum violation of other available constraints; this has been the intention in this review.

Any data that are to be used in a time-scale must meet a number of requirements; they must be reproducible, by the same and by different methods, and they must be reproducible from different specimens, rock types and minerals from the same locality and from other localities at the same horizon.

On a more optimistic note; the work of Hellmann & Lippolt (1981) demonstrates that the application of modern analytical techniques to other, well-defined igneous horizons within sedimentary sequences of known biostratigraphic age could yield equally impressive results in terms of the detailed correlation of the geological time-scale. Bentonites, in particular, would warrant further investigation in the Carboniferous, Permian and Triassic, since the frequent presence of high temperature, primary volcanic phenocrysts of minerals such as sanidine, biotite, apatite and zircon means that the problems of alteration and loss of radiogenic daughter products, which are so common in otherwise suitable lava horizons, may be avoided. Bentonites may be particularly suited to the use of a variety of different dating techniques depending on the volcanic source; as well as conventional K-Ar, Rb-Sr and $^{40}Ar/^{39}Ar$ methods, recent work by Ross *et al.* (1982) has shown that fission track dating may contribute useful data for bentonites in the older part of the Phanerozoic time-scale.

A detailed, fully integrated geological time-scale for the Carboniferous, Permian and Triassic will only be possible when many more precise, accurate, well-correlated age determinations are available from those areas where the stratigraphic record is well-known. The time-scale proposed here for the Carboniferous, Permian and Triassic is based on a relatively small number of ages that meet a number of basic criteria; the advent of new age data may well result in revision of the ages proposed in this review.

Note added in page proof

Since the completion of the literature survey on which this account is based a number of relevant papers have appeared but the results reported in these do not satisfy our criteria for use in time-scale construction; for example, the stratigraphic control on the results reported by Lippolt *et al.* (1983) from Permian lavas in south-western Germany, and by Hess *et al.* (1983) from tuffs in Carboniferous rocks in the same region is inadequate.

Harland *et al.* (1982) and Carr *et al.* (1984) have reviewed aspects of the Phanerozoic timescale, and Foster (1983) has collated lithostratigraphic, biostratigraphic and radiometric data from the Permian and Early Triassic sequences of Queensland, Australia.

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Comments on the geochronology of the Carboniferous to Triassic times

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Thanks to the relatively numerous and more or less reliable radiometric dates obtained from the Carboniferous to Triassic sequences it is possible to choose the best calibration points before trying to draw a picture of the corresponding numerical time-scale. In this context, the analysis of Forster & Warrington in the introduction to their chapter is a most welcome reminder of the criteria for data selection, sometimes neglected or forgotten by other authors.

Taking this into account, one may be sure that the portion of the time-scale proposed is probably one of the most objectively founded in this book for: (1) The authors are fully aware of the various uncertainties arising in the exercise of assessing a calibration point. (2) They actually use their own recommendations, when possible, and try to discuss each of the different areas of uncertainty (i.e.) the stratigraphical and the geochemical as well as the analytical ones in order to weight the reliability of a given date. (3) They have found in the literature a large number of data from which they selected 45. These are supported by other results which, although only briefly mentioned and less significant, are useful to reinforce their arguments.

The result of this considerable work is a time-scale which certainly is a very good reference. The minor remarks proposed here must be considered therefore as a modest complement of information to emphasize some of the limited differences between this work and the synthesis edited by the author in 1982 (Odin 1982).

The selection of items

The author's first general remarks concern the choice of the items and other results quoted by Forster & Warrington, relative to some others originally found in the (French) literature but also quoted and discussed in detail (in English) in Odin (1982). Indeed, the choice of the 'best' calibration points remains partly a question of subjectivity. However, it is somewhat disappointing to observe that very few references for calibrating that portion of the time-scale come from French workers, in spite of the actual abundance of published radiometric studies related to the subject. The author would like to submit five examples which could have been considered as useful as many of those quoted by Forster & Warrington for different parts of the time-scale.

The basal Carboniferous times

Peucat & Charlot (1982) have proposed a preliminary series of results obtained from the Huelgoat granite (Armoricain Massif) which leads to an apparent age of 336 ± 13 Ma for post-Famennian–pre-Strunian times, actually more or less *at* the Devonian–Carboniferous boundary (see details and references in NDS 229). This seems to imply that the concerned boundary must be younger than 349 Ma.

Charlot & Vidal (1982, NDS 133) have presented an abstract of their available data on volcanics overlying rocks affected by late Devonian tectonism and covered themselves with fossiliferous beds of early Tournaisian and early Visean age, respectively in the Laval Basin and Chateaulin Basin. The obtained ages of 342 ± 13 Ma and 331 ± 6 Ma seem to be of some use for the calibration of the numerical age of the Tournaisian stage boundaries. The two references quoted above suggest an age younger than 350 Ma for the Devonian–Carboniferous boundary. But the more recent researches undertaken in 1982–1983 by Montigny in the Vosges (NE France) lead this author to suggest an age certainly older than 340 Ma and possibly older than 350 Ma for the boundary early Visean–middle Viséan (see abstracted results in the Bulletin of information IGCP Project 196 n°2).

The Carboniferous–Permian boundary

De Souza (1982, NDS 231) has gathered a series of radiometric results obtained from Stepbanian (Carboniferous) to Autunian (Permian) volcanics from Southern France. The apparent age of 283 ± 5 to 301 ± 5 Ma might have been quoted to further document or comment upon the numerical age of about 290 Ma for the Carboniferous–Permian boundary, in spite of the small remaining uncertainty on the stratigraphy of some of these French samples.

The base of the Triassic times

Montigny *et al.* (1982, NDS 158) have abstracted their radiometric results on Werfenian (= Scythian = early Triassic) gabbroic plutons from Yugoslavia. The gabbro intrusion is not precisely located *inside* the Werfenian stage; however, the preferred analytical result of 248 ± 7 Ma is probably an interesting reference to estimate the numerical age of the Permian–Triassic boundary in the absence of results obtained from rocks stratigraphically better located than this one. The analytical data are not very numerous for the gabbro, but it is clearly located nearer the Permian–Triassic boundary than the Neara volcanics (item 27 of Forster & Warrington and NDS 194 for other details), than the Kinkin beds (item 29 and NDS 195) in Australia, or than the Puesto Viejo formation of Argentina (item 28). Concerning this last formation, the author must recall that, depending on the palaeontologists, and even for a given author, the stratigraphical attribution varies from the top Scythian to the top Ladinian. This is clearly discussed in the abstract NDS 186 and the author recommends that this sample should not be weighted heavily for time-scale calibration (much less than NDS 158 quoted above).

The base of the Jurassic times

The most interesting analytical results available to estimate the age of the Triassic–Jurassic boundary are certainly those obtained from the Newark group of the Hartford basin in the eastern USA, (item 39 of Forster & Warrington and also Webb 1982, abstracts NDS 202-203). However, there are few stratigraphically well located Hettangian rock dates in the literature. Some are presented in the abstract NDS 213 by Baubron & Odin (1982); the available K-Ar data lead to a tentative age of 194 ± 7 Ma for the Hettangian from the French Massif Central. This age needs confirmation due to a possible rejuvenation observed in some of the dated samples, but it is possibly of interest to support other results.

Although not one of the above quoted dates is perfect, some of them are satisfactory and the author suggests that they are of nearly similar standard compared with those proposed by Forster & Warrington. Most of them support the ages proposed by these authors.

The plus or minus

There is another question which appears to he of some importance in the utilization of the portion of the time-scale in the form proposed by Forster & Warrington in this volume. It concerns the estimated uncertainty (±) given for their boundary age recommendations. The author has already tried to point out this unfortunate choice following the oral presentation of their paper in 1982. It is regretted that the authors have kept their systematic ± 5 Ma for all the discussed boundaries. The actual knowledge of the numerical age of the different boundaries is absolutely not equivalent. This is not very important for the small number of well documented boundary ages for which Létolle & Odin (1982) have proposed ± 4 Ma; for example the Anisian–Ladinian boundary is very well documented and may be confidently located between 229 and 237 Ma. The Triassic–Jurassic is well known too, but the Carnian–Norian boundary is much less accurately known. An interval of time of 16 Ma was accepted in 1982 as most

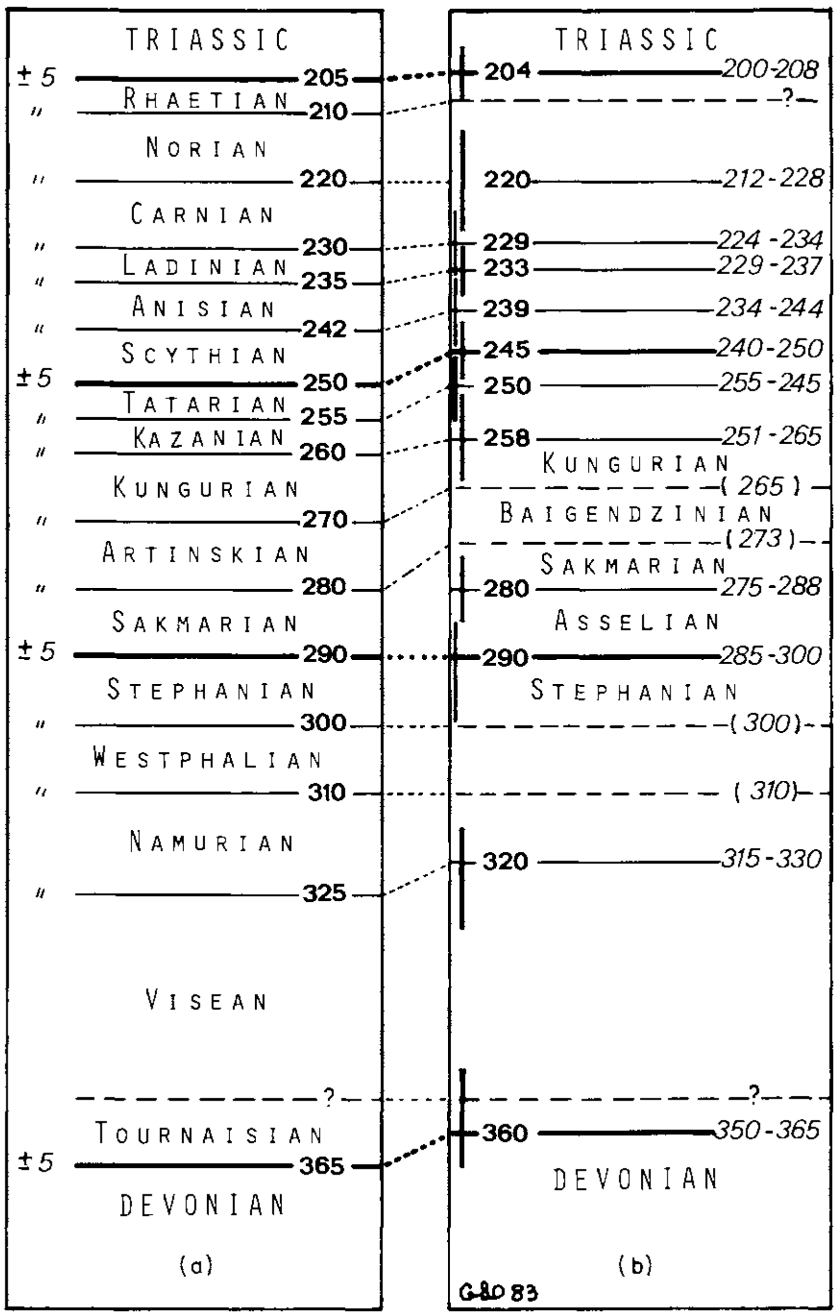

FIG. 1. Comparison of the numerical ages of the Carboniferous, Permian and Triassic stage boundaries according to the (a) scale of Forster & Warrington 1985 and (b) the combined scales of De Souza (1982), Odin & Gale (1982), Odin & Létolle (1982). Note the nearly complete agreement; the small differences mostly lie in the estimates of the uncertainties linked with each recommended number.

probably including that boundary (between 228 and 212 Ma) because the data are not sufficient to allow the lowering of that interval (225 to 215 Ma) as accepted by Forster & Warrington.

The remaining uncertainties are even more underestimated for the pre-Kazanian times (Carboniferous and so called early and middle Permian stage boundaries). For example, there are no accurate radiometric data able to help in the estimate of the numerical age of the upper and lower limits of the 'Artinskian' stage (= Baigendzinian as recommended by Waterhouse 1978). In the accepted range of 280 ± 5 Ma to 270 ± 5 Ma, the authors quote five items (15 to 20). Two are from plutonics from Sweden (items 19-20) with a very inaccurate stratigraphical control, i.e. somewhere in the Sakmarian or possibly younger. Moreover, the same analytical results may be interpreted in several ways; a single Rb-Sr isochron age of 270 ± 2 Ma may be split in four lines corresponding to ages ranging from 269 ± 7 to 278 ± 7 Ma (see NDS 240). Two other items are on Australian samples (items 17 and 18). The stratigraphical data only indicate that the volcanics, and the related pluton, are older than a 'probably Sakmarian marine fauna'. One may note that the corresponding apparent ages of 270 ± 2 (!) Ma and 275 ± 6 Ma for the pluton and the volcanics respectively argues for a lower boundary of the Baigendzinian ('probably' located above these samples) younger than these ages and not at 280 ± 5 Ma as accepted by Forster & Warrington. Item 16 at 282 ± 10 Ma corresponds to a badly defined stratigraphical location, post early Permian. Item 15 at 286 ± 7 Ma is for a post-Westphalian 'probable' Autunian flora. These two ages can hardly allow the definition of any of the Baigendzinian boundaries with a precision of ± 5 Ma. The author suggests therefore, that Forster & Warrington have given too precise an estimate in spite of their very complete critical review.

The author's last comments will be to underline how the choice of a too small ± could be supported by further results and how it could also prove to he incorrect in the future. The boundary concerned is that of the Devonian–Carboniferous. The estimate of 365 ± 5 Ma proposed hy Forster & Warrington is essentially based on the numerous results available from Australia (item 2) and on lavas from the Scottish Borders. The Scottish lavas dated at 361 ± 7 Ma may well be Carboniferous as accepted by the authors but definitive arguments which reject a latest Devonian stratigraphical age are not known by the author so that, given the ± and the stratigraphy, this age is not fully incompatible with a Devonian–Carboniferous boundary at 360 Ma or less.

Concerning the Australian Frasnian 'Cerberean volcanics' and the post Cerberean (but still Devonian) later plutonic rocks (see NDS 235), Forster & Warrington use the radiometric data from these rocks to place the Devonian–Carboniferous boundary at 365 Ma. This is older than the data accepted by the authors of the original studies: both Richards & Singleton (1981) and Williams *et al.* (1982) proposed a boundary *younger than* 360 Ma. The author usually considers it very important to take into account the conclusions of the original authors of the studies themselves when evaluating radiometric results (except when obvious errors exist) because they actually know the field and the reliability of their equipment better than it is possible to express in a paper. This is one of the reasons why the author has gathered, in his own synthesis, abstracts written or cosigned by the original authors themselves. In the synthesis by Gale & Odin (1982) it was emphasized that 'The granites (post Cauldron intrusions: NDS 135) are 360 to 365 Ma old and are stratigraphically clearly younger than the volcanics (item 2 of Forster & Warrington and NDS 134). Richards (pers. comm., June 1981) also emphasizes that following the granites, there is quite a thickness of sediment before we come to indubitably Carboniferous fishes (the Broken River fauna). This is why Marsden (1976) proposed an age as young as 345 Ma for the boundary, although 350–355 Ma might be correct'. According to this comment, the top Devonian might well be younger than 360 Ma, the implicit youngest age accepted when one proposes 365 ± 5 Ma.

However, the possibility of a limit younger than 360 Ma is also shown by the results proposed from Russian late Devonian rocks at 335 ± 10 Ma to 355 ± 20 Ma. The same possibility is shown in the abstracts NDS 133 and NDS 229 quoted above, and by early to very early Carboniferous French occurrences not considered by Forster & Warrington. The author would agree if individual analytical results are not accepted, but the series of results, as a whole, does seem to be meaningful.

The author therefore suggests that the Devonian–Carboniferous might be younger than the younger limit proposed by Forster & Warrington and recommends the use of the interval of time 350 to 365 Ma to locate the boundary (Fig. 1).

Conclusion

Once more, it must be underlined that the authors work is notable for its thoroughness and objectivity and is a good reference for people who need numbers. It should be suggested, however, that the uncertainty linked with these recommended numbers is probably too low in several occasions.

From this review and the figure proposed, it appears that four of the 16 stratigraphic units considered need to be improved with new analytical results. They are the Tournaisian, the Westphalian, the Baigendzinian and the Norian stages. The Tournaisian and Westphalian stages could be easily studied in different countries of Western Europe and the presently working IGCP Project 196 will encourage researches on these questions.

Reply by S. C. Forster and G. Warrington

The authors thank Dr Odin for his kind remarks regarding their criteria for the selection of radiometric data and the proposed time-scale for Carboniferous to Triassic times. In view of the rather different data base and interpretation favoured by Dr Odin it is indeed remarkable that we have arrived at such similar conclusions.

The radiometric dates quoted by Dr Odin from French workers were considered by the authors, together with many other relevant data published in the French literature, but were not suitable for inclusion in the final synthesis as they failed to meet the criteria established in the paper.

Regarding the errors ascribed to the stage boundaries: using the data available at present, quoted above in the paper, ± 5 Ma is thought to be a reasonable estimate of the errors associated with those boundaries. The authors do not share Dr Odin's confidence in further refining the error on some boundaries to less than $\pm$ 5 Ma.

The age of the Devonian–Carboniferous boundary will obviously continue to be a discussion point and will only be resolved by further research, initiated in new localities, on suitable rocks located close to the boundary. At present all estimates of the age of the boundary rely heavily on the Australian data; this is unsatisfactory since the majority of the results are from granite plutons, which do not provide the most accurate calibration points on the geological time-scale.

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Jurassic to Paleogene: Part I
Jurassic and Cretaceous geochronology and Jurassic to Paleogene magnetostratigraphy

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SUMMARY: From the Jurassic onwards the relatively new technique of magnetostratigraphy becomes of increasing importance in geochronology and hence demands extended treatment. The original intention was to include, in one large contribution, accounts of Jurassic to Paleogene geochronology based on biostratigraphy and radiometric dating, together with separate accounts of magnetostratigraphy of deep-sea cores and land sections. Whereas, however, the Jurassic and Cretaceous chronology could be reviewed fairly concisely, this did not prove possible for the Paleogene, and a detailed discussion of biostratigraphic correlation problems and dating was necessary. Furthermore, the revised chronology is based on a new magnetostratigraphic analysis integrated with the other data. To have included the Paleogene geochronology with the rest of the material would have destroyed the overall balance and anticipated the more general and introductory accounts of magnetostratigraphy. Consequently our report is divided into two parts, the second part dealing exclusively with the new detailed analysis of the Paleogene. There is inevitably overlap with regard to the Cretaceous–Tertiary boundary but since this boundary is one of exceptional geological interest there can be little objection to thorough analysis by independent workers.

In part I Jurassic and Cretaceous geochronology is reviewed successively. These sections are followed by an introductory account of magnetostratigraphy and then a general review of magnetostratigraphic investigations for the Jurassic to Paleogene interval. The concluding section deals with some geological implications of the results reported earlier.

The ages discussed here in relation to the Jurassic and Cretaceous time-scale have all been recalculated to accord with the decay constants recommended during the 25th International Geological Congress (Steiger & Jäger 1977). Furthermore the discussion relies heavily on the recent compilation of time-scale data by Odin (1982); ages itemized and discussed in volume two of Odin's work are here further referred to by their item number with the prefix NDS (Numerical Dating in Stratigraphy). Author's contributions are identified by initials at the end of each section.

A. H.

Jurassic time-scale

Jurassic radiometric dating is based upon the analysis of (a) igneous intrusive rocks in the midst of Mesozoic strata, (b) intercalated volcanics and (c) glauconite extracted from biostratigraphically dated sediments. All the glauconite and the volcanic biotite dates are derived from potassium-argon isotope ratios but some intrusives have also been dated using rubidium-strontium or uranium-lead ratios.

In addition to analytical error there are three principal sources of error. Although acid intrusives are mineralogically the most favourable for dating purposes, because in at least some cases different isotopic methods may be used for cross-checking purposes, the stratigraphic interval during which they were intruded is usually if not characteristically known only within rather wide limits. For example, Howarth (1964) cited a biotite date from the Shasta Bally batholith as indicating an early Tithonian age, but in fact the batholith can only be dated stratigraphically as post-Kimmeridgian – pre-Barremian (Armstrong 1978).

Secondly, the potassium-argon method of dating is complicated by the possibility if not likelihood of argon loss. This point is well illustrated by a series of analyses of basalt lava specimens from the East Moray Firth basin of the North Sea. Although the stratigraphic evidence indicates that they were extruded no later than the Bathonian, the $^{40}Ar/^{40}K$ age range is 165 to 109 Ma (Harrison *et al.* 1979). Quite clearly, the majority if not all the results must be considered only as minimum ages, with some giving gross underestimates.

The third source of error lies in the determination of decay constants. Revision of these in recent years has led to the alteration of earlier-cited dates. When it is borne in mind that the number of well established, reliable and precise dates for the Jurassic is regrettably small, it would seem over-optimistic to attempt at this time a high measure of precision for stage-boundary determinations.

Van Hinte (1976a) should be consulted for a comprehensive historical review of Jurassic radiometric dating. For our purposes we may start with the Geological Society's first effort at establishing a Phanerozoic time-scale. For the Jurassic, Howarth (1964) accepted only twelve of the then-published radiometric dates as being sufficiently accurate to be used. A time span of 135 to 190–195 Ma was proposed and the eleven Jurassic stages were assumed to be of equal duration, 5 Ma, except for three (Oxfordian and Toarcian, 6 Ma, Hettangian, 4 Ma) for which no reason was given.

In an amendment to the Geological Society time-scale, Lambert (1971) considered that none of the glauconite dates cited by Howarth was reliable and argued that the stratigraphic location of the other radiometric dates was vague. He suggested rounding off to 5 Ma the ages of all major pre-Tertiary boundaries and proposed ? 135 Ma for the Jurassic–Cretaceous boundary; for the Triassic–Jurassic boundary he could not decide between 195, 200 and 205 Ma.

Van Hinte (1976a), using a slightly modified sequence of stages, adopted on the basis of this earlier work lower and upper limits for the Jurassic of 192 and 135 Ma, together with 165 Ma mid Jurassic (Bajocian–Bathonian boundary) age. His further subdivisions were based on the assumption that ammonite zones (± 1 Ma) were a better indicator of duration than stages, which are merely assemblages of variable numbers of zones.

Van Hinte's time-scale has been widely utilized in the last few years but, as Armstrong (1978, 1982) has pointed out, he only used dates published before 1971 and did not even recalculate the dates to a single set of decay constants. Armstrong (1978) utilized newly determined decay constants for K, Rb and U to recalculate all the earlier dates in a comprehensive survey. The primary effect of this for the Jurassic was to revise upwards the upper and lower boundaries to 144 and 211 Ma. On the basis of further dating of igneous rocks in British Columbia, Armstrong (1982) has slightly lowered the date for the start of the Jurassic to 208 Ma. Kennedy & Odin (1982) give a full account of earlier dating for the Jurassic and arrive at a similar figure (205 Ma) for the start of the Jurassic. For the end of the period, however, they utilize European glauconite dates to obtain a figure of 130 Ma, which is significantly lower than Armstrong's.

New analysis of data

Because, as noted, existing radiometric data are inadequate to determine Jurassic stage boundaries with satisfactory precision, it seems pointless in the absence of further results to add to the recent reviews of Armstrong (1978, 1982) and Kennedy & Odin (1982). It was felt to be more useful to attempt to refine Van Hinte's (1976a) method of subdividing the Jurassic column into biostratigraphic units and then to use this subdivision to compare the somewhat divergent results obtained by the above-cited workers.

It is necessary initially to settle upon best estimates for the lower and upper boundaries of the period. Data from igneous rock determinations, principally from western North America but also from southeast Asia and Australasia, have led to a consensus that the Triassic–Jurassic boundary lies between about 200 and 210 Ma. Thus Webb McDougall (1981) proposed 200 and Kennedy & Odin 205, while Armstrong (1982) has reduced his estimate to 208 from an earlier one (1978) of 211. Forster & Warrington (this volume, and the authors propose a figure of 205 ± 5 Ma.

The situation for the Jurassic–Cretaceous boundary is much less satisfactory. Recalculation of Howarth's (1964) date, due to decay constant revision, gives 138 Ma, and Armstrong (1978) has proposed an even higher figure of ~ 144 Ma, principally on the basis of intrusives in western North America which are rather imprecisely dated biostratigraphically. Kennedy & Odin (1982), on the other hand, lower the age of the boundary to 130 ± 3 Ma. In support of this claim they cite a number of glauconite dates from the Portlandian of Europe ranging between 129 and 134 Ma, and a date for the Shasta Bally batholith of 131 Ma. With regard to the last cited date, Armstrong (1978) not only cites a higher figure of 139 but states that the batholith can only be dated as post Kimmeridgian–pre Barremian. In view of the possibility of argon loss the glauconite dates are more likely to be underestimates than overestimates, so more credence should be placed on the date of 134 Ma obtained from the *Hectoroceras* zone of the Ryazanian (Sandringham Sands) of Norfolk by Dodson *et al.* (1964; NDS 134).

The authors propose a date of 135 ± 5 Ma. This is a tentative compromise between A. H. and J. M. H. and the former would have preferred a higher figure.

As already pointed out, Van Hinte (1976a) rightly focused on ammonite zones rather than stages in his effort to achieve a Jurassic time-scale, but biostratigraphical subdivision in the classic sections of NW Europe can be carried further, to the level of subzone. Table 1 gives the number of zones and subzones currently recognized in the British Jurassic and which are widely recognizable across NW Europe (Cope *et al.* 1980). It will be seen that, whereas for the Hettangian to the Oxfordian the number of subzones varies between two and three times the number of zones, for the two youngest stages zones and subzones are virtually equivalent, indicating a different approach adopted towards biostratigraphic subdivision. Therefore the basic units of Jurassic time, here termed *chrons* and presented in the right hand column of Table 1, are not entirely subzones. Van Hinte's estimate of ~ 1 Ma for the mean duration of ammonite zones remains unchanged.

The only exception to total correspondence with the minimal biostratigraphic divisions given in the middle column of Table 1 concerns the Bathonian. This is because the British Batbonian is for the most part developed in facies in which

TABLE 1. Number of zones, subzones and chrons per Jurassic stage, based on Cope *et al.* (1980).

STAGES	ZONES	MINIMAL STRATIGRAPHIC UNITS (USUALLY SUBZONES)	SELECTED CHRONS
Portlandian	9	9	9
Kimmeridgian	13	16	16
Oxfordian	8	16	16
Callovian	6	14	14
Bathonian	8	11	16
Bajocian	7	16	16
Aalenian	3	8	8
Toarcian	6	16	16
Pliensbachian	5	15	15
Sinemurian	6	17	17
Hettangian	3	6	6
TOTAL	74	144	149

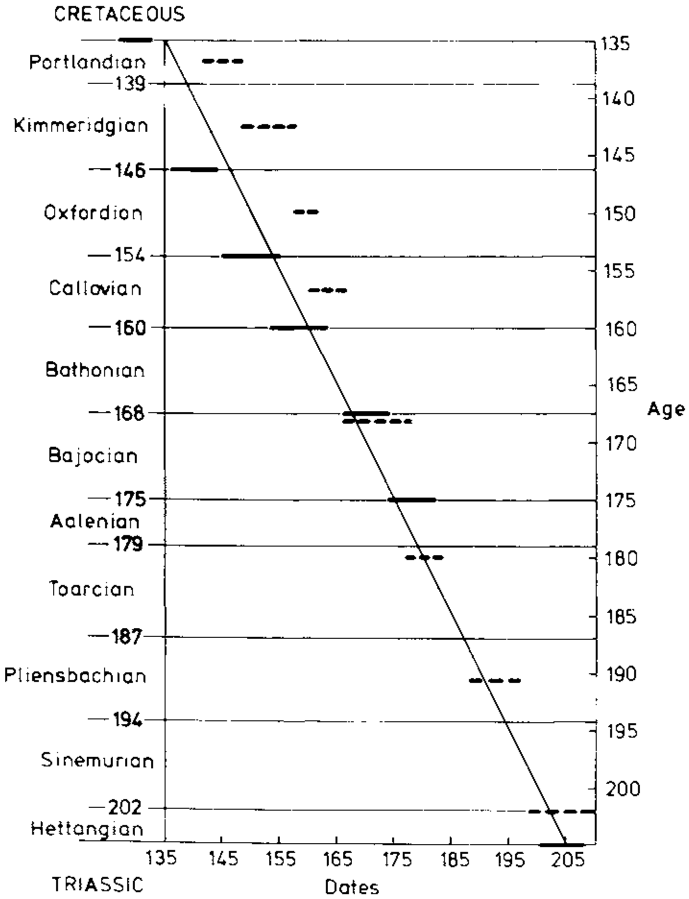

FIG. 1. Stage and system boundary dates for Jurassic, ± 5 Ma. Solid horizontal bars represent results of Kennedy & Odin (1982), broken horizontal bars represent results of Armstrong (1978).

ammonites are sparse or absent. It is not unreasonable to presume that further subdivision would be possible if there were a widespread occurrence of marine shales as in the Lias. Furthermore, overall similarity of stratal thickness in many regions suggests that the Bathonian and Bajocian are roughly time equivalent. The change in question is an inconsiderable one, eleven subzones being converted to sixteen chrons to match the Bajocian, and does not therefore seriously affect the results of this analysis.

The presumption that the ammonite chrons are roughly time equivalent needs to be tested against radiometric data. This is done in Fig. 1, in which age is plotted against measured dates. The slope of the line relates to the vertical (= time) spacing of stages, which is calculated according to the number of chrons. The continuous horizontal bars give the stage boundary confidence limits of Kennedy & Odin (1982); prior to the Bajocian their dating is more tentative, with no confidence limits being given, and so their results are not included. The broken horizontal bars give the more tentative age ranges for Jurassic stages of Armstrong (1978); his new results for the early Jurassic (Armstrong 1982) do not require any adjustment of the earlier figures.

The 'chron line' and the two sets of results match well up to the Bathonian but Kennedy & Odin's dates depart increasingly from Armstrong's from the Callovian onwards, with the line effectively bisecting them. Because Kennedy & Odin's late Jurassic dates are based on glauconites it seems likely that they are systematically underestimated by several percent. This can only be confirmed by more and better data for the Upper Jurassic, for instance from as yet undated lavas intercalated between ammonitiferous strata in the Southern Andes. The overall correspondence is close enough, however, to use the line to establish provisional stage boundaries as indicated in Fig. 1, albeit within rather wide confidence limits.

As the Jurassic chrons do seem to be broadly time-equivalent this implies that the ammonite turnover resulting from evolution and extinction was essentially regular in time, signifying a kind of clock. Such an inference is supported for most of the period by stage-by-stage analysis of stratal thickness in Britain (Hallam & Sellwood 1976). More data are required, however, to confirm the suggestion.

A. H.

Cretaceous time-scale

The time-scale used here is not derived from any other single scale elsewhere but is based on evidence from a variety of sources. The principal papers used have been those by Obradovich & Cobban (1975) and Kennedy & Odin (1982) but consideration has also been given to the general surveys by Casey (1964), Lambert (1971), Zotova (1972), Van Hinte (1976b), Kauffman (1977), Armstrong (1978), Lanphere & Jones (1978) and Odin & Kennedy (1982), in addition to various papers dealing with individual parts of the system.

Cretaceous radiometric dates are subject to a number of inaccuracies and limitations that are not obvious from a simple table. Almost all the dates are based on $^{40}Ar/^{39}Ar$ or $^{40}Ar/^{40}K$ ratios dependent on the breakdown of ^{40}K. Re-measurement of the constants for these isotopes during the last few years has itself increased many of the dates published up to the mid 1970s by some 2.5%. Those from the USSR need to be reduced by 2.5%, but even the adjusted dates by Zotova (1972) are at least 3 Ma older than the scale proposed here. Most of the potassium occurs in one or other of two very different ways: in autochthonous sedimentary glauconites, or in biotite or sanidine in tuffs. The general problems associated with dating from glauconite are discussed by Odin (1978 and 1982). In spite of the precautions taken by Odin and his co-workers and the reasonable internal consistency of their results, comparison with dates from tuffs still suggests that there may have been a frequent small loss of argon giving rise to dates that are too low. The problems associated with dates from biotite and sanidine have been discussed by Folinsbee *et al.* (1983), Obradovich & Cobban (1975) and McDougall (1978).

Further problems for some Cretaceous examples come from difficulties of assigning reliable or internationally agreed biostratigraphic data. There is a good discussion of these by Kennedy & Odin (1982) and only critical examples will be mentioned here. Thus arguments over the biostratigraphic definition of the Jurassic–Cretaceous boundary are hardly relevant when it cannot be fixed radiometrically more closely than about ± one age.

In general the earlier early Cretaceous ages (Berriasian to Barremian) have a reliability of only ± 5 Ma. The Aptian dates are little better, but those of the Albian are probably correct to ± 3 Ma. The late Cretaceous dates are now reliable to about ± 1 Ma as straight radiometric values, but for the late Turonian or Coniacian to Maastrichtian ages the biostratigraphic uncertainties are of the same order of size as the potential radiometric errors. Whatever their actual accuracy, there is a considerable range in the confidence held by different authors in Cretaceous radiometric dates, from Kauffman (1977) who gave a 'generalized' scale with some dates to the nearest 50 000 years, to Naidin (1981) who has cautiously refrained from giving his own table of dates at all (see also discussion in Naidin 1982).

Late Cretaceous

The dates, still largely based on Obradovich & Cobban (1975), are mainly from K/Ar ratios in carefully selected biotites and sanidines from volcanic ashes in the Upper Cretaceous of the western interior of the United States (chiefly Montana, but also Wyoming, Colorado and South Dakota).

66 Ma — The end of the Maastrichtian (= Cretaceous–Paleogene boundary)

There are no radiometric dates close to this boundary from Europe. In the western interior of North America the sea retreated out of the region during the Maastrichtian. Hence the boundary lies within non-marine sediments and has been based on the highest occurrence of large reptiles – dinosaurs in popular parlance. The biostratigraphy of these Triceratops Beds is best known in Montana, Saskatchewan and Alberta where Lambert's review of the data (1971) gave a date of 65.6 Ma. Other dates have been obtained near Golden, Colorado, on a biotite-bearing dacite pumice (Obradovich & Cobban 1975; Newman 1979): one from 35 m above the Cretaceous–Paleogene boundary gave 65.9 Ma (NDS 103). These are minimum ages for the boundary and the probable date is 66 Ma.

The reptile-boundary in that region coincides with that based on palynomorphs (Newman 1979), but extending this date to Europe and other regions with marine sequences across the boundary is not as easy as is sometimes assumed. The usual molluscan-correlators are nearly impotent. The highest ammonite fauna in the western interior is found in the Fox Hills Formation, known chiefly in South Dakota and Wyoming. Jeletzky (1962) considered these ammonites to be early Maastrichtian, but from a detailed analysis of the assemblage in 1968 Waage concluded that one could not date this formation more accurately than plain Maastrichtian (on the occurrence of *Sphenodiscus* and *Scaphites* (*Hoploscaphites*), both long-ranging genera). The overlying non-marine beds, variously called Hell Creek, Lance, or Edmonton Formations, and which may be diachronous, contain the terminal dinosaur faunas (Jeletzky 1960). These are overlain in turn by the Cannonball, Ludlow and Tullock Formations; the first of these has been dated accurately in North Dakota by Fox & Olsson (1969) to the foraminiferal Zone of *Globigerina edita*, the basal Subzone of the Paleocene in the critical Zumaya section of north Spain (Hillebrandt 1965). There is thus no doubt that the latest North American dinosaurs are pre-Paleocene on the European marine scale, and their extinction could be simultaneous on the biostratigraphic scale with that of the well-known nanoplankton-planktic foraminiferal extinctions.

Attempts have been made to check the correlation between Europe and North America on the geomagnetic polarity scale. The boundary at Gubbio in northern Italy is in a zone of reversed (negative) magnetic polarity (between normal intervals 29 and 30 on the scale of Lowrie & Alverez 1981), whilst the boundary taken in New Mexico, Alabama and various places in the western interior of North America is in a zone of normal (positive) polarity (Butler *et al.* 1977; Kent 1977; Russell 1979). Assuming that the difference between Italy and these North American localities does not involve more than adjacent polarity intervals, the chances are that this would not mean a time-difference of more than 200 000 years. However, Archibald *et al.* (1982) now claim that the Z lignite on the boundary (fixed biostratigraphically by vertebrates) in Garfield County, Montana, lies in different polarity zones at localities only 15 km apart. Since the biostratigraphic boundary on vertebrates still agrees with that on palynomorphs as far south as Golden in Colorado (Newman 1979), there is an urgent need to re-check polarity boundaries. In the meantime it is probably not safe to try to make geomagnetic polarity correlations between Europe and North America.

71.5 Ma — The Maastrichtian–Campanian

Obradovich & Cobban (1975) obtained dates in the western interior of the United States for the Zones of *Didymoceras nebrascene* (74.0 Ma), *Exiteloceras jenneyi* (73.8 Ma) and *Baculites compressus* (73.3 Ma) and for the unquestioned Maastrichtian Zone of *Baculites grandis* (70.2 Ma). These are in general agreement with those obtained by Folinsbee *et al.* (1966) in Alberta. Unhappily there is considerable uncertainty on the biostratigraphic position of the boundary in this region because the ammonite faunas are indigenous. If one takes Jeletzky's ammonite boundary (also used by Kauffman 1977), which according to Cobban lies above the Zone of *Baculites eliasi* and below the Zone of *Baculites baculus*, one gets a date around 70.7 Ma. Cobban's own boundary would be below the Zone of *Baculites reesidei* which would give a date of approximately 71.9 Ma. Pessagno (1969) chose a boundary based on the foraminiferal succession in Texas–Arkansas that would correspond, according to Cobban, to beneath the Zone of *Didymoceras stevensoni* and a date of nearly 74 Ma. Glauconite from the Atlantic coast (biostratigraphic correlation better than in the western interior) suggests a maximum age of 71.6 ± 2.7 Ma (Owens & Sohl 1973).

Kennedy & Odin (1982; NDS 139) quote Priem *et al.* (1975) as giving glauconite dates from Limburg (Belgium-Netherlands) for the base of the Maastrichtian there at 71.5 ± 2.5 Ma. Actually only their samples two and seven are well fixed biostratigraphically and these are from a considerable way above and below the boundary. It is not clear how Kennedy & Odin have calculated their figure, but accepting the two good dates and extrapolating to the boundary gives 71.3 Ma.

It is suggested that 71.5 Ma be used here.

84 Ma — The Companian–Santonian

Obradovich & Cobban (1975) obtained a date on a biotite from the Zone of *Desmoscaphites bassleri* of 84.4 ± 1.6 Ma (NDS 107). This zone contains the free-swimming crinoids *Uintacrinus* and *Marsupites* that are accepted in northern Europe as marking the Upper Santonian (Rawson *et al.* 1978), and Obradovich & Cobban suggested 82 Ma for the boundary, which would now be 84.1 Ma. However, there are other dates from North America that suggest the boundary may be later or earlier than this. William & Baadsgaard

(1975) dated beds in Saskatchewan equivalent to somewhere in the range of the Lower Campanian Zones of *Scaphites hippocrepis* III to *Baculites mclearni* of 82–83 Ma, whereas Obradovich & Cobban had obtained dates around 80–82 Ma for the same interval, i.e. 1–2 Ma less, so that the boundary might be 85–86 Ma. Kennedy & Odin (1982; NDS 163) quote a basal Campanian lava in Texas as only 81.5 ± 3 Ma.

In Europe there was a date obtained by Evernden *et al.* (1961) from a Lower Campanian glauconite at 500 m depth in a core near Hannover that gave a date of 83.5 Ma (recalculated).

At present 84 Ma still seems the most sensible value to take.

88 Ma — The Santonian–Coniacian

Apart from a date of 84.6 Ma for the top zone of the Santonian, Obradovich & Cobban (1975; NDS 107) obtained no other dates within the Santonian. Williams & Baadsgaard (1975) had too wide a range of dates for the mid-Santonian to be of any use. Evernden *et al.* (1961) obtained an age of 85 Ma for a Lower Santonian glauconite from a core near Salzgitter in Germany, but Kennedy & Odin (1982) regard this as untrustworthy. In taking the figure for the boundary at 88 Ma the author is not indicating accuracy but merely 'guestimating' from the thickness of sediments and the number of zones that can be recognized.

89.5 Ma — The Coniacian–Turonian

There is considerable uncertainty at present on where this boundary lies in the western interior of the United States. However, no-one doubts that the Zone of *Inoceramus deformis* is well within the Coniacian, and probably the lower half. Obradovich & Cobban (1975) obtained a date of 89 Ma for this zone, whilst Williams & Baadsgaard (1975) in Saskatchewan give 91.2 to 92.2 Ma. Kennedy & Odin (1982; NDS 108) quote 88.9 ± 1.5 Ma for this same zone in Montana.

There are also a fair number of Coniacian dates from glauconites in Europe given by Kennedy & Odin (1982), the most reliable being from Belgium at 88.1 ± 3.0 Ma and 88.7 ± 2.2 Ma. All these results point to the base of the Coniacian being around 89.5 Ma.

91 Ma — The Turonian–Cenomanian

Obradovich & Cobban (1975) dated a bentonite from the lower part of the Zone of *Inoceramus labiatus*, i.e. very close to the boundary, as 91.1 Ma. Williams & Baadsgaard (1975) obtained figures of 91.2 to 92.3 Ma in Saskatchewan which they considered to be close to those of Obradovich & Cobban, but their sample was biostratigraphically appreciably younger, probably by several ammonite zones. Kennedy & Odin (1982; NDS 118) quote ages from five biotites of the *labiatus* Zone in Alaska as 91.5–93.6 Ma.

This is the boundary that is well correlated biostratigraphically between North America and Europe (Wright & Kennedy 1981), but the European dates from glauconites are slightly younger (Kennedy & Odin 1982). These authors quote five analyses for the boundary made by Kreuzer *et al.* in NW Germany at 89.0–91.2 Ma (NDS 226). They also quote late Cenomanian dates from the Paris Basin ranging from 89.5 ± 3.3 Ma to 91.2 ± 2.0 Ma (the last from an Rb/Sr ratio) (NDS 59 and 81).

One can be reasonably confident in a figure of 91 Ma for this boundary.

95.5 Ma (? 95) — The Cenomanian–Albian

There are difficulties in finding the biostratigraphic Albian–Cenomanian boundary in the western interior of North America, partly from a scarcity of fossils, partly because in Montana and southern Canada the ammonites differ from those elsewhere in the United States and Europe. It is possible that the neogastroplitid zone of *N. maclearni* marks the base of the Cenomanian (Warren & Stelck 1969). For this zone Folinsbee *et al.* (1963) suggested an approximate date of 94 Ma, which with the new constants would be 96.4 Ma. This fits with the date of 97.6 ± 2.0 Ma that Obradovich & Cobban (1975; NDS 111) obtained for the zone of *Neogastroplites cornutus*, four zones earlier. Kennedy & Odin (1982) have recognized these and other data from that region with the help of further information from Baadsgaard, and conclude that the Albian–Cenomanian boundary probably lies between 93 and 96 Ma; they prefer the younger end of this scale, but they have taken the *Neogastroplites maclearni* Zone to be late Albian.

Early Cenomanian glauconite from Devon, England, is reported to be 94–95 Ma (Kennedy & Odin 1982; NDS 96). In the Paris Basin Juignet *et al.* (1975) considered the boundary to lie between 94.5 and 96.5 Ma, and suggested 95 ± 1 Ma for this date; this is confirmed in the more recent paper by Odin & Hunziker (1982; NDS 62–68, 85).

In view of the uncertainty of the biostratigraphic dating in Montana and Canada and the extensive data from Europe, Kennedy & Odin's figure of 95 Ma seems sensible, although in our opinion the American evidence still points to 96 Ma.

Early Cretaceous

Evidence from North America peters out below the Upper Cretaceous. Lanphere & Jones (1978) had some dates from California and Alaska but most of them are badly placed biostratigraphically. We are largely dependent on glauconite dates from Europe where there are many gaps in the record. Although such dates can be too high, as Kennedy & Odin (1982) have stressed, it still seems more likely that they are minimum figures.

107 Ma — The Albian–Aptian

Kreuzer & Kemper (in Odin 1982; NDS 143) have dated a basal Albian glauconite from Germany as 106.0 ± 0.8 Ma. Kennedy & Odin (1982) quote early Albian glauconites from Aube in France: the higher was 102 ± 3.1 Ma and the lower 108.6 ± 3.3 Ma. The end of the early Albian in the Boulonnais has been dated by Elewaut *et al.* (in Odin 1982; NDS 78) at 100 Ma by K/Ar and 104.8 Ma by Rb/Sr.

Kennedy & Odin consider that the boundary lies between 106 and 108 Ma, and we follow them in taking it at 107 Ma.

114 Ma — The Aptian–Barremian

A glauconite date by Conrad & Odin (in Odin 1982; NDS 73) for the latest Barremian in SE France gave an apparent age of 110.7 ± 3.6 Ma, but according to Kennedy & Odin (1982) the specimen was tectonized and the age must be regarded as a minimum. However, Kennedy & Odin give a number of dates from Aptian glauconites in Germany, France and England that range from 106 to 114 Ma, and suggest 112 ± 2 Ma for the boundary.

Lanphere & Jones (1978) quote a U/Pb date from zircon in intrusions into the Ailisitos Formation in NW Mexico; in the area studied this formation yields Aptian fossils so that their date of 114 ± 2 Ma might be considered as a minimum age for the Aptian, but the biostratigraphy is too weak to make it of much value.

Jeans *et al.* (1982) have now obtained consistent results from the youngest sanidines in two samples of the Upper Aptian fullers earth of Surrey, England; both are dated as 112 ± 1 Ma. This deposit is so high in the Aptian (Zone of *Parahoplites nutfieldiensis*) that it seems unlikely that the base of the stage could be younger than 115 Ma, but it is suggested that 114 Ma be used until some reliable dates are available close to the boundary.

118 Ma — The Barremian–Hauterivian

All Cretaceous dates earlier than the Aptian must be regarded as approximate. For example, all the Barremian dates obtained so that where the possibility of inheritance is unlikely (e.g. 108.3, 111, 110.7 Ma) are figures that fall into the Aptian as here dated. The figure of 118 Ma suggested here for the boundary is guesswork.

122 Ma — The Hauterivian–Valanginian

The Hauterivian of south-east France has yielded dates that suggest a figure between 115 and 121 Ma for this boundary

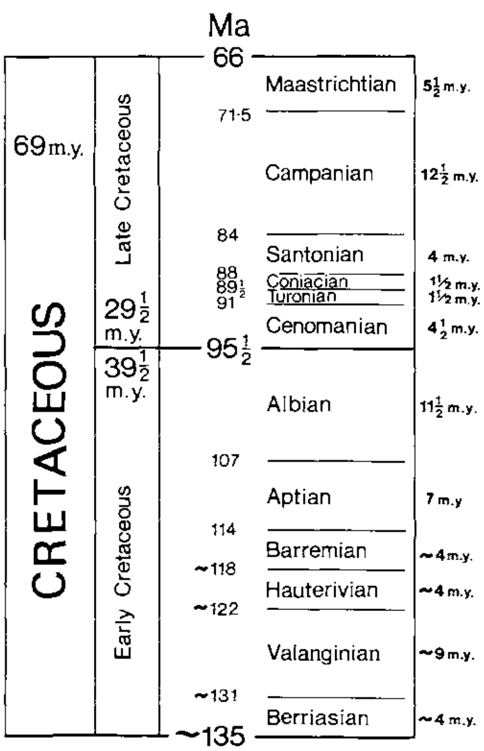

FIG. 2. Stage and system boundary dates for Cretaceous with stage durations on right. Upper Cretaceous dates ± 1 Ma, Lower Cretaceous dates ± 5 Ma.

TABLE 2. Number and estimated duration of Cretaceous ammonite chrons

			Ammonite chrons recognized (a)	Average length (b)
	(Palaeocene)			
		66 Ma		
5½ Ma	Maastrichtian		2–7	3/4–2 3/4 Ma
		71.5 Ma		
12½ Ma	Campanian		4–18	0.7–3 Ma
		84 Ma		
4 Ma	Santonian		2–6	2/3–2 Ma
		88 Ma		
1½ Ma	Coniacian		2–4	3/4–1 1/2 Ma
		89½ Ma		
1½ Ma	Turonian		3–8	1/5–1/2 Ma
		91 Ma		
4½ Ma	Cenomanian		4–12	1/3–1 Ma
		95½ Ma		
11½ Ma	Albian		7–22	1/2–1 2/3 Ma
		107 Ma		
7 Ma	Aptian		5–17	0.4–1.4 Ma
		114 Ma		
4 Ma	Barremian		2–7	(1/2–2 Ma)
		118 Ma		
4 Ma	Hauterivian		7	(0.6–Ma)
		122 Ma		
9 Ma	Valanginian		6–14	(2/3–1 1/2 Ma)
		131 Ma		
4 Ma	Berriasian (Ryazanian)		2–7	(0.6–2 Ma)
		135 Ma		
	(Portlandian) (Volgian) (Tithonian)			

(Kennedy & Odin 1982; NDS 74). Markedly younger dates come from the Speeton Clay of Yorkshire for the Hauterivian (108 to 117 Ma, Curry & Odin 1982; NDS 72) and these must be regarded as minima. The figure suggested here of 122 Ma is again guesswork.

131 Ma — The Valanginian–Berriasian

This is guessed to be 131 Ma.

The Cretaceous: some comments

A Cretaceous time-scale, based on the above data is summarized in Fig. 2; Table 2 gives age estimates of Cretaceous ammonite chrons, which present a more variable picture than those of the Jurassic. No distinction has been made between zones and subzones. In general the smaller number of chrons can be distinguished almost world-wide, or at least through a major faunal province, such as the Tethys. The large number of chrons may only be recognizable over a quite limited area, say 100000 km^2, because of ammonite provincialism. The Lower Cretaceous numbers are based on Rawson (1981), Casey (1961) and Owen (1976). The Upper Cretaceous numbers are based on various papers and unpublished work by and with W. A. Cobban, W. J. Kennedy and C. W. Wright.

The figures of column (b) are often approximated; obviously anyone can divide column 1 by column 4. The values for the four oldest ages are put in brackets because they are extremely approximate. One of the main uses of these figures is to give an indication of the accuracy in years of intercontinental correlations. Thus a reconstruction of continental positions in the Santonian may be comparing data that were actually 2 Ma apart if there were a mistake of only one ammonite zone in correlation, and few post Coniacian correlations are as reliable as that. For the Campanian–Maastrichtian there are now nanoplankton zones of only about 1.5 to 2 Ma duration (see Thierstein 1976, 1981; Sissingh 1977; Perch-Nielsen 1979). Upper Cretaceous globotruncanid zones are longer than these, at least below the Maastrichtian, usually more than 3 Ma. Many globotruncanid species are rare outside the Tethys; the same geographical limitation applies to the larger benthic foraminifera (Dilley 1971, 1973; Sigal 1977). Boreal foraminiferal zonations in the Chalk of Europe (e.g. Pozaryska 1956 and Peryt 1979 in Poland; Robaszynski 1980 in NE France; Hart *et al.* 1981 in England) are better. Koch (1977) has erected *Stensioeina* zones for the Turonian to Lower Campanian that allow correlations to ± 1 Ma whilst the *Bolivinoides* zones for the Campanian–Maastrichtian allow correlations to ± 2 Ma.

J. M. H.

TABLE 3. Chron sequence based on magnetic reversals for the Cenozoic and Mesozoic

Cenozoic and late Cretaceous

CHRON	CALIBRATION MA LKC-77	REVERSAL HISTORY
C1	0	C1.4320R, C1.5494N, C1.5864R
C2	1.62	C2.0N, C2.2658R
C2A	2.41	C2A.0N, C2A.3185R, C2A.3704N, C2A.4370R, C2A.5111R, C2A.6741N
C3	3.76	C3.0N, C3.06618R, C3.1544N, C3.2500R, C3.3529N, C3.4044R, C3.4706N, C3.6103R
C3A	5.12	C3A.ON, C3A.1360R, C3A.2480N, C3A.4000R, C3A.7520N, C3A.8480R
C4	6.37	C4.0N, C4.0686R, C4.1372N, C4.5196R, C4.5882N, C4.6372R
C4A	7.39	C4A.0N, C4A.3053R, C4A.5053N, C4A.5895R, C4A.8000N, C4A.8842R
C5	8.34	C5.0N, C5.06226R, C5.07782N, C5.2023R, C5.2062N, C5.3152R, C5.3268N, C5.4358R, C5.4436N, C5.5447R, C5.5914N, C5.6109R, C5.7860N, C5.8132R
C5A	10.91	C5A.0N, C5A.0528R, C5A.0909N, C5A.1701R, C5A.2727N, C5A.2815R, C5A.3079N, C5A.3226R, C5A.3871N, C5A.4399R, C5A.4985N, C5A.5748R, C5A.6452N, C5A.7654R, C5A.8006N, C5A.9384R
C5B	14.32	C5B.0N, C5B.0714R, C5B.1929N, C5B.2929R
C5C	15.72	C5C.0N, C5C.2158R, C5C.2446N, C5C.3741R, C5C.4245N, C5C.5611R
C5D	17.11	C5D.0N, C5D.3333R, C5D.5490N, C5D.5686R
C5E	18.13	C5E.0N, C5E.6585R
C6	18.95	C6.0N, C6.7134R
C6A	20.52	C6A.0N, C6A.1628R, C6A.2965N, C6A.4883R, C6A.6046N, C6A.6977R, C6A.8139N, C6A.8721R
C6B	22.24	C6B.0N, C6B.5694R
C6C	22.96	C6C.0N, C6C.0746R, C6C.1272N, C6C.2325R, C6C.3465N, C6C.4210R
C7	25.24	C7.0N, C7.1222R, C7.2000N, C7.5333R
C7A	26.14	C7A.0N, C7A.3878R
C8	26.63	C8.0N, C8.0601R, C8.1203N, C8.6842R
C9	27.93	C9.0N, C9.3727R, C9.4099N, C9.6708R
C10	29.57	C10.0N, C10.2013R, C10.2403N, C10.4026R
C11	31.11	C11.0N, C11.3130R, C11.3652N, C11.6783R
C12	32.26	C12.0N, C12.1700R
C13	35.26	C13.0N, C13.095R, C13.1300N, C13.3000R
C15	37.26	C15.0N, C15.2500R, C15.2727N, C15.5114R
C16	38.14	C16.0N, C16.1644R, C16.2808N, C16.4794R, C16.5137N, C16.8014R
C17	39.60	C17.0N, C17.5111R, C17.5500N, C17.6667R, C17.7055N, C17.9000R
C18	41.40	C18.0N, C18.1899R, C18.2236N, C18.4093R, C18.4388N, C18.6245R
C19	43.88	C19.0N, C19.2474R
C20	44.85	C20.0N, C20.3699R
C21	49.04	C21.0N, C21.4985R
C22	52.31	C22.0N, C22.3485R
C23	54.29	C23.0N, C23.1163R, C23.1628N, C23.6512R
C24	55.58	C24.0N, C24.0744R, C24.1715N, C24.3301R
C25	58.67	C25.0N, C25.2950R
C26	60.06	C26.0N, C26.1221R
C27	62.38	C27.0N, C27.2708R
C28	63.34	C28.0N, C28.6900R
C29	64.34	C29.0N, C29.5437R
C30	65.37	C30.0N, C30.9456R
C31	66.84	C31.0N, C31.3093R
C32	69.20	C32.0N, C32.0951R, C32.1859N, C32.7438R, C32.8843N, C32.9008R
C33	71.62	C33.0N, C33.6052R
C34	79.65	C34.0N

Cretaceous–Jurassic

CHRON	CALIBRATION MA LH-75	REVERSAL HISTORY
CM0	108.19	CM0.0R, CM0.1851N
CM1	112.62	CM1.0R
CM2	113.14	CM2.0N
CM3	114.05	CM3.0R
CM4	116.75	CM4.0N
CM5	118.03	CM5.0R, CM5.7841N
CM6	118.91	CM6.0R, CM6.1875N
CM7	119.27	CM7.0R, CM7.5532N
CM8	120.21	CM8.0R, CM8.4627N
CM9	120.88	CM9.0R, CM9.5755N
CM10	121.94	CM10.0R, CM10.2402N, CM10.4916R, CM10.5251N, CM10.7654R, CM10.7821N
CM10N	123.73	CM10N.0R, CM10N.2482N
CM11	125.10	CM11.0R, CM11.2248N, CM11.2481R, CM11.4341N, CM11.7984R, CM11.8488N
CM12	127.68	CM12.0R, CM12.4292N, CM12.484, CM12.5982N, CM12.7899R, CM12.8584N
CM13	129.87	CM13.0R, CM13.6136N
CM14	130.75	CM14.0R, CM14.5638N
CM15	132.63	CM15.0R, CM15.2627N
CM16	135.18	CM16.0R, CM16.6129N
CM17	136.42	CM17.0R, CM17.7250N
CM18	138.82	CM18.0R, CM18.2552N, CM18.3333R, CM18.3802N
CM19	140.74	CM19.0R, CM19.3121N, CM19.5202R, CM19.5607N
CM20	142.47	CM20.0R, CM20.4405N
CM21	144.74	CM21.0R, CM21.2083N, CM21.8977R, CM21.9205N, CM21.9432R, CM21.9659N
CM22	147.38	CM22.0R, CM22.5537N, CM22.6384R, CM22.7571N
CM23	149.15	CM23.0R, CM23.2230N, CM23.2432R, CM23.7365N
CM24	150.63	CM24.0R, CM24.2048N, CM24.2190R, CM24.3333N, CM24.4048R, CM24.5524N, CM24.7524R, CM24.8381N
CM25	152.73	CM25.0R, CM25.1648N, CM25.2637R, CM25.3132N, CM25.3791R, CM25.4286N, CM25.5165R, CM25.5824N, CM25.6483R, CM25.6978N, CM25.7472R, CM25.7967N, CM25.8461R, CM25.8791N
CM26	154.55	CM26.0R, CM26.4286N
CM27	154.97	CM27.0R, CM27.3461N
CM28	155.49	CM28.0R, CM28.1805N
CM29	156.82	CM29.0R, CM29.07457N
CM30	165.00(?)	

Problems in magnetostratigraphic time-scale calibration

Chron nomenclature for the Cenozoic and Mesozoic

Magnetostratigraphy provides a high resolution chronological technique to correlate, on a global scale, locally observed events. Recent advances in technology including the development of the cryogenic magnetometer and the hydraulic piston corer have greatly enhanced our ability to determine the magnetostratigraphy of older sediments. It is apparent from the literature that the present nomenclature is inadequate for the expression of magnetostratigraphic correlations. Long and often contrived verbiage is necessary to describe the magnetostratigraphic location of observed events. Because the pre-late Miocene to Jurassic polarity history has been derived from the marine magnetic anomaly record, some magnetostratigraphic correlations are described in terms of the correlative magnetic anomaly. Though such references are convenient, they are not logically correct. Magnetic anomalies are observed physical phenomena and not chronological events. Though the anomaly pattern presents a record of the geomagnetic polarity history it is also a function of many other variables including the local magnetic field vector and local crustal structure.

A division of the Cenozoic and Mesozoic geomagnetic polarity history into chrons based on the magnetic anomaly nomenclature is reasonable for two reasons. First, the anomaly nomenclature is based on the correlation of easily observed features in the magnetic reversal pattern; this is a characteristic which will aid magnetostratigraphic studies. Secondly, the development of another nomenclature to describe the same phenomenon is unnecessary.

During Leg 73 of the Deep Sea Drilling Project (DSDP) the hydraulic piston corer was successfully employed to recover long intervals of the Neogene and Paleogene geomagnetic reversal history. Since no magnetic time units had been defined for the Paleogene, it was necessary to develop a nomenclature to accommodate the new material. To this end, Tauxe *et al.* (1984) have named the Paleogene and late Cretaceous chrons after the correlative magnetic anomaly nomenclature (Heirtzler *et al.* 1968; LaBrecque *et al.* 1977). A particular chron is defined as extending from the youngest reversal boundary of one named anomaly to the youngest reversal boundary of the next anomaly. The intervening polarity history was defined by a chron named for the correlative magnetic anomaly. A letter C (for chron) prefixes the associated anomaly number to eliminate confusion with the pre-existing Neogene chron nomenclature. An N (normal) or R (reversed) is suffixed to define whether the observed polarity interval within the chron is Normal or Reversed. Cox (1982) has developed a similar nomenclature and has

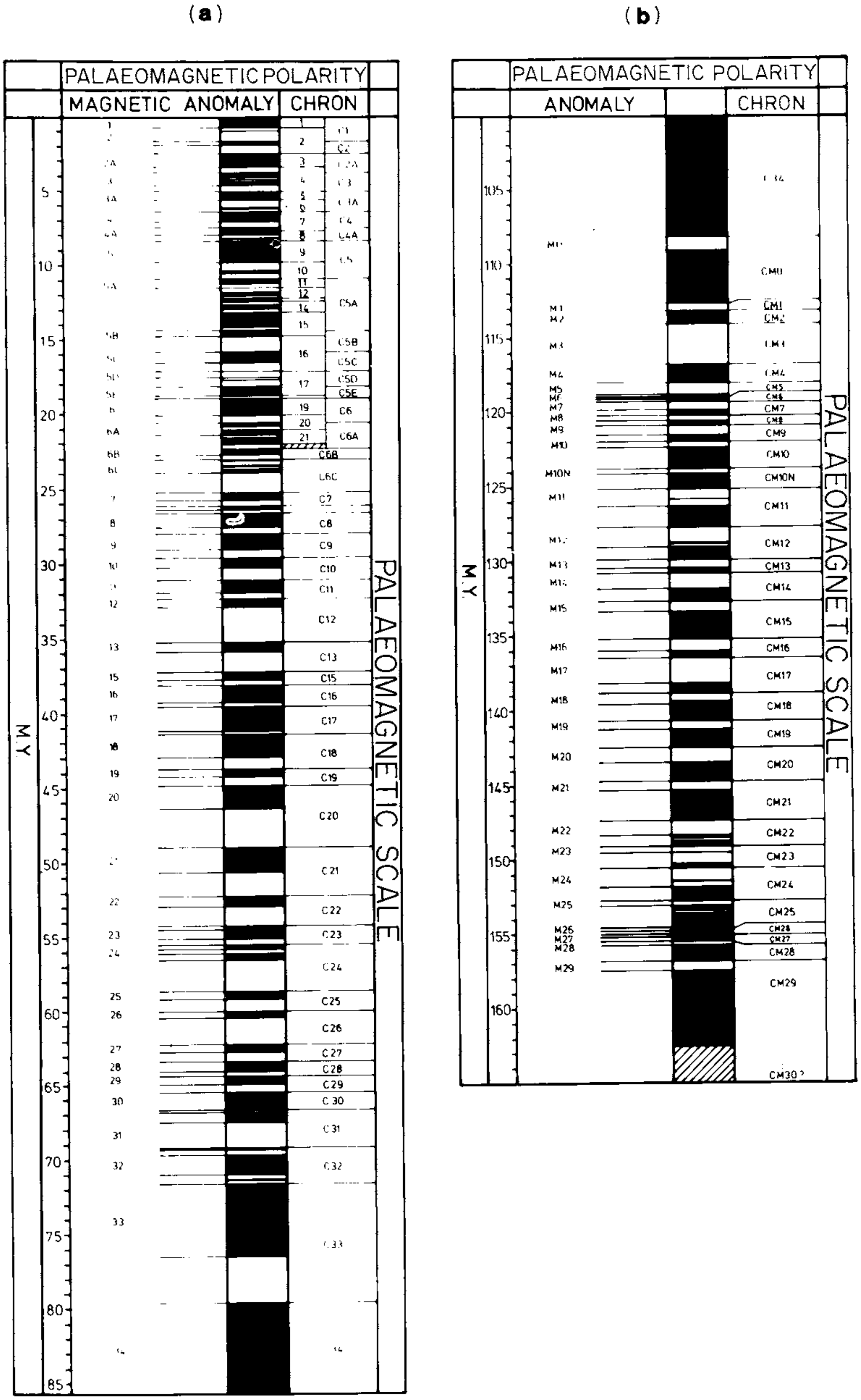

FIG. 3. Chron subdivisions of the Cenozoic and Mesozoic reversal pattern. Absolute ages are from La Brecque *et al.* 1977 and Larson & Hilde 1975. Reversal locations are tabulated in Table 3.

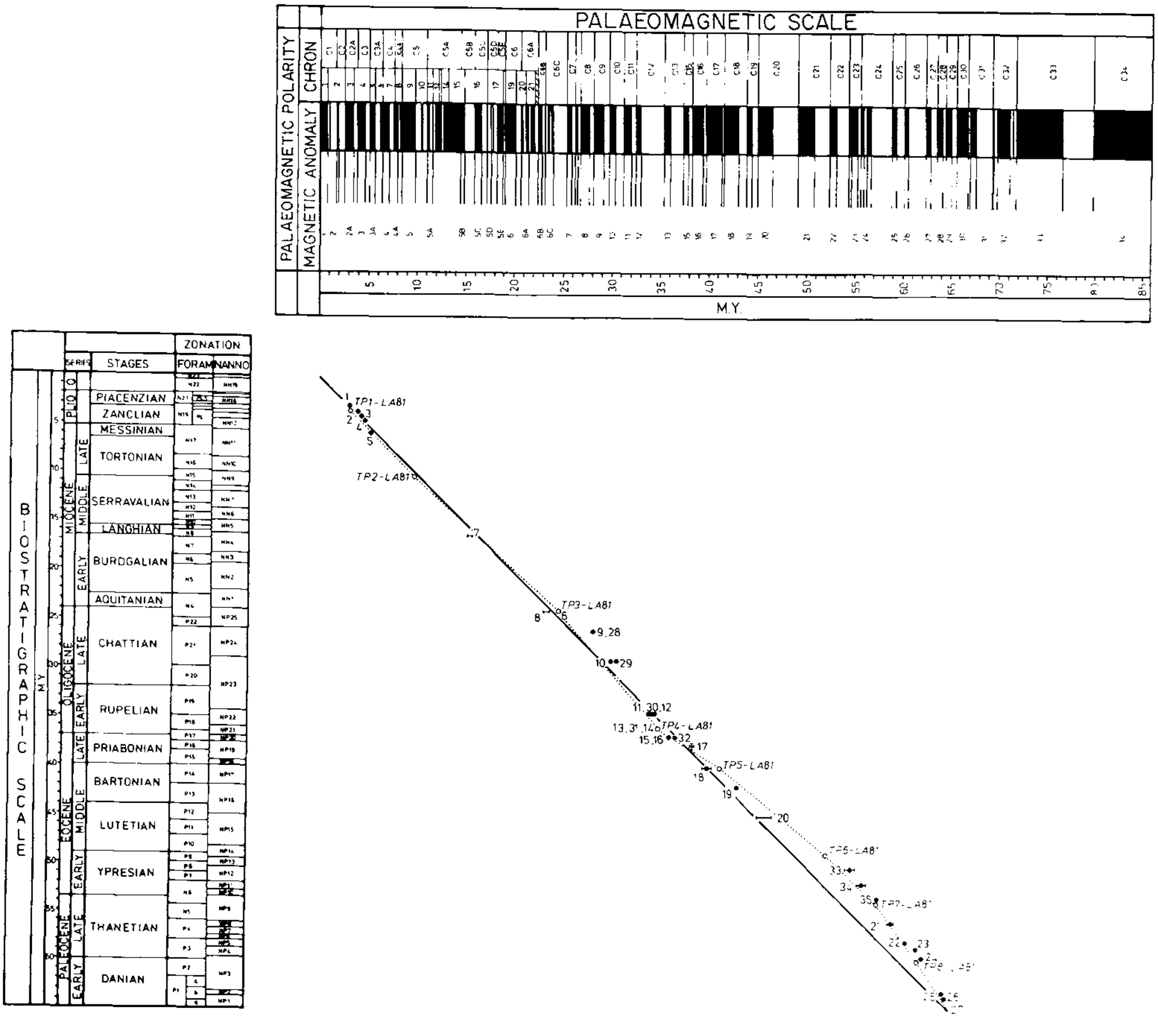

FIG. 4. Observed nanofossil zonal correlations with magnetostratigraphy. Reference numbers refer to listing in Table 4. Dotted line represents revised magnetostratigraphic correlations of Lowrie & Alvarez (1981).

extended this nomenclature to the Neogene and Mesozoic reversal history. The nomenclature of Cox provides for revisions of the polarity history by suffixing additional symbols to the chronal names.

We suggest an extension of this scheme which is numerical in nature and which may locate arbitrary events (e.g. polarity reversal, biostratigraphic zonal boundaries, etc.) with unlimited resolution with respect to the chronal boundaries.

The chron is the basic unit of geomagnetic field history (Anonymous 1979). Furthermore, the position of an event is defined as the relative position in time or distance between the younger and older chronal boundaries. The relative position of an event is expressed as a decimal appended to the chron name. For example, the location of the Cretaceous–Tertiary Boundary is defined by planktonic foraminifera at C29.75R in Gubbio (Alvarez *et al.* 1977). The K/T boundary is defined by nanofossils at C29.5R in DSDP hole 524 (Hsü *et al.* 1982), and the K/T boundary defined by dinosaur extinctions as C28.5R to C21.1N in the San Juan Basin (Butler *et al.* 1977; Lindsay *et al.* 1978). The location of an event can be expressed with any degree of resolution necessary, it is independent of the absolute age assigned to a given chron. Figure 3(a) displays the chron divisions suggested by Tauxe *et al.*, (in press), and Cox (1982) for the Cenozoic and late Cretaceous. Figure 3(b) presents the Mesozoic chron boundaries modified from Cox (1982). This suggested chron nomenclature will be used throughout the following discussion.

Biostratigraphic correlations — the Cenozoic and Mesozoic

The correlation of magnetic and biostratigraphic time-scales have been the object of great interest since the recognition of the Cenozoic polarity time-scale. Proper correlations of

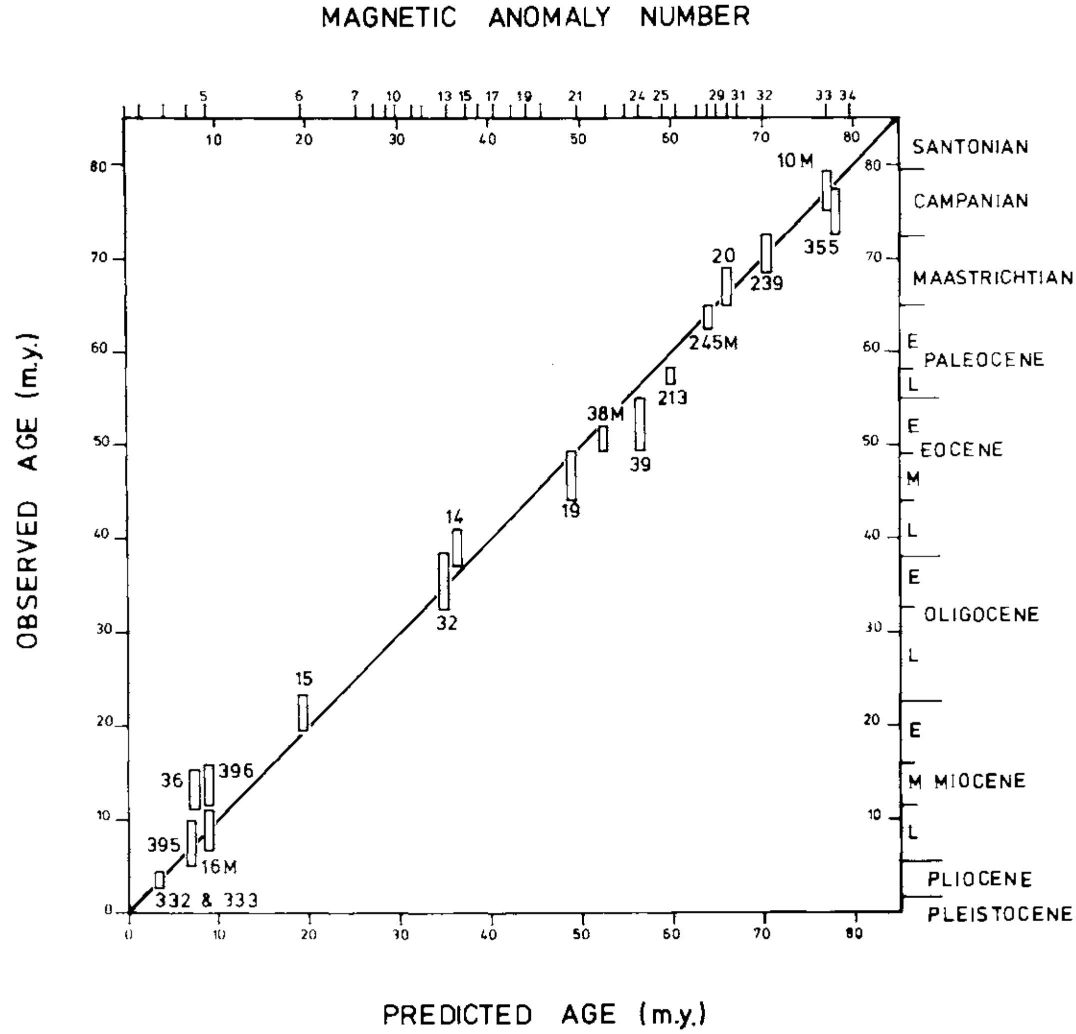

FIG. 5. DSDP bottom hole correlations from La Brecque *et al.* 1977.

the magnetic and biostratigraphic time-scales is important to the interrelationship of plate tectonic events and palaeoenvironmental phenomena. The evolution of oceanic circulation patterns and the attendant climatic development are presumably related to the development of ocean basins which is measured in terms of the geomagnetic reversal pattern (e.g. Pitman & Talwani 1972). Secondly, most reliable radiometric dates are related to the biostratigraphic time-scale, therefore the calibration of the geomagnetic reversal history to radiometric ages is accomplished through biostratigraphic correlations (LaBrecque *et al.* 1977; Ness *et al.* 1980; Lowrie & Alvarez 1981)

Correlations between magnetostratigraphic and biostratigraphic chronologies may be accomplished through magnetostratigraphic studies, e.g. Lowrie *et al.* (1980), Butler & Coney (1981), Lowrie *et al.* (1982), Tauxe *et al.* (in press), the dating of sediments in contact with ocean basement magnetized during a given magnetic chron (Maxwell *et al.* 1970) and the interpolation/extrapolation between known calibration points. Of these techniques, only the first is now precise enough to improve our knowledge of the correlation scheme for the Cenozoic and Mesozoic.

The most extensive magnetostratigraphic record yet developed is from the Gubbio region of the Umbrian Alps (Lowrie & Alvarez 1981). Correlations of the planktonic foraminiferal zones to the magnetostratigraphic record have been carried out for this zone and accepted radiometric ages for the biostratigraphic boundaries have heen applied to the chron boundaries. The development of the hydraulic piston corer has also permitted the Glomar Challenger to acquire magnetostratigraphic data for a good deal of the Cenozoic and late Cretaceous. The ocean floor biostratigraphic record is generally better preserved than the land record due to the less severe lithification of oceanic sediments. The seafloor sedimentary data allow the correlation of the high resolution nanoplankton zonations to the magnetostratigraphy which enhances the resolution of magnetostratigraphic/biostratigraphic correlations.

The observed occurrence of nanoplankton zonations and related magnetic stratigraphy determined for the Cenozoic of the eastern South Atlantic (Tauxe *et al.* 1984; Poore *et al.* 1984; Percival 1984) and for the Umbrian sequences (Lowrie *et al.* 1982) are shown in Fig. 4 and detailed in Table 3. A comparison of these zonations to DSDP bottom hole ages for the same time period (Fig. 5) demonstrates the increased resolution of magnetostratigraphy correlations. *The systematic miscorrelation between time-scales near to the Paleocene–Eocene boundary is evident from both illustrations. This miscorrelation is due to errors in the absolute age assignments to either the magnetic polarity sequences or the biostratigraphic zonations.* Implicit in the magnetostratigraphic scale of Figs 4 and 5 is the assumption of constant spreading in the South Atlantic for the Cenozoic. *The biostratigraphic scale is determined for sparsely located radiometric ages (Hardenbol & Berggren 1978). Therefore, the determination of a corrected scale is not presently obvious.*

The magnetostratigraphic/biostratigraphic correlations of Lowrie & Alvarez (1981) are displayed in Fig. 4 *without regard to their absolute ages*. As indicated in Lowrie & Alvarez (1981), the revised scale does correct the major systematic miscorrelation of the Paleocene. However, there does remain some systematic discrepancy within the Eocene. These systematic discrepancies should be removed after the acquisition of additional magnetostratigraphic data.

The assumption of constant sea-floor spreading in the South Atlantic is implicit in the earlier magnetostratigraphic time-scales (Heirtzler *et al.* 1968; LaBrecque *et al.* 1977) because the ages of the various Cenozoic and late Cretaceous anomalies were derived by simple extrapolation of Pliocene spreading rates. Later time-scales (Ness *et al.* 1980; Lowrie & Alvarez 1981) have assumed that the biostratigraphic ages are a more accurate measure of absolute ages and have recalibrated the magnetic reversal ages to adjust the mid-Paleogene miscorrelations.

We shall investigate the validity of the assumption of a constant spreading rate in the South Atlantic. The relative motion of two plates (A and B) may be described by a pole of relative rotation and an angular velocity about that pole for a given time interval (T_2-T_1). The spreading rate at a given point P on plate A is determined by the angular distance from P to the appropriate pole of rotation or:

V_P $= W_{AB}\, R_E \sin\theta$
W_{AB} = relative rotation rate between plates A and B
R_E = Earth's radius
θ = Angular distance between point P and the pole of rotation.

Deviations in either W_{AB} or θ_P will result in variations of the spreading rate at point P and hence departures from the assumption of constant spreading rate in the Heirtzler *et al.* (1968) model for the South Atlantic. In Fig. 4 we have plotted the variation in θ_P for points located along the Vema 20 profile of the South Atlantic.

It is interesting to note that the Vema 20 profile has remained at an angular distance of 90° ± 15° for most of the Cenozoic. Since the rate of opening of a point P at a ridge segment varies as sin θ_P, the 30°S profile of Vema 20 is at minimum sensitivity to variations in the pole rotation during the Cenozoic. Figure 6 shows a plot of the expected spreading

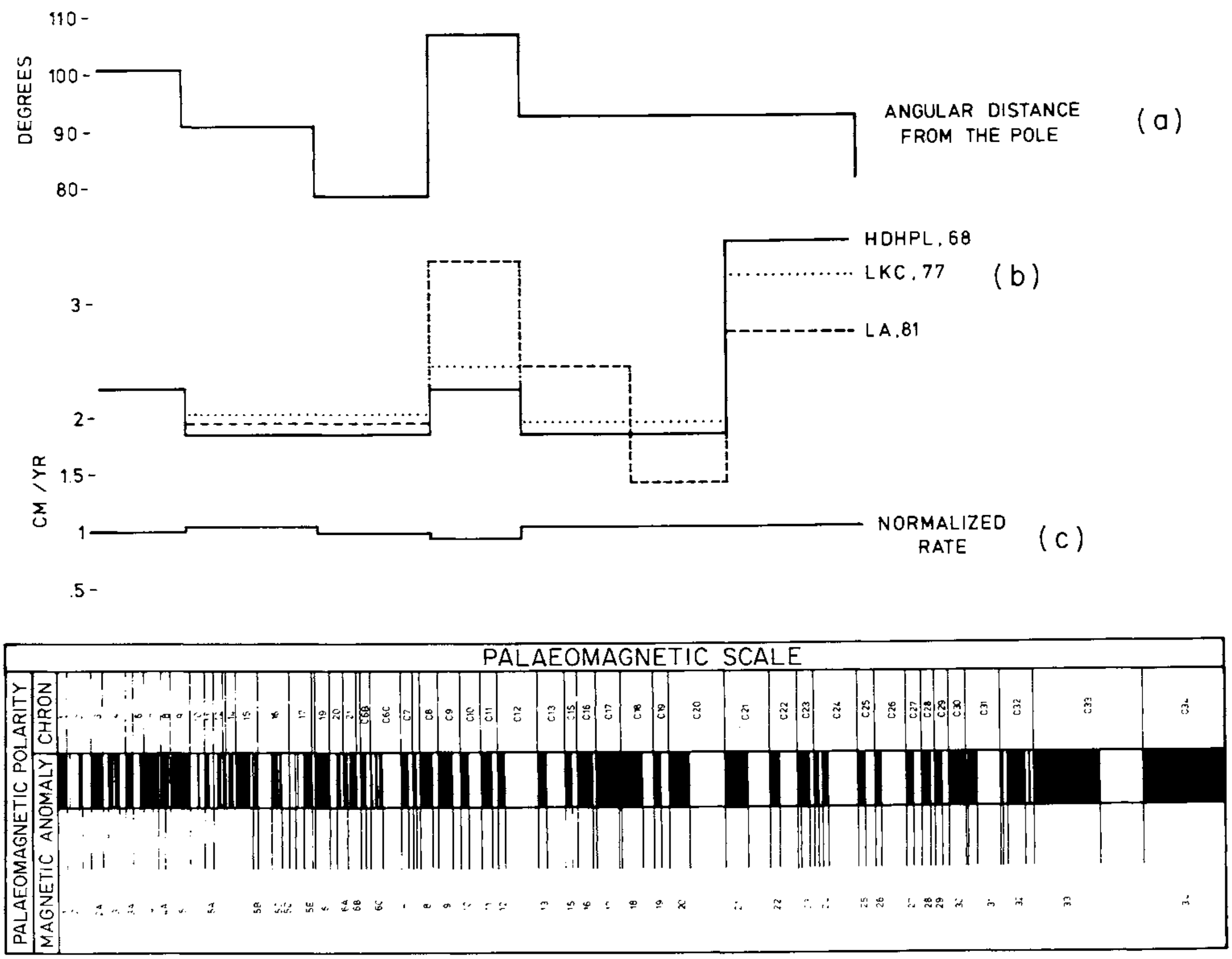

FIG. 6. (a) Variation in the location finite difference pole of rotation with respect to anomalies at 30°S. (b) Plot of spreading rate predicted for the South American, for the Cenozoic and late Cretaceous at 30°S. Rates are determined from poles of rotation calculated from Ladd (1974) and this article. Intervals are 0-C5, C5-C6B, C6B-C13, C13-C24, C24-C31, C31-C34. (c) Variations in spreading rates predicted by the time-scales of Heirtzler *et al.* (1968), La Brecque *et al.* (1977) and Lowrie & Alvarez (1981).

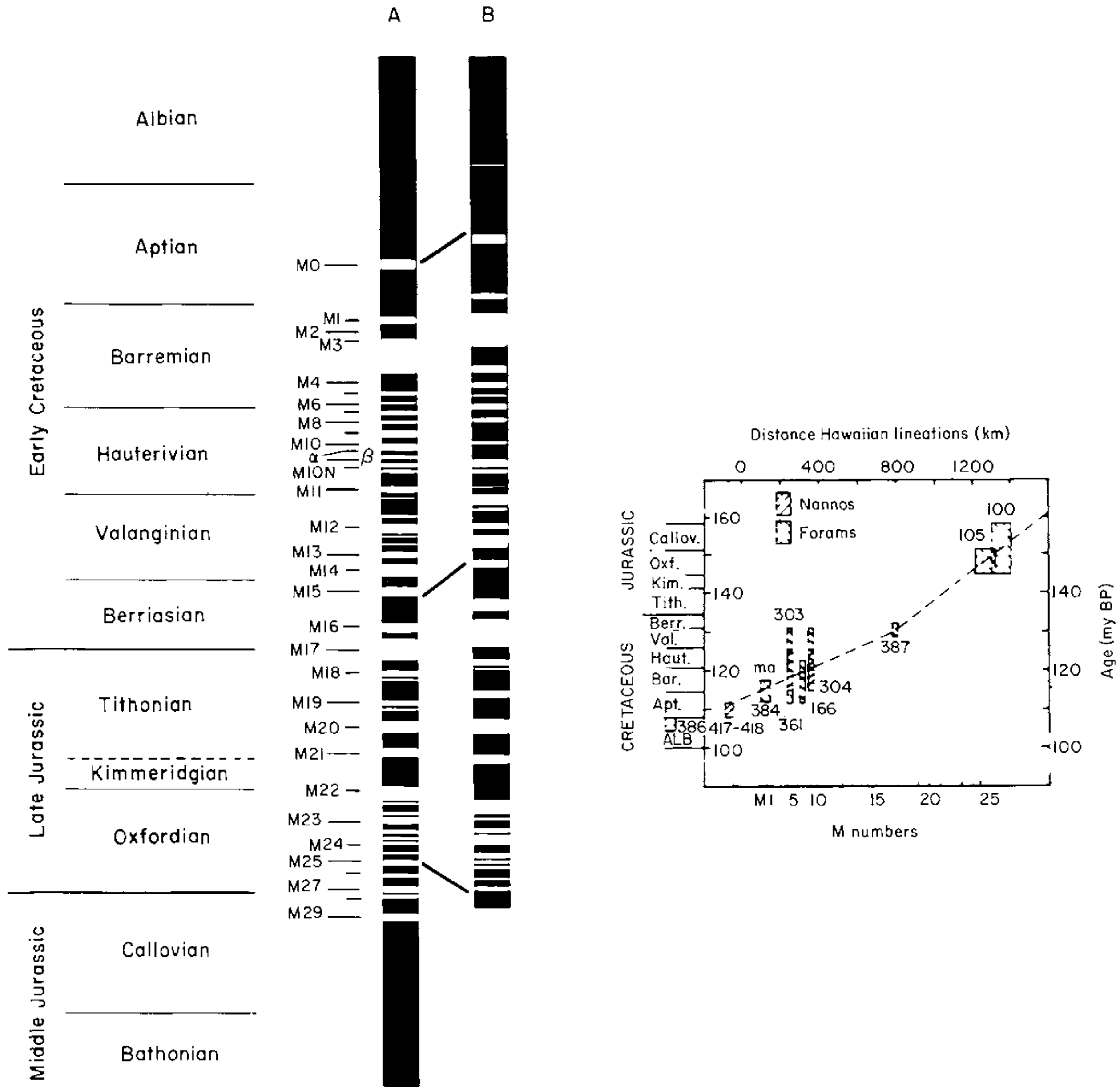

FIG. 7. Mesozoic polarity time-scales adapted from (a) Larson *et al.* (1981) and (b) Vogt & Einwich (1979). The DSDP bottom hole correlations are from Vogt & Einwich (1979).

rate assuming W_{AB} is constant and the pole of opening for Africa-South America is allowed to vary as observed by matching the anomalies of the African and South American plates (Ladd 1974; this chapter).

Figure 6 displays spreading rates normalized to the observed post Miocene rates. Note that the changes in pole of rotation result in less than a 4% variation in spreading rate for the entire Cenozoic. Variations in angular velocity of course may vary independently and we have no known reliable measure of this variable in the sea-floor data.

Also shown in Fig. 6 are the spreading rates predicted by the Heirtzler *et al.* (1968), LaBrecque *et al.* (1977), and Lowrie & Alvarez (1981) time-scales. It is interesting that the Heirtzler time-scale is not constant for the Cenozoic as was assumed in the construction but varies by 10% over the time-scale. This presumably is an artifact of local variations in spreading rate within the V-20 profile. This error, however, is not reflected in the magnetostratigraphic correlations. In fact the biostratigraphic correlations require an even greater variation in rate as demonstrated by the rates predicted by the Lowrie & Alvarez (1981) time-scale.

The authors therefore conclude that the observed systematic miscorrelations in the Paleogene between magnetostratigraphic and biostratigraphic scales cannot be explained by variations in the South Atlantic poles of rotation.

The Mesozoic reversal pattern is described by Larson & Hilde (1975) and Cande & Mutter (1978) – and tabulated in Table 3. The magnetostratigraphic to biostratigraphic correlations given in Fig. 7 we derived from the studies of Lowrie *et al.* (1980) and the DSDP bottom hole ages (Vogt & Einwich 1979). The assignment of radiometric ages is more difficult. The most accurate time-scale to date appears to be the scale of Larson *et al.* (1981) because it includes the M25-29 reversal pattern and recent DSDP drilling results.

Detailed chron structure

A second aspect of the geomagnetic reversal pattern is the development of detailed reversal history at the chron level. This knowledge is developed solely from the geomagnetic reversal pattern as recorded by the sea-floor spreading pattern. A caveat must be stated here since non-uniformities in the spreading history including asymmetric spreading, spreading centre jumps, changes in poles of rotation all contribute to the possible errors which may distort the recorded pattern. LKC-77 incorporated all revisions as of 1977. Subsequently, chrons C11, C12 and C19, C25, C26, C27 have been examined and revised (Cande & Mutter 1982; this paper). These have been included in Table 3. The technique is simply to assume local variations in spreading rate are small over chronal intervals (approximately 1–5 Ma) and that averaging over many profiles will compensate for remaining errors. A proper definition of the reversal pattern is crucial to proper magnetostratigraphic correlations and the determination of rates of chronal resolution levels. Studies of geomagnetic field histories are also affected to a lesser degree by errors in the detailed chronal structure.

J. L. LaB.

A summary of magnetic stratigraphy investigations for the Paleogene, Cretaceous and Jurassic

The sequence of geomagnetic polarity reversals interpreted from oceanic magnetic anomalies for the Jurassic, Cretaceous and Paleogene has been investigated independently by magnetic stratigraphy studies in sediment cores from the DSDP and in pelagic limestone sequences exposed on land. A major impediment to successful investigations in DSDP cores has been the frequently poor core recovery which resulted in unfortunately large hiatuses in the sediment record. The drilling process also introduced undesirable disturbance of the magnetization.

The geomagnetic field has only two stable polarity states and a transition from one to another has in itself no diagnostic character. Successful magnetostratigraphic correlation depends on the identification of a sequence of polarity intervals of different durations which form a recognizable pattern, or characteristic 'fingerprint'. Sedimentary hiatuses, changes in sedimentation rates, or non-representative sampling (as occasioned by poor DSDP core recovery) can spoil the pattern of polarity epochs. This has been, until recently, a drawback of magnetic stratigraphy studies on drilled DSDP cores. In spite of this problem Hailwood *et al.* (1979) correlated the Neogene magnetic stratigraphy in North Atlantic DSDP cores with the magnetic polarity time-scale for the last 10 Ma and produced a revised polarity time-scale for the whole of the Cenozoic era. Keating & Helsey (1978) also confirmed the main features of geomagnetic polarity history during the Cretaceous in DSDP

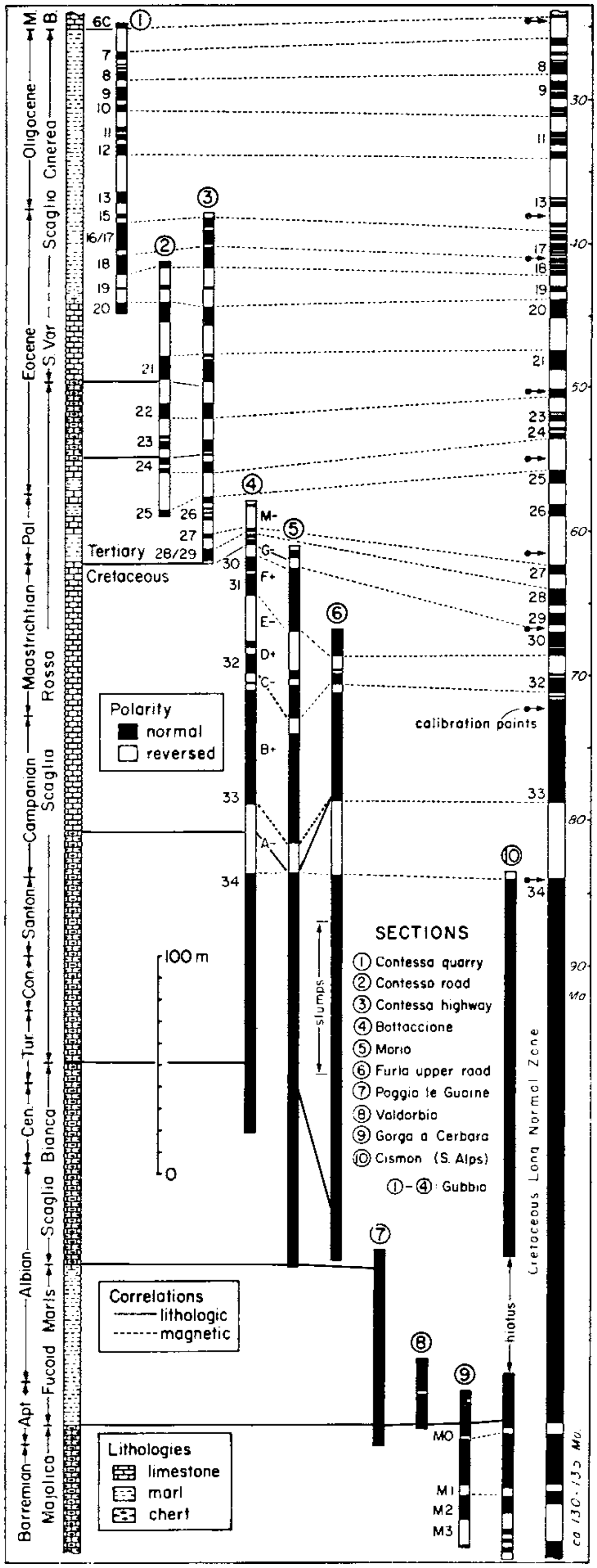

FIG. 8. Paleogene and Cretaceous magnetostratigraphic results from 10 sections in Umbria and the Southern Alps. Left-hand columns: biostratigraphic age (M. = Miocene), formation names (B. = Bisciaro, S. Var. = Scaglia Variegata) and lithology. *Right-hand column: oceanic magnetic reversal sequences as redated by Lowrie & Alvarez (1981).* Numbers refer to standard magnetic anomaly identification. (Figure from Lowrie & Alvarez 1981).

cores. The recent introduction of a hydraulic piston corer on DSDP Legs has resulted in improved recovery, and undisturbed magnetizations. This has enabled Tauxe *et al.* (1980, 1984) to confirm the Paleogene and late Cretaceous magnetic polarity history in marine sediments from DSDP Leg 73.

Magnetic stratigraphy in pelagic limestone sections

In order to preserve a recognizable 'fingerprint' reversal pattern a sedimentary sequence must be deposited continuously and reasonably uniformly over periods of several million years duration. The natural remanent magnetization, aside from easily removable 'soft' secondary components, must date from close to the time of deposition of the sediment. These conditions are met by the Mesozoic and Paleogene pelagic limestones of South Tethyan provenance which are now widespread throughout the Mediterreanean realm. Their post-compactional sedimentation rates are typically around 10 m/Ma, and a sampling interval of 0.5–1.0 m gives a resolution of about 50–100 Ka. This is approximately equivalent to the time represented in the thinnest strip of magnetized oceanic crust, the magnetic anomaly of which can be resolved at the oceanic surface (Cox 1969). As a result of their suitable properties the South Tethyan limestones have been intensively investigated in recent years. Magnetic stratigraphies have been closely tied to palaeontological dating schemes in at least 22 sections (Figs 8, 9, 10). The oldest (early Jurassic) sections pre-date the onset of sea-floor spreading following the breakup of Pangaea; for them there is no corresponding oceanic reversal sequence. The late Jurassic, Cretaceous and Paleogene sections provide important correlations for dating the oceanic magnetic anomaly sequences.

Paleogene and late Cretaceous magnetic stratigraphy

Thick exposures of the pelagic carbonate rocks of the Umbrian sequence have been intensively investigated in the Contessa and Bottaccione gorges near to Gubbio in the northern Apennines of Italy. The magnetic stratigraphies of these sections give an almost one-to-one correlation with the Paleogene magnetic reversal sequence determined from sea-floor magnetic anomalies (Lowrie *et al.* 1982; Napoleone *et al.* 1983). The Contessa section has been dated by zonations of planktonic foraminifera and calcareous nanofossils; the Bottaccione section was dated by zoning the planktontic foraminifera. Important locations of the Paleogene biostratigraphic stage boundaries have been determined in these sections (Fig. 8). The Oligocene–Miocene boundary was found to be within a negative polarity interval and just older than anomaly 6C (Lowrie *et al.* 1982). Anomaly 14, which does not show on all oceanic magnetic profiles (LaBrecque *et al.* 1977), also had no expression in the Contessa section; the Eocene–Oligocene boundary was located in the negative polarity interval between anomalies 13 and 15 (Lowrie *et al.* 1982). The Paleocene–Eocene boundary was firmly established in the lower part of the long negative interval separating anomalies 24 and 25 (Lowrie *et al.* 1982; Napoleone *et al.* 1983). This key location agrees with the placement of the boundary in vertebrate-bearing continental sedimentary sequences in Wyoming (Butler & Coney 1981; Butler *et al.* 1981). It also agrees very closely with magnetostratigraphic correlations in DSDP cores from Leg 48 (Hailwood *et al.* 1979) and Leg 73 (Tauxe *et al.* 1984).

This placement of the Paleocene–Eocene boundary is substantially different from that in previous magnetic polarity time-scales; Heirtzler *et al.* (1968) placed it within anomaly 21, LaBrecque *et al.* (1977) placed it within anomaly 23 and Ness *et al.* (1980) placed it within anomaly 24. However, previous locations of the boundary were not the result of direct magnetostratigraphic correlations but derived from interpolations between the ages allotted to anomalies in these successive magnetic polarity time-scales.

The Tertiary–Cretaceous boundary has now been established to occur within the negative polarity interval separating anomalies 29 and 30. This location was first established in the Gubbio Bottaccione section (Premoli Silva *et al.* 1974; Roggenthen & Napoleone 1977; Lowrie & Alvarez 1977a, 1977b) and has since been confirmed in other Umbrian sections at Moria (Alvarez & Lowrie 1978), at Furlo (Alvarez & Lowrie 1984) and in the Gubbio Contessa section (Lowrie *et al.* 1982). It has been found in the same location in sections in the Southern Alps (Channell & Medizza, 1981) and in DSDP sediments (Tauxe *et al.* 1984). The boundary has been located within a normal polarity zone corresponding to anomaly 29 in vertebrate-bearing sections from the SW United States (Butler *et al.* 1977). The discrepancy may be due to misadjustment of the vertebrate and microfossil biostratigraphic schemes (Butler *et al.* 1977), or to the presence of a hiatus at this level (Alvarez & Vann 1979).

The magnetic stratigraphy of the Umbrian sections clearly matches the oceanic reversal sequence not only in the Paleogene but also in the late Cretaceous (Fig. 8). In the Cretaceous it is dominated by a long period of normal polarity correlated with the Cretaceous magnetic quiet zone in the oceanic magnetic anomaly sequence (Fig. 8). This quiet zone is therefore interpreted as due to sea-floor spreading during a lengthy interval of constant, normal geomagnetic polarity, as interpreted by Helsley & Steiner (1969). This quiet interval terminated with the reversal which gives rise to marine magnetic anomaly 34; its age is slightly younger than the Santonian–Campanian boundary (Lowrie & Alvarez 1977a, 1977b; Channell *et al.* 1979; Channell & Medizza, 1981).

Early Cretaceous and late Jurassic magnetic stratigraphy

Early Cretaceous magnetostratigraphic sections at Cismon in the Southern Alps (Channell *et al.* 1979) and at Poggio le Guaine, Valdorbia and Gorgo a Cerbara in Umbria (Lowrie *et al.* 1980) document that the Cretaceous long normal interval began in the Early Aptian and lasted through the Albian, Cenomanian, Turonian, Coniacian and Santonian stages (Figs 8, 9). There have been several reports of short-duration reversed epochs within the quiet interval (Keating & Helsley 1978; Hailwood *et al.* 1979; VandenBerg & Wonders 1980; Lowrie *et al.* 1980) but these have not yet been correlated between different magnetostratigraphic sections and have no consistent expression in the oceanic magnetic anomaly record. This absence of a global character makes it questionable whether they reflect real geomagnetic reversals.

The youngest anomaly (M0) of the M-sequence of marine magnetic anomalies (Larson & Hilde 1975) is just younger than the Aptian–Barremian boundary (Figs 8, 9). The 'fingerprint' reversal pattern for anomalies M0 to M4 can be correlated adequately with magnetostratigraphic results. For

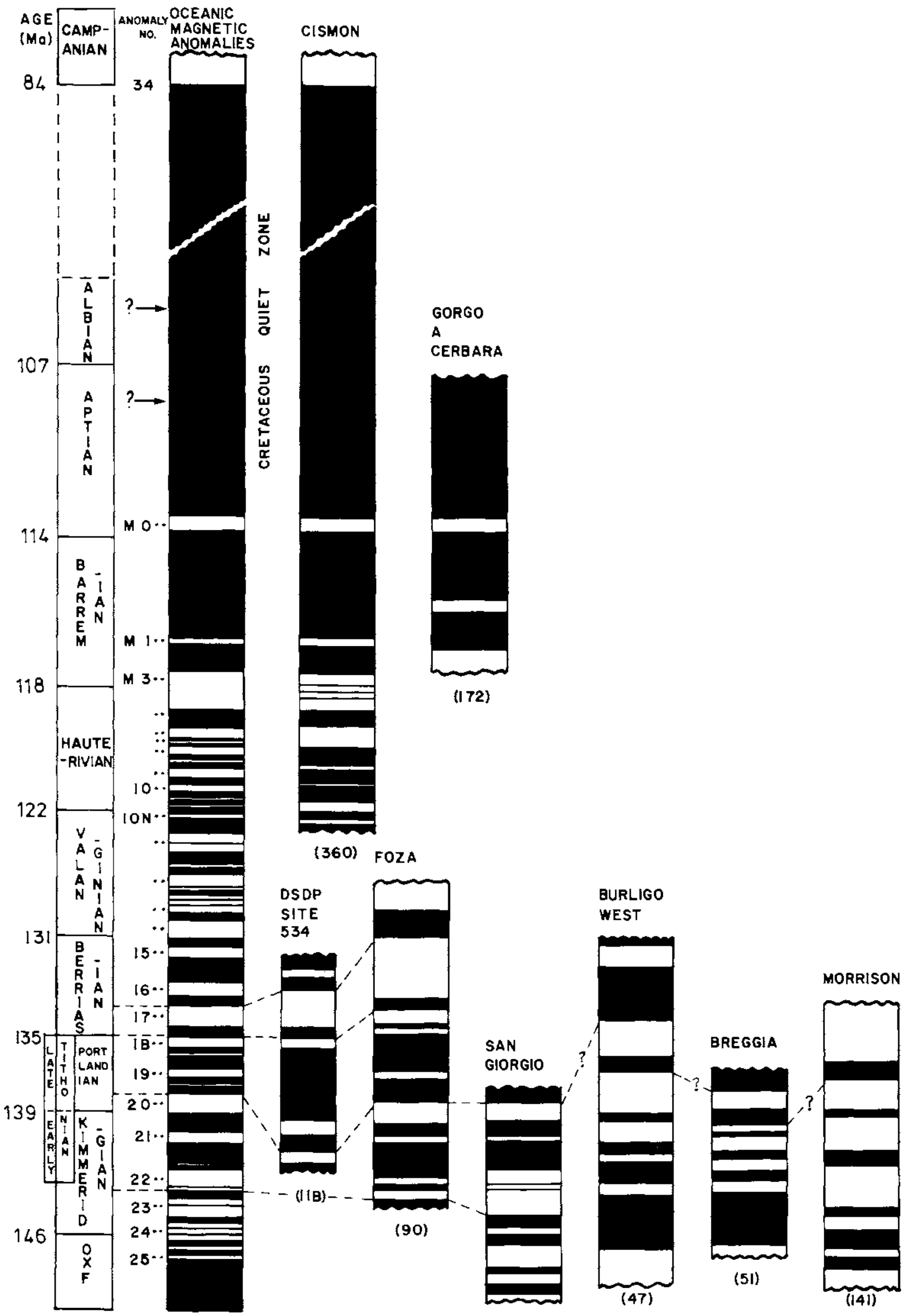

FIG. 9. Early Cretaceous and late Jurassic magnetostratigraphic sections and their correlation with the geomagnetic polarity interpreted from M-sequence oceanic magnetic anomalies (Larson & Hilde 1975). *The absolute ages associated with stage boundaries are adopted uncorrected from Van Hinte (1976a, b).* The six polarity stratigraphies at the base of the figure have been expanded for clarity. Dashed tie lines are primarily biostratigraphic and secondarily magnetostratigraphic, and correlate the magnetozones to the oceanic sequences. (From Channell *et al.* 1982.)

older M-sequence reversals there is not an unambiguous correlation with the magnetostratigraphic record in continental outcrops or DSDP cores. However, the Hauterivian and Valanginian stages are characterized by mixed polarity magnetozones corresponding to oceanic magnetic anomalies M5 to M15. New results from DSDP site 534 (Ogg 1983) and from sections in the southern Alps at Foza and San Giorgio (Ogg 1981) corroborate the anomaly reversal sequence from M16 to M22 (Fig. 9). This locates the position of the Cretaceous–Jurassic boundary (indicated by a sudden increase in abundance of the nanofossil *Nanoconus coloni*) near to the reversal boundary which marks the young end of anomaly M18.

These direct correlations and age determinations for anomalies M0 to M4 and M16 to M22 have important consequences for the dating of M-sequence anomalies. Previously, these were dated by determining the biostratigraphic ages of the oldest sediments overlying basaltic basement at DSDP sites drilled on known anomalies (Larson & Pitman 1972; Larson & Hilde 1975; Vogt & Einwich 1979). These age estimates are imprecise because (1) nanofossil and foraminiferal ranges are particularly broad for the early Cretaceous and late Jurassic, (2) the ranges of the observed nanofossils and forams are often incompatible, and (3) the ages are minimum ages only because some time may elapse before the basalt layer acquires a permanent sediment blanket. In the time-scale of Larson & Hilde (1975) a middle Aptian age was associated with anomaly M0, the Cretaceous–Jurassic boundary was associated with anomaly M16 and a middle Tithonian age was associated with anomaly M18. The revised dating on the basis of continental magnetostratigraphic investigations allocates the anomaly sequence M0-M17 to the early Cretaceous Barremian, Hauterivian, Valanginian and Berriasian stages, requiring an increase in previous estimates of early Cretaceous sea-floor spreading rates.

The correlation of other late Jurassic magnetic anomalies older than M22 is less satisfactory. It is difficult to make one-to-one correlations between the Southern Alpine sections at Breggia and Burligo West (Ogg 1981), the Morrison formation in the western United States (Steiner 1980) and the oceanic reversal pattern (Fig. 9). The Tithonian and Kimmeridgian stages are characterized by well-defined positive and negative magnetozones, corresponding to the oldest of the M-sequence anomalies (the Kimmeridgian stage as used here employs the British convention in which the Upper Kimmeridgian is equivalent to the Lower Tithonian, rather than the French convention where the Kimmeridgian stage separates the Tithonian and Oxfordian stages).

The oceanic anomaly sequence is well established to anomaly M25 (Larson & Hilde 1975) and less well-defined older lineations as far as M29 have been postulated (Cande *et al.* 1978). Oceanic crust older than the M-sequence anomalies is characterized by the Jurassic Quiet Zone in which lineated anomalies are absent. The explanation of this quiet zone is not certain. Steiner (1980) concluded that North American sedimentary sections do not support the existence of an extended period of normal polarity in the Jurassic, but Heller (1977, 1978) found that the Malm α, β and γ (equivalent to Oxfordian and early Kimmeridgian) stages in marine limestones from southern Germany are normally magnetized. Magnetostratigraphic results from sections at Valdorbia and Fonte Avellana in the Umbrian Apennines (Fig. 10) show that most of the early Kimmeridgian, the Oxfordian and Callovian stages are represented by normal magnetic polarity (Channell *et al.* 1982). Unfortunately biostratigraphic control in these sections is poor and the stage boundaries cannot be unambiguously located. However, this interval of normal polarity appears to he largely restricted to the Oxfordian and Callovian stages.

The Bathonian to Sinemurian stages of the Middle and early Jurassic have been investigated in several magnetostratigraphic sections (Fig. 10). There is no corresponding oceanic magnetic record because sea-floor spreading following the break-up the Pangaea had not been initiated or had not yet generated a typical oceanic crust. However, mixed polarities characterize the Bathonian stage in the Erbezzo South section in the southern Alps (Ogg 1981) and set the older limit for the Jurassic quiet interval. Part of the Bajocian has not yet been investigated in magnetostratigraphic sections. Overlapping sections at Fonte Avellana, Valdorbia and Cingoli in Umbria (Channell *et al.* in press), at Breggia in the southern Alps (Horner & Heller 1981) and at Bakonycsernye in Hungary (Marton *et al.* 1980) show magnetozones of alternating magnetic polarity. The Pliensbachian section at Bakonycsernye can be correlated with the corresponding part of the Breggia section; both have well-developed magnetic stratigraphies and good ammonite biostratigraphies. It is not yet possible to correlate clearly the reversal stratigraphies in the other sections with each other.

W. L., J. E. T. C.

Some geological implications of the revised time-scale

The revised Jurassic to Paleogene time-scale is presented in summary form in Table 4, utilizing data presented in both parts of this report. Because late Cretaceous events have aroused considerable interest in recent years and because they can be dated with comparatively high precision some estimates are presented here of geologically significant rates of change.

Late Cretaceous events

The late Cretaceous sedimentary trough of the United States Western Interior is one of the most fully researched regions in the world. Sedimentation rates varied from a maximum of 155 m/Ma in Utah and Wyoming to a minimum of 17 m/Ma in North Dakota. This may be compared with a mere 3.6 m/Ma for the Cenomanian Lower Chalk of north Norfolk, England.

Gill & Cobban (1973) calculated rates of advance and retreat of the strandline at zonal level in Montana and Wyoming on the western border of the Western Interior seaway. Transgressions during the Campanian–Maastrichtian ranged over 95 to 150 km per Ma. In general regressions were 55 to 115 km per Ma but for half a million years the later Fox Hills regression was moving at the rate of more than 800 km per Ma.

With regard to rates of eustatic sea-level change, there are considerable difficulties in assigning absolute figures to past sea levels (for discussions, see Hays & Pitman 1973; Bond 1978; Hancock & Kauffman 1979; Watts & Steckler 1979; Schlanger *et al.* 1981). The minimum estimate for the overall rise of sea-level from the beginning of the Albian is 100 m (Schlanger *et al.* 1981), the maximum 650 m (Hancock & Kauffman 1979). Revision of the very crude figures given in Fig. 5 of Hancock & Kauffman gives the following values.

1. Transgression from early Albian to earliest Turonian: 4.3 m per Ma if the total Cretaceous rise was only 100 m; 28 m per Ma if the total rise was 650 m.
2. Regression from Early Turonian to early late Turonian: 25 m/Ma (for 100 m); 160 m/Ma (650 m).
3. Transgression from late Santonian to late Campanian: 1.5 m/Ma (100 m); 10 m/Ma (650 m).

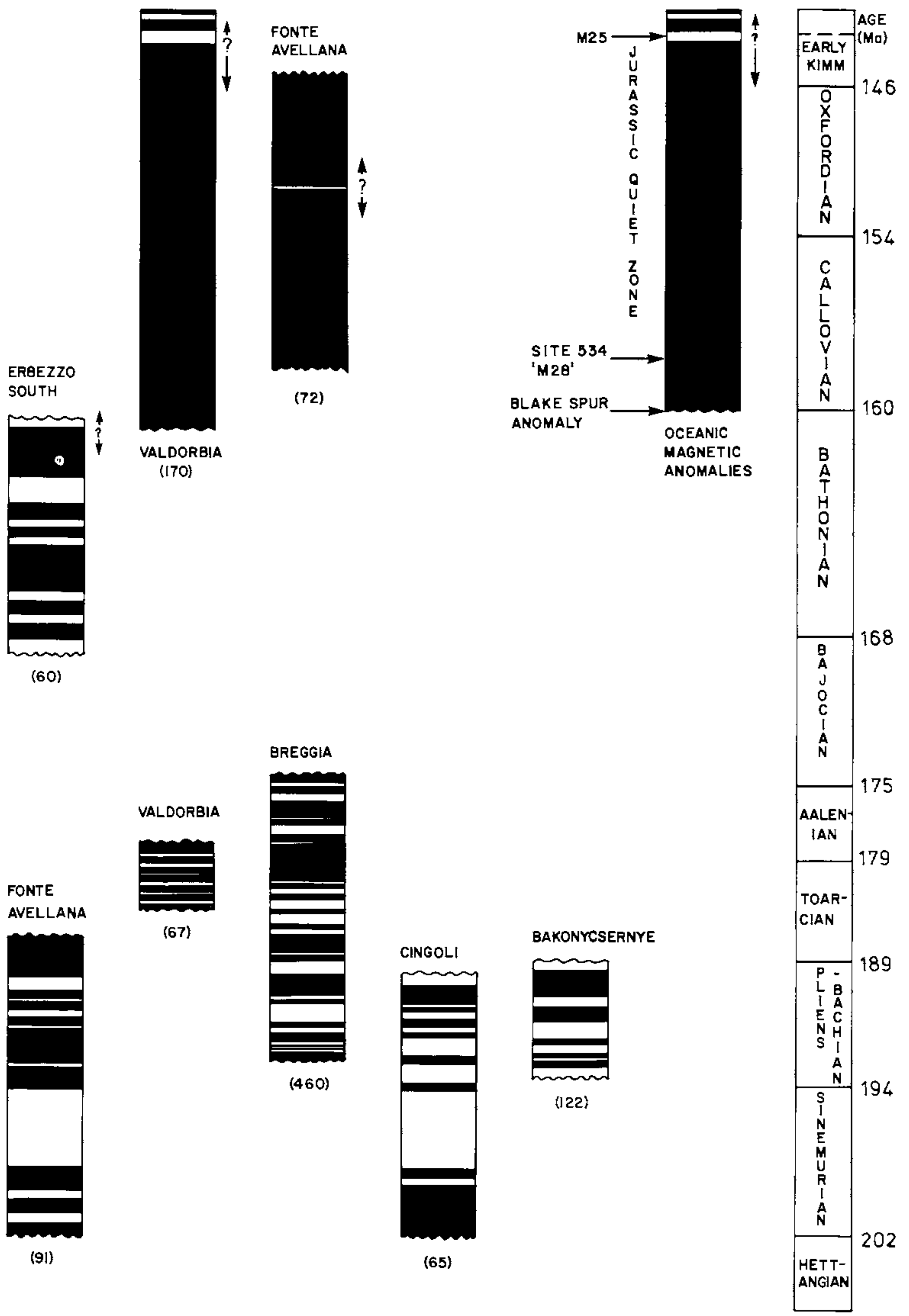

FIG. 10. Middle and early Jurassic magnetostratigraphic sections confirm an extended period of normal polarity in the Oxfordian and Callovian, and mixed polarity in the early Jurassic stages (from Channell *et al.* 1982). Ages of Jurassic stages are after Van Hinte (1976a).

4. End Maastrichtian regression is very difficult to quantify, but possibly 25 m/Ma (100 m); 170 m/Ma (650 m).

Late Cretaceous ammonite history

The first Hoplitaceae, prominent superfamily in the boreal Albian, appeared in the Valanginian (if the Eodesmoceratidae are primitive hoplitacids (Rawson 1981)) but according to Casey (1965) and Wright (1981) the earliest is a *Cleoniceras* in the middle *tardefurcate* Zone of Germany at approximately 107 Ma. The last, *Hoplitoplacenticeras*, just survived into the Maastrichtian, but is poorly dated biostratigraphically at 70 ± 2 Ma giving a total span approximately 37 Ma.

TABLE 4. Summary of revised dates for Jurassic–Paleogene time interval, to nearest million years

		24
	Chattian	
OLIGOCENE		30
	Rupelian	
		37
	Priabonian	
		40
	Bartonian	
EOCENE		44
	Lutetian	
		52
	Yypresian	
		58
	Selandian	
PALEOCENE		64
	Danian	
		66
	Maastrichtian	
		72
	Campanian	
		84
	Santonian	
		88
	Coniacian	
		90
	Turonian	
		91
	Cenomanian	
CRETACEOUS		95
	Albian	
		107
	Aptian	
		114
	Barremian	
		118
	Hauterivian	
		122
	Valanginian	
		131
	Berriasian	
		135
	Portlandian/Volgian	
		139
	Kimmeridgian	
		146
	Oxfordian	
		154
	Callovian	
		160
	Bathonian	
JURASSIC		168
	Bajocian	
		175
	Aalenian	
		179
	Toarcian	
		187
	Pliensbachian	
		194
	Sinemurian	
		202
	Hettangian	
		205

Proleymeriella, the earliest of the vast superfamily Acanthocerataceae, appears at the base of the Albian (107 Ma). The last is *Sphenodiscus* which reaches the late Maastrichtian but not the top of the stage with certainty which is approximately 67 Ma giving a total span of 40 Ma.

Hamites, the first of the Turrilitaceae, appeared in the middle of the early Albian (Casey 1961) at approximately 104 Ma. The last *Baculites* and *Diplomoceras*, survived to the end of the Cretaceous, 66 Ma giving a total span of 38 Ma.

As regards the rate of extinction, Hancock (1967) and Wiedmann (1969) showed that there was a decline in variety amongst the ammonites that was spread over the whole of the late Cretaceous. New data, principally from Wright (1981), shows that there was a rather steady decrease in variety from the Albian with 33 families to the Santonian with 13. This is equivalent to percentage decreases in the number of families known from the whole Cretaceous (52) in the range of 1.9–3.3% per Ma. But only one family disappeared in the Santonian, and one more in the Campanian, equivalent to percentage decreases of only 0–2% Ma.

According to Wiedmann (1969) and Birkelund (1979) the highest Maastrichtian of Zumaya in northern Spain and in Denmark together contain seven, possibly eight families. It is impossible at present to determine the speed of this final disappearance of the ammonites, hut it must have taken less than 1 Ma.

The end-Cretaceous mass extinction event in general is fully discussed in part 2 of this report by Berggren *et al.*

J. M. H.

Sea-floor spreading and the disintegration of Pangaea

Recent drilling at site 534 of Leg 76 of the DSDP in the Atlantic east of Florida, has confirmed a proposal, hased primarily on biogeographic evidence, that true ocean floor did not start to he generated between North America and Africa until late Middle Jurassic times (Hallam 1980). This event, which marks the beginning of the disintegration of Pangaea, has been dated as approximately basal Callovian (Sheridan *et al.* 1982). This contrasts with the Toarcian opening time estimated by Pitman & Talwani (1972) on the basis of extrapolation back through time of sea-floor spreading rates inferred from younger Mesozoic magnetic anomalies.

The revised Atlantic opening data accounts for the otherwise puzzling lack of deep Atlantic magnetic anomalies during the early Jurassic to Bathonian interval of mixed polarity reported here by Lowrie & Channell, and implies a much more rapid rate of late Jurassic spreading than the ~ 1 cm/a inferred by Larson & Pitman (1972); Sheridan's revised figure is 3.76 cm/a.

Our revisions also require an increase over previous estimates of early Cretaceous spreading rates, such as ~ 1 cm/a in the Atlantic (Larson & Pitman 1972). This implies a corresponding reduction in late Cretaceous spreading rates, which has a bearing on the Hays & Pitman (1973) hypothesis tbat tbe late Cretaceous eustatic rise of sea-level was due to a rapid increase in the rate of sea-floor spreading. Hays & Pitman utilized the Larson & Pitman (1972) data, suggesting for example that the spreading rate in the North Atlantic increased from 1.1 to 3.3 cm/a from the early to the late Cretaceous. With the revised time-scale and new magnetic data these figures are altered respectively to ~ 1.6 and ~ 2.5 (see Baldwin *et al.* 1974, Fig. 1).

The much more modest rate of increase implied by our new data suggests that other factors might have played a more significant role in promoting the eustatic rise, such as an increase in the length of the ocean ridge system, as

Gondwana disintegration got fully underway (Hallam 1977, 1980), or the development of swells on the Pacific and Caribbean floors (Schlanger *et al.* 1981).

The following best age estimate in Ma of significant events in the breakup of Pangaea is based on Hallam (1980), Kristoffersen (1978), Norton & Sclater (1979), Sheridan (*et al.* 1982) and Talwani & Eldholm (1977), in conjunction with the revised chronology. The dates must obviously be regarded only as approximate.

1. Opening of Atlantic between North America and Africa; 160 Ma.
2. East Gondwana (Madagascar, India, Australia, Antarctica) started to separate from West Gondwana (Africa, South America); 148 Ma.
3. Spreading started in South Atlantic; 125–130 Ma.
4. Northwards extension of Atlantic between Ireland and Newfoundland; 90–95 Ma. Westward jump of N. Atlantic spreading centre and commencement of Labrador Sea opening; 80 Ma.
5. India moved away from Madagascar; 80–90 Ma.
6. Norwegian Sea commenced opening; 56 Ma.
7. Australia separated from Antarctica; 53 Ma.

A. H.

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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-

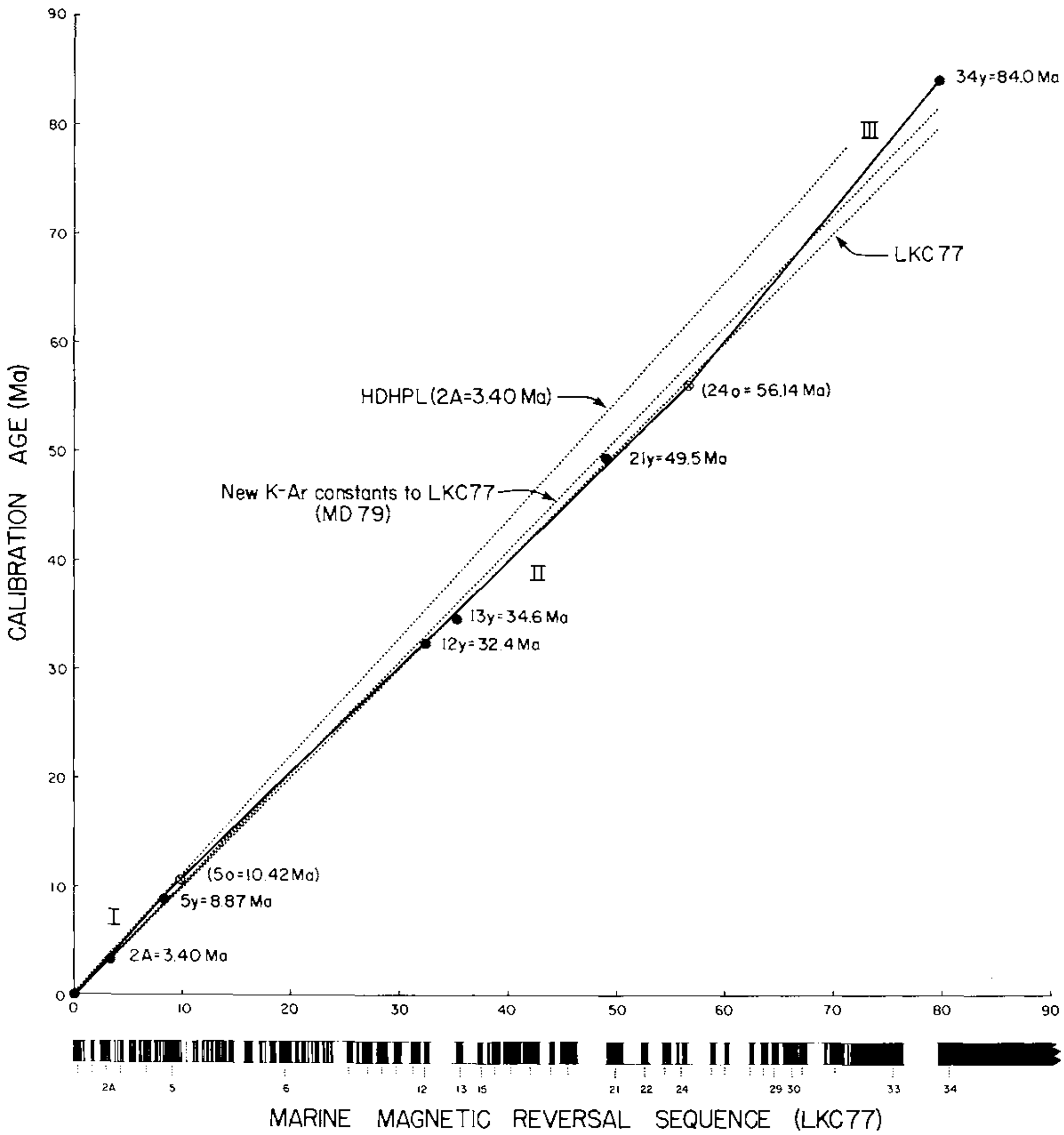

FIG. 1. Revised age calibration of marine magnetic reversal sequence from LaBrecque *et al.* 1977 (LKC77). Solid lines are three linear apparent age-calibration age segments (I, II, and III) which satisfy calibration tie-points indicated by solid circles (see text for details). The two open circles with 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 I. 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 hiochronologic 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-

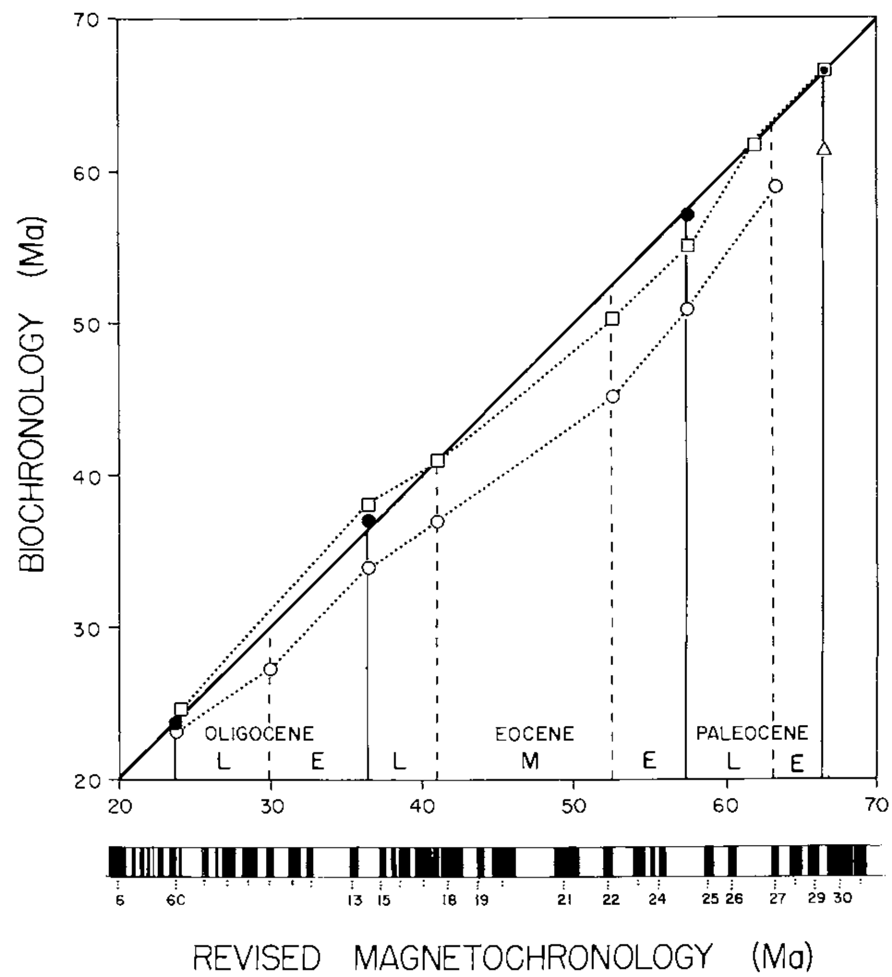

FIG. 2. Comparison of various biochronological estimates of Paleogene epoch and intra-epoch boundaries within magnetochronological framework provided by correlation to revised geomagnetic polarity time-scale. Solid circles: this paper. Open circles: from Odin & Curry (1981). Squares: recalculated from 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 biostratigraphic 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 bighest 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 magnetobiostratigraphic 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, Bishopstone 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. kuepperi* 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

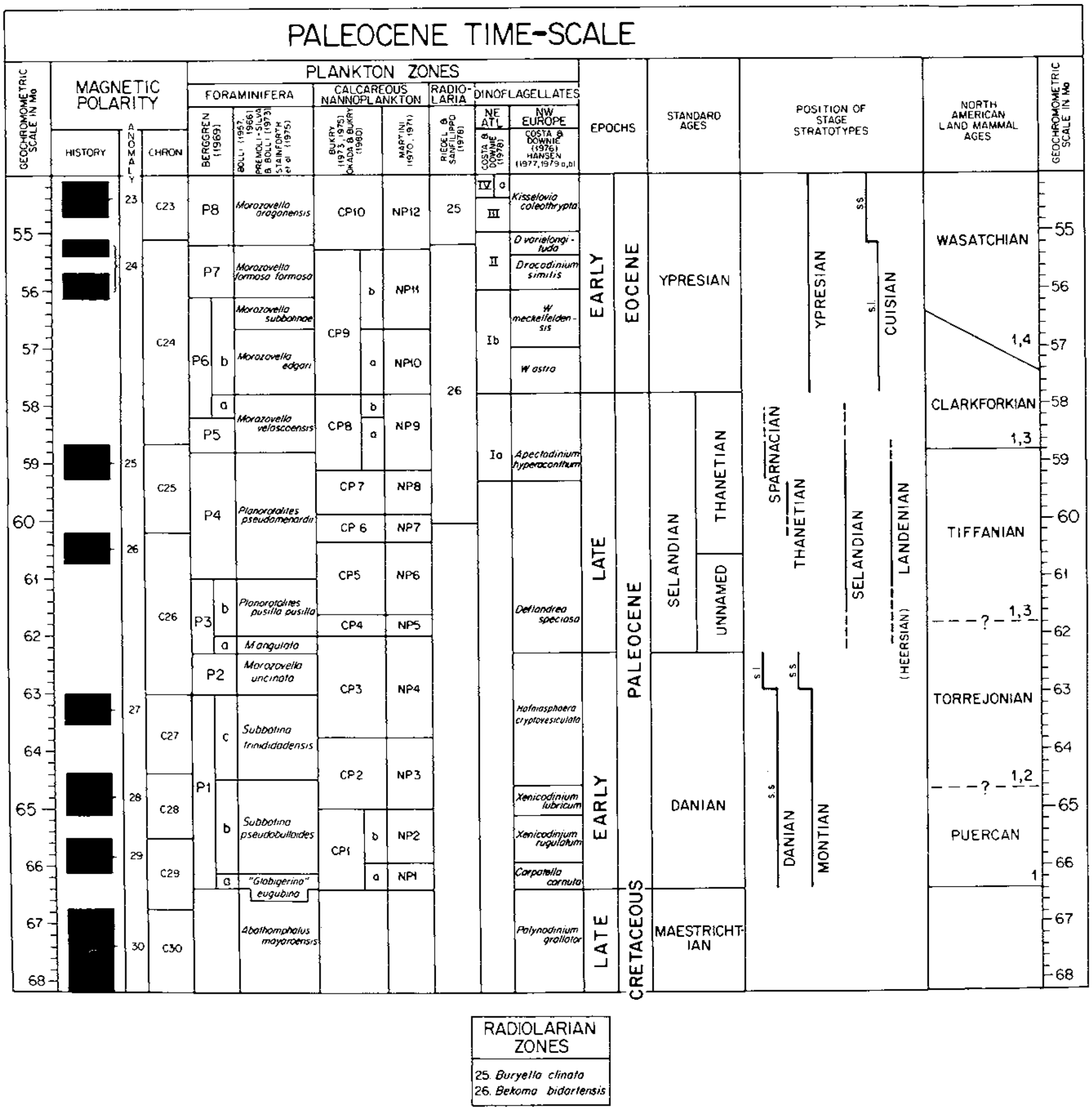

FIG. 3. Paleocene geochronology. The geochronologic scale at the margins of the figure is derived from the magnetic polarity chronology which is in turn derived from palaeontologically and/or palaeomagnetically controlled radiometrically dated calibration points in the late Neogene, early Oligocene, 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 magnetostratigraphic 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 hy 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, heen 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 desertilfraasi* 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* (Ial) 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 magnetostratigraphic 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 uncertainty' (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 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 houndary 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 (Cf_2) of Rose (1980), and therefore tbe Paleocene–Eocene boundary is located between lower and middle Clarkforkian (Cf_1/Cf_2) 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

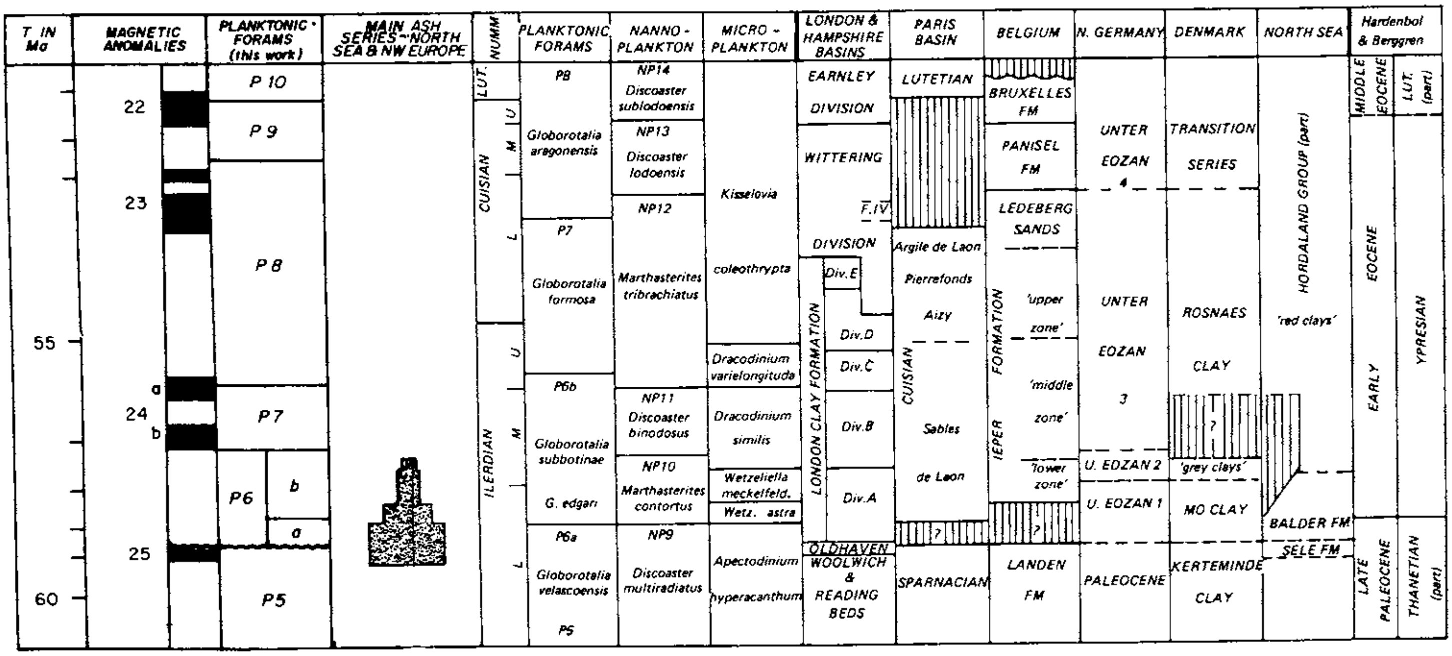

FIG. 4. Correlation of uppermost Paleocene and lowermost Eocene stratigraphic succession in northwest Europe (modified from King 1981, text-Fig. 52). In the left hand side of the figure we show the relationship between planktonic foraminiferal zones to the magnetobiochronologic scale developed in this 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-palaeooceanographic expression of the eustatic sea-level fall in deep sea (i.e. bathyal) deposits and trace this level into the unconformity 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 *Nannotetrina* sp. as low as the lower part of anomaly 21 correlative in the Contessa Road section (*op. cit.*) which suggested 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 sublodoensis* (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 *Amphimorphina 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 correlatives 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

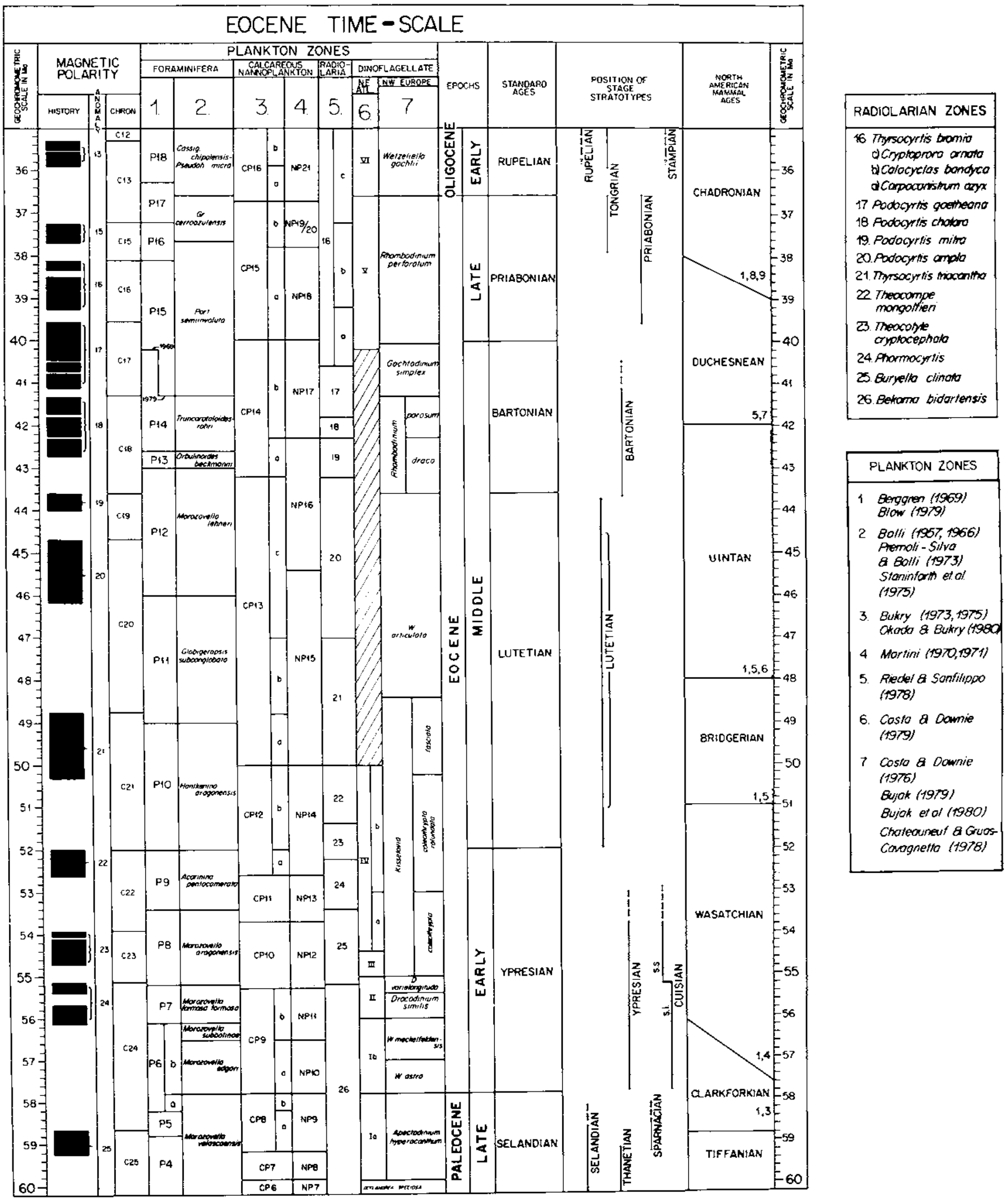

FIG. 5. Eocene geochronology (explanation as in Fig. 3).

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 hoth 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 hy 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 ohtained K-Ar (glauconite) dates of 37.6 Ma on the Pachuta Member (Jackson Formation), 37.9 Ma on the Sbubuta 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 dehris in the Wheatland Formation;
2. the assumed hiostratigraphic accuracy of a Refugian benthic foraminiferal fauna. Paleogene henthic 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 hiostratigraphic 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 microtektites 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 propiniquity.
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 metbods (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 sbort 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 hy 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 microtektite 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. Tbe corrélation 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 magnetochronologic 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 magnetochronology 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

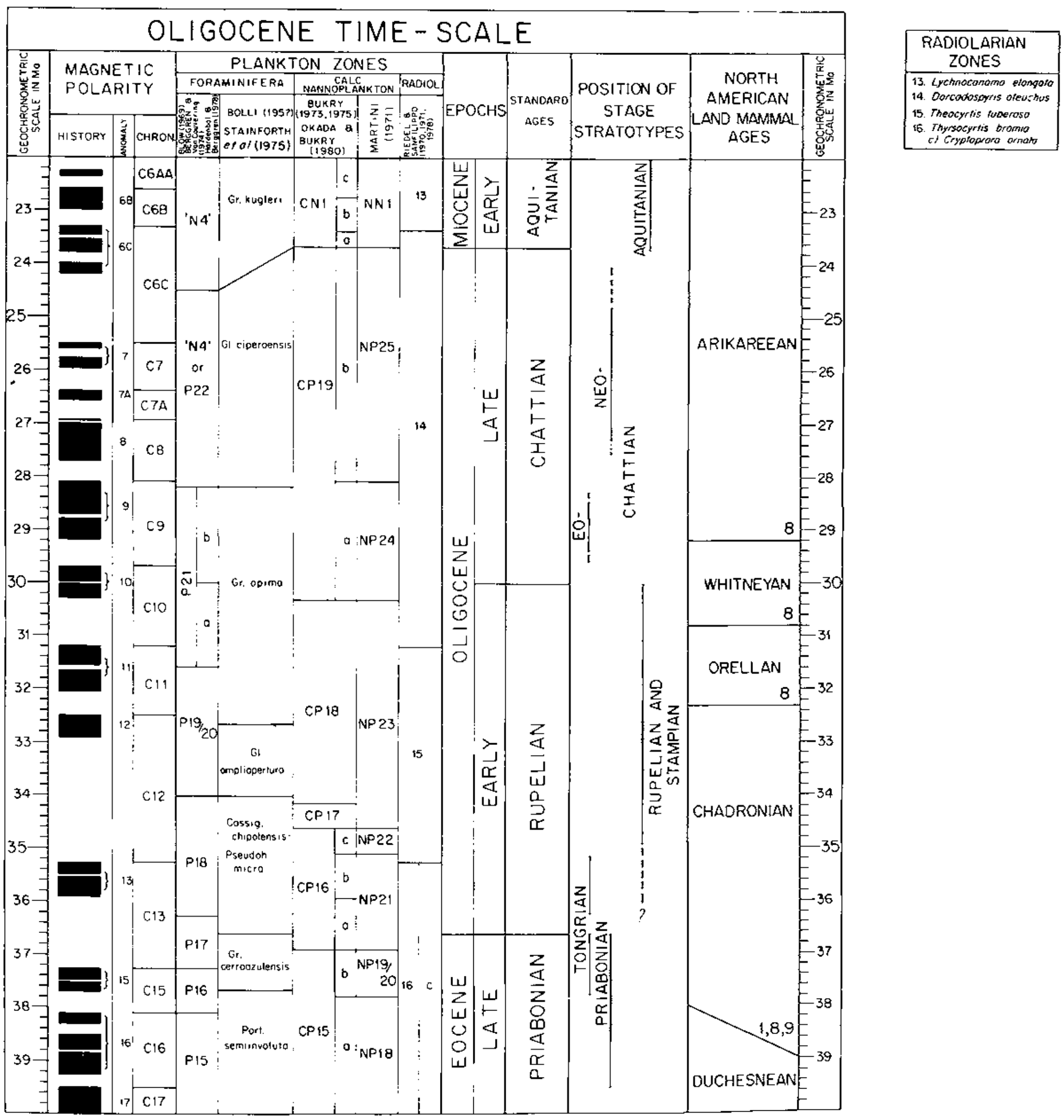

FIG. 6. Oligocene geochronology (explanation as in Fig. 3).

(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 terrestrial 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 biostratigraphy 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 hy 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 magnetostratigraphic 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 hasis 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 hiostratigraphically 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.).

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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 o (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 $\pm^{1}_{0.5}$ 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)

EOCENE

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 associated 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 &

OLIGOCENE

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.

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

OLIGOCENE

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 magnetochronology 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).

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Concerning the numerical ages proposed for the Jurassic and Cretaceous geochronology

G. S. Odin

In the presentation by Drs Hallam and Hancock some numbers are proposed to calibrate the Middle and Late Mesozoic time. For this period some (if not all) stage boundaries are actually not precisely located from a purely stratigraphical point of view, and, strictly speaking, no definitive numbers can be assigned to them. Consequently, different acceptable solutions can be proposed concerning the numerical age of these boundaries. However, some considerations developed by Hallam and Hancock do not correspond to the present state of knowledge resulting, in my view, in either incomplete or sometimes obsolete conclusions. These considerations are discussed below.

Age of the Jurassic–Cretaceous boundary

The use of radiometric dates obtained from glaucony is certainly not easy, and the older time-scales, particularly in the sections based on glauconies, must now be revised because of the systematic choice by earlier workers of the oldest available ages in instances where several glaucony ages were available.

For glauconies, as well as for other geochronometers, a diffusion loss of argon was *systematically* assumed for the youngest of a series of apparent ages. It is known that such argon loss exists in *specific conditions* which can usually be identified (Odin 1982, p. 307–85). If the conditions of burial, tectonization (and sometimes weathering) are not apparent in the dated outcrop, it is not correct to assume the possibility of argon loss without noting at the same time the possibility of obtaining apparent ages higher than the time of deposition as demonstrated by Odin & Dodson (1982).

Hallam notes on page 119: 'In view of the possibility of argon loss the glauconitic dates are more likely to be overestimated than underestimated'. This is incorrect.

Page 120: 'Because ... late Jurassic dates are based on glauconites it seems likely that they are systematically underestimated'. This is incorrect.

Page 120: 'Comparison with dates from tuffs still suggests that there may have been a frequent small loss of argon'. This is more correct but needs supporting examples; to my knowledge, the literature does not give any convincing examples in epicontinental, unweathered, shallow buried and non-tectonized deposits.

Page 122: 'Although such dates can be too high, it still seems more likely that they are minimum figures'. This also appears more correct but needs a comment as follows: Yes, glaucony apparent ages may be low compared to the time of deposition, but this must be *deduced* from each specific outcrop history before dating. The discussion could be similar for all other geochronometers. A correct comment is proposed on pages 122–3. 'The latest Barremian in SE France gave an apparent age of 110.7 Ma but ... the specimen was tectonized and the age must be regarded as a minimum'. I have proposed in the past a similar comment for this possible calibration point (Odin 1982, NDS 73).

An example of this distortion consequent upon the selection of the oldest available glaucony apparent age concerns the Jurassic–Cretaceous boundary. This example has already been fully discussed in Kennedy & Odin (1982) and the corresponding abstracts. Hallam and Hancock seem to deduce their tentative compromise of 135 ± 5 Ma essentially from the apparent age of 134 Ma obtained from the Ryazanian glaucony (*Hectoroceras* zone) from Norfolk (S 8008). The author's comments are as follows:

1. Against this single glaucony date Kennedy & Odin (1982, p. 579) have emphasized that at least six other glaucony apparent ages suggest a younger age: they are two late Portlandian glauconies from France measured in Berne at 131.0 and 129.4 Ma (NDS 75, NDS 76); three glauconies from the Portland Beds giving ages of 132–132 and 134 Ma (Dodson *et al.* 1964, S 8003 and S 8004; Obradovich 1964, KA 523) and one glaucony from the *Paracraspedites* zone of Norfolk giving an age of 135 Ma (Dodson *et al.* 1964, S 8005). Statistically, the six late Jurassic ages are younger than the single early Cretaceous age of 134 Ma.
2. The statistics, however, are not the only argument. The author also deliberately did not mention the analytical uncertainty which indicates that an age of 134 ± 3 (1σ) for S 8008 may well be 131 or 137 Ma. The apparent ages obtained from samples stratigraphically older suggest that the younger age limit for S 8008 (131 Ma) may be the more appropriate.
3. The main argument is of a geochemical nature. The author has shown that the confidence level in a glaucony apparent age depends largely on the potassium content when similar ambient conditions exist for the different samples. Here, the glauconies are all in a very similar environment, they are from the same area, some of them being collected from nearly the same outcrop. Consequently, if we have to eliminate one deviating sample *on geochemical grounds*, it is just the one *preferred* by Hallam and Hancock. This appears to be a demonstration of what the author considers to be an error due to subjectivity when objective arguments such as the potassium contents are available. The slightly 'too old' number obtained for S 8008 although not fully irreconciliable, is probably due to the frequently observed trend due to argon inherited from the substrate (see demonstration in Odin & Dodson 1982). The author considers that the fully coherent apparent ages obtained for the late Jurassic glauconies reinforce this interpretation for the Ryazanian glaucony.

The other dates published in the literature can hardly help in choosing between 130 and 135 Ma for location of the boundary. To the author's knowledge, there are no arguments at the moment for accepting an age older than 135 Ma (which is implicit in the estimate of 135 ± 5 Ma proposed by Hallam and Hancock) for the Jurassic–Cretaceous boundary. J. Hancock seems to have a similar opinion.

Age of the Cretaceous–Tertiary boundary

It must be recalled that there is no agreement between the apparent ages obtained from Colorado, which suggest a boundary at about 66 Ma (NDS 103) and those obtained from Alberta and Montana (NDS 126, 127). The more recent studies by the Edmonton laboratory published first in 1979 and 1980 clearly supersede the ones discussed in Lambert

(1971) quoted by Hallam *et al.* both because they are more recent and were obtained from the same outcrop, and because they were obtained using three different dating methods (Rb-Sr + K-Ar + U-Pb) on *three* different geochronometers (biotite, sanidine, zircon). This kind of study is nearly a model for time-scale calibration and the conclusions of Baadsgaard & Lerbekmo (1980) (see data in NDS 126 — NDS 127) must be emphasized: they lead to an age of 63.5 with a very low analytical and geochemical uncertainty for the Cretaceous–Tertiary boundary as accepted in that area.

If a choice is needed the more fully documented Canadian ages are to be quoted as a first choice. However, the results discussed during a recent presentation in Copenhagen (Lerbekmo & Baadsgaard 1983) have shown that some question of strictly analytical calibration still exists. Summarily, at least the K-Ar apparent ages of about 63.5 Ma of Edmonton are possibly analytically too young by about 1 Ma. While waiting for more details, the author suggests that the interval of time (64–66 Ma) most probably includes the boundary as drawn in North America. The authors' comment that there is very little security in correlating this North American boundary (presently defined there on continental vertebrates and palynomorphs) either with pelagic facies controlled with foraminifera or with the type locality in Denmark is agreed with by this author. The palaeomagnetic records are as ambiguous in Denmark as in the North American continental outcrops.

Age of the Turonian–Coniacian boundary

The author accepts the observation and decision of the experts that the *Inoceramus deformis* zone is now recognized as of Coniacian age. The considered date (quoted 89 ± 1.5 Ma in Odin & Obradovich (NDS 108)) is therefore a minimum age for the Turonian–Coniacian boundary and this is the same conclusion reached by Obradovich & Cobban (1976). The dates from Williams & Baadsgaard (1976) are even higher at 90–92 Ma for Coniacian samples. Those obtained from European Coniacian glauconies are either younger at 86–87 Ma (NDS 83) or similar at 90.5 Ma (NDS 60). The age of 88.1 ± 3.0 and 88.7 ± 2.2 Ma quoted in Kennedy & Odin (1982) are for *Turonian* glauconies from Belgium; this is possibly not clear in the paper quoted but clear enough in the abstracts (see NDS 82 and 164). There is no reason to consider these two as the most reliable. All these glauconies are *evolved* to *very-evolved*.

The picture given by these data is not very precise and it is suggested that the authors quote a rather large ± for their chosen estimate. The author would personally recommend now the interval of time 87–90 Ma to locate this boundary.

Other remarks

Concerning Fig. 2 of Hallam and Hancock the author would argue that it is imprudent to propose such precise estimates. Although the numbers proposed for most stage durations are certainly possible, they give an optimistic idea of the actual state of knowledge. The relative error bars are especially large for the short stages. Actually, we know nothing about the relative duration of the Turonian–Coniacian–Santonian stages, although the numbers proposed by Hallam and Hancock seem to show the Santonian about as long as the Cenomanian or as long as the two Turonian and Coniacian stages together.

There is a good example of the misguided use of these numbers (stage durations) in Table 2 of Hallam and Hancock. The authors neglect that they themselves have assigned a ± 1 Ma for their estimates of the stage boundary ages (this is especially optimistic for the two Turonian and Coniacian boundaries). They use these (uncertain) stage durations proposed in their Fig. 2, without error bar, to calculate the *mean biozone* durations. They clearly emphasize that the ammonite zone number may vary depending on outcrop, authors, etc . . . for a single stage and they calculate the 'extreme' possibilities of *mean zone* duration taking into account only the different possible numbers of biozones. However, as for the number of zones there are also different *possibilities* for the stage duration. Therefore, an actual view of what is known about *mean zone* duration must also take into account the different possible stage durations. As an example, the Coniacian stage may well be half as long or three times longer than accepted by the authors (i. e. 0.75 to 3.5 Ma) instead of 1.5 Ma according to the available radiometric data. This corresponds to a factor of variation of six to be added to the author's estimate for the *mean zone* durations in this stage.

In order to diminish the extremely divergent (and virtually useless) figures obtained from this exercise, the author proposed in October 1983 in Copenhagen (Odin 1983) to consider only either long stages, or groups of short stages for calculating the rates of geological processes (Odin 1983). In doing this, the relative error margin becomes much smaller, and the *mean zone* durations become more significant. The grouping proposed is shown in Fig. 1, together with the interval of time in which the author has proposed to locate the Cretaceous stage boundaries (Kennedy & Odin 1982). The author must reiterate that although recommended numbers are suggested in this figure (thick numbers to the left), any number included in the time intervals shown at the right of the stage sequence is equally possible given the radiometric data available. Summarily, it is proposed that for calculating *mean zone* durations, or rate of accumulation, etc . . . only the large square limits should be used. In passing, it must be noted that a similar example, i. e. absence of evaluation of an uncertainty for the rate of a geological phenomenon, is given in the same paper in Fig. 6. A sea-floor spreading rate estimate must take into account and show the known uncertainty in time-scale calibration.

Finally, the author would like to emphasize that, although the above exercise may (almost accurately) be undertaken for the Cretaceous System events, nothing equivalent can be done for the Jurassic System. Because of this a request must be addressed to stratigraphers and more generally to people working frequently in the field: better than any discussion, rediscussion, re-evaluation or recalculation of previous data, the best way to improve time-scale calibration is to find *new* dateable material, if possible better than that previously used.

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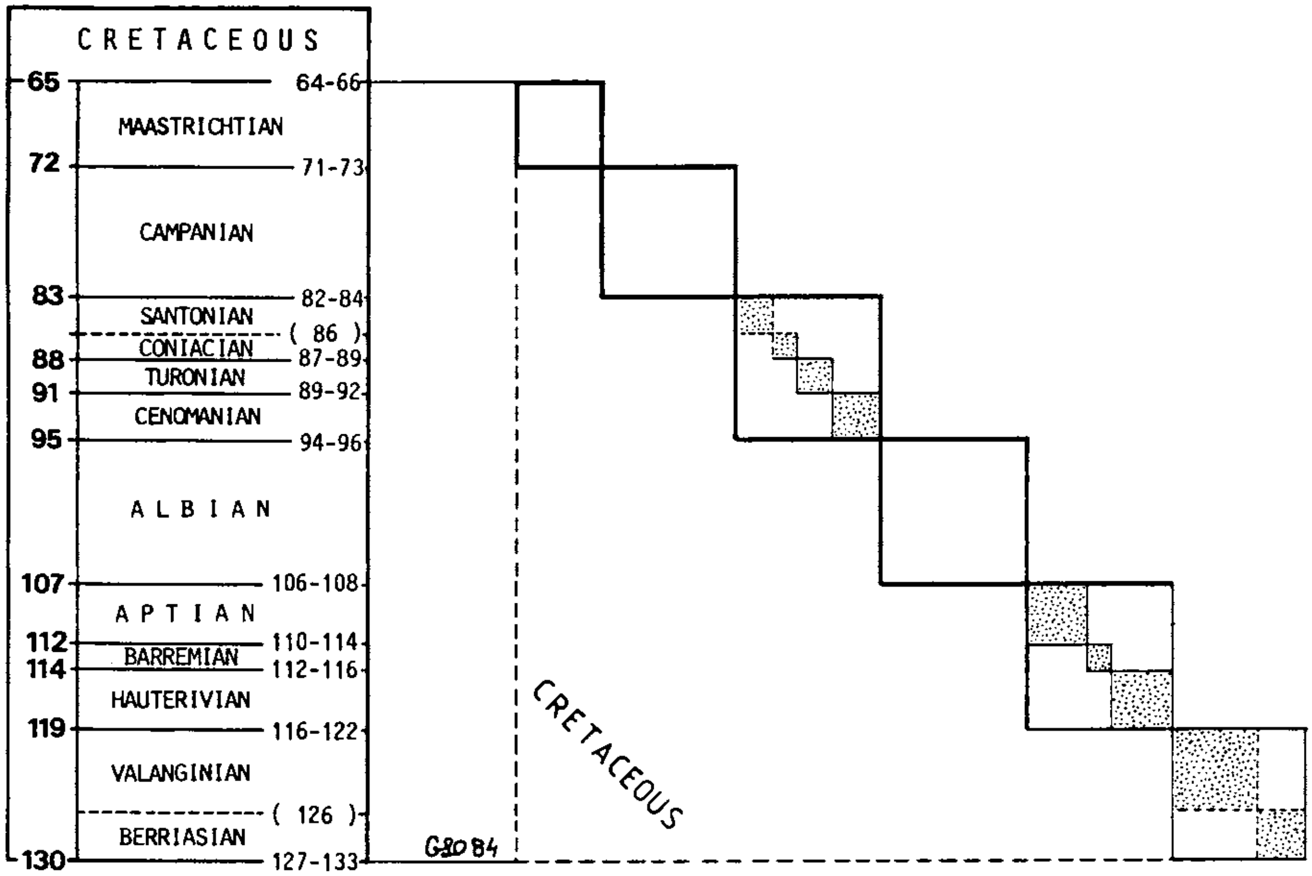

FIG. 1. The Cretaceous time-scale. The left column shows, to the left, numbers suggested to locate the boundaries; to the right the intervals of time actually defined according to radiometric dates. The age of the limit may be anywhere between the two numbers shown including them. The diagram to the right shows the duration of groups of stages for which the extreme limits are known with enough security to permit the definition of a correct duration allowing calculating meaningful speeds or durations of geological phenomena. From the base right to the top left, the Berriasian plus Valanginian stages have a total duration of about 11 Ma (130–119 Ma); Haute-rivian to Aptian about 12 Ma (119–112 Ma); Albian alone, about 12 Ma; Cenomanian to Santonian, about 12 Ma; Campanian, about 11 Ma; Maastrichtian, about 7 Ma. This defines the large squares. The small dotted squares inside the large ones have a relatively badly defined duration because their boundary ages are not located precisely enough to calculate a duration with a practical meaning.

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The Neogene: Part 1

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SUMMARY: The Neogene is taken to include the Miocene, Pliocene, Pleistocene and Holocene, and a well established sequence of marine stages now exists from the early Miocene Aquitanian to the Calabrian in the early Pleistocene. Up-to-date reviews are provided for the classification of continental Pleistocene deposits, oxygen and carbon isotope stratigraphy, organic geochemical stratigraphy and the letter classification of the East Indies Neogene. Major boundaries of the Miocene have been narrowed down to durations of about 1 Ma or less and to progressively shorter durations for the Pleistocene to Holocene boundaries. Future work will have to identify and date these boundaries in stratotype sections. Authors contributions are identified by initials at the end of each section.

Introduction

Hoernes (1856)* first used Neogene to distinguish the Eocene faunas from those of the Miocene and Pliocene, and Berggren & Van Couvering (1974) reiterated the opinion of Denizot (1957) that the Neogene in its original sense covers the interval from the base of the Miocene to the Holocene. The review of the historical usage of the Neogene by Cati *et al.* (1981) confirmed the opinions expressed by Denizot and Berggren & Van Couvering.

In the United Kingdom there are only a few on-shore Miocene and Pliocene deposits while there are far better Pleistocene and Holocene sequences. This disproportionate record may have influenced George *et al.* (1968) when they included only the Miocene–Pliocene in the Neogene and left the Pleistocene–Holocene without an equivalent time-stratigraphic term. This restricted use was also applied by Curry *et al.* (1978) in their correlation of the British Tertiary rocks. Nevertheless, beyond the shores of Britain there has been a move towards a return to the original Hoernes meaning of the Neogene. Thus Brouwer (1963) in describing the Cenozoic rocks of the Netherlands included the Pleistocene in the Neogene and further afield, in a correlation of the Atlantic, Mediterranean and Indo-Pacific Neogene, Berggren (1981) also included the Miocene, Pliocene, Pleistocene and Holocene in the Neogene.

Therefore for this work which has a world-wide aspect, the Neogene is used as a term which includes the rocks from the base of the Miocene to the present.

Stages

Unlike the Palaeogene, the Neogene international stages are reasonably well established from the early Miocene to the early Pleistocene (Fig. 1). When the stratotypes are examined and compared with better and more continuous records in deep-sea deposits, some intervals have been found to be missing; the relationships of the stratotypes to the more complete stratigraphic records is shown in Fig. 1 and is after Berggren (1981).

Above the Calabrian in the Pleistocene–Holocene there appears to be no internationally acceptable set of marine stages but various *local stages* are described by D. Q. Bowen later in this chapter. Similar local stages for the Neogene have been developed in Australia, New Zealand and California, while in the Far East there is a letter stage classification (Te_5-Th) for the Neogene which will be described by C. G. Adams.

Zones

There has been a proliferation of zonal schemes since the 1950s and these include those based on the following microfossils: planktonic foraminifera, nannofossils and Radiolaria. The relationship of these zones to the Neogene time-scale is dealt with by W. A. Berggren *et al.* in Part 2.

Ages of boundaries

The ages of Neogene boundaries are based on radiometric dates (Berggren 1969, 1972; Odin 1978; Kreuzer *et al.* 1980) or on a combination of palaeomagnetic stratigraphy and radiometric ages (La Brecque *et al.* 1977; Ness *et al.* 1980; Berggren *et al.* this volume).

It is important to understand the methods used by those compiling time-scales from palaeomagnetic and radiometric data. La Breque *et al.* (1977) used two calibration points, namely the base of anomaly 2A at 3.32 Ma and anomaly 29 at 64.9 Ma, and the dates of the other anomalies in between were 'determined by interpolation between the two anomalies'. Berggren *et al.* in this chapter have used a similar method for the Neogene with the following age calibration tie-points: the base of anomaly 2A is at 3.40 Ma, and the top of anomaly 5 is 8.87 Ma. The Palaeogene–Neogene boundary is at 23.7 Ma within anomaly 6c.

In this chapter there are detailed and up-to-date contributions on various aspects of the Neogene and the authorship of each section is given. The setting for the Neogene geochronology is by D. G. Jenkins; this is followed by a description of the Pleistocene of the continents by D. Q. Bowen and the letter classification of the Neogene from SE Asia by C. G. Adams. Results of more specialized research topics are given by N. Shackleton on oxygen and carbon isotope stratigraphy and by S. C. Brassell on organic geochemical stratigraphy; conclusions are by D. G. Jenkins.

D. G. J

* References for this chapter are to be found at the end of The Neogene: Part 2.

ERA	SYSTEM		SERIES		STAGES	STRATOTYPES
CENOZOIC	Q	NEOGENE	HOLOCENE			
			PLEISTOCENE	E	CALABRIAN	
	TERTIARY		PLIOCENE	L	PIACENIAN	
				E	ZANCLEAN	
			MIOCENE	L	MESSINIAN	
					TORTONIAN	
				M	SERRAVALIAN	
					LANGHIAN	
				E	BURDIGALIAN	
					AQUITANIAN	
		PALEOGENE	OLIGOCENE		CHATTIAN	

FIG. 1. The Neogene is equivalent to the late Tertiary and Quaternary (Q); marine stages are shown, together with intervals represented by stratotypes.

The Pleistocene of the Continents:

Despite the immediate redundancy of existing continental classification schemes once the deep-sea ^{18}0 scales were correlated with stratigraphic sequences on the continents (Fink & Kukla 1977; Kukla 1977) there has been a marked lack of progress towards the definition of formal stratigraphic units (Hedberg 1976) to place continental classifications on a satisfactory basis. Because of the continuing climatic basis for Pleistocene subdivisions the major stratigraphic units continue to be cold (glacial) and temperate (interglacial), and are accorded Stage status. Definition of Global Standard Stages, however, is distant and satisfactory stratigraphic subdivision of provincial regions not yet secured. Because the duration of most cold (glacial) events is only about 100 ka, and intervening temperate (interglacial) ones 10–30 ka, problems of inter-regional correlation without the aid of a palaeontology with first appearances and extinctions is acute not least because of the highly diachronous response of lithology and the biota to climatic changes. Moreover, repeated constellations of similar species occurred, thus further hindering correlation and classification.

Comprehensive correlation between the oceans and continents is, by comparison of differently derived scales, assisted by point-datings provided by palaeomagnetism and a variety of dating methods. ^{18}0 scales do not correspond exactly to continental ones variously constructed in different regions. For example, Shackleton (1969) argued that the Eemian Interglacial of NW Europe, which is defined on the basis of its vegetational history, corresponds only to substage 5e of the deep-sea scale. Further, whereas the Matuyama-Brunhes boundary is located in Stage 19 of deep-sea core V28-V238 (Shackleton & Opdyke 1973), that is, within an interval of relatively low continental ice-volume (an interglacial), it occurs within a cold (glacial) stage in East Germany (Helmian) and in 'Glacial A' of the Dutch Cromerian Complex (Fig. 2). It is also to be noted that, whereas the ^{18}0 scale is a surrogate for continental ice-volume (glaciation) it does not indicate the proportional ice-volume distribution of ice-sheets on the continents at any given time. Thus, arising from different atmospheric circulation regimes, the regional accumulation of ice is likely to have differed between different glacial events. This factor, related principally to an availability of precipitation (Ruddiman *et al.* 1980), is known to be of major importance during shifting ice-accumulation patterns during the Last Glaciation (Boulton 1979). Despite overlapping dating methods, restricted applicability in time and space hinder progress: ^{14}C dating, ubiquitous in application, is restricted to *c.* the last 50 ka (with extension to 70 ka by enrichment); ^{230}Th/^{234}U dating is confined to the 'coral belt' and speleothems and travertines in mid-latitudes, which, notably in caves, are often difficult to relate to the local geology; K-Ar and fission track dating are restricted to areas of vulcanicity and zones of tephra fall-out. Amino acid (Wehmiller 1982) and thermoluminescence dating (Wintle & Huntley 1982) are being increasingly applied. TL dates in Eastern Europe form the basis of correlation in Fig. 2.

Satisfactory agreement has not yet been reached on defining the base of the Pleistocene. A date of *c.*2.5 Ma seems appropriate if based on European mammalian faunas (Zagwijn 1974); around about the top of the Reunion magnetic event (2.01 Ma) if based upon the first appearance of northern 'guests' such as *Arctica islandica* in Mediterranean waters (Arias *et al.*, 1980); but at 1.7 Ma, the top of the Olduvai event, if based on first appearance and last appearance datums in deep-sea sediments (Berggren *et al.* 1980).

The Late Pleistocene

Correlation of the continental and oceanic records is sufficiently well established to allow the use of oxygen isotope stages in referring to continental events. Substage 5e, for example, is widely represented along world coastlines by marine units which show that global sea-level was slightly above present at 125 ka. On rapidly uplifting coastlines estimates of global sea-level fluctuations during the last glaciation are possible. Particularly successful have been studies on Barbados coral terraces using $^{230}Th/^{234}U$ for dating and ^{18}O analysis for correlating emerged reef-crests with the deep sea record (e.g. Fairbanks & Mathews 1978); on the Huon Peninsula, New Guinea, close to a subduction zone (Bloom *et al.* 1974; Chappell 1974); Timor (Chappell & Veeh 1978); and Haiti (Fairbanks 1982). On Bermuda $^{230}Th/^{234}U$ dates on submerged speleothems has shown the pattern of sea-level change during Stage 5 and that after substage 5e the Bermudan carbonate platform was not submerged further (Harmon *et al.* 1981). Thus correlation between the sea-level record and the deep sea isotope stratigraphy appears satisfactory. But, other dates from the US Atlantic Coastal Plain show high sea-levels after substage 5e (Cronin *et al.* 1981) which, together with current debate on continental ice-volume during the last glaciation, has led to questioning the use of ^{18}O scales as a sea-level surrogate (e.g. Williams *et al.*, 1981; Andrews 1982).

On the continents, substage 5e is generally taken to correspond with the Eemian (last interglacial) of NW Europe. But the Grande Pile, Vosges, sequence has shown two further 'interglacials', each with Mixed Oak Forest (the classical palaeobotanical definition of 'interglacial' in NW Europe), after the classical Eemian (Woillard 1978). Whether these should be accorded interglacial Stage status or correlated with Amersfoort and Brørup interstadials farther north is debatable and further highlights the difficulties in using facies-floras in correlation and classification. Within the range of ^{14}C dating and beyond using the 5e marker as a control, satisfactory intercontinental and land-sea is secure in outline from Stage 5 to the present (Fig. 2). Long sequences for this purpose occur at: Grande Pile, France (Woillard 1978, 1979), Clear Lake, California (Adam *et al.* 1981), Searles Lake, California (Smith 1979), Lynch's Crater, Queensland (Kershaw 1974), Lake Biwa, Japan (Horrie 1976), the Sabana de Bogota, Colombia (van der Hammen *et al.* 1971) and Philippi, eastern Macedonia (Wijmstra 1969). Furthermore, the records are matched by ^{18}O scales through the Greenland (Daansgard *et al.* 1969) and Antarctic (Johnsen *et al.* 1972) ice caps and the sea-level record of the Huon Peninsula (Bloom *et al.* 1974).

The base of the last glaciation is placed at 120 ka so as to include sub-stages 5a to 5d of the ^{18}O deep-sea scale. In the USA the Sangamon Interglacial Stage is now restricted to 130–120 ka (although palaeosols of that name are both younger and considerably older than that restricted time span) and a new subdivision, the Eo-Wisconsin, recognized between 120 and 70 ka (Richmond 1982). The USA scheme also recognizes a Middle Wisconsin (55–30 ka), a subdivision recognized generally (Fig. 6) although not always with the same precisely dated boundaries: for example, in the British Isles, the Middle Devensian is defined between 50 and 26 ka (Mitchell *et al.* 1973). This interval of generally ameliorated temperatures (interstadial or 'interstadial complex') corresponds to stage 3 of the ^{18}O deep-sea scale. But workers in Siberia ascribe the Karginsky event at this time to full interglacial status. Other interstadial events occur in the early part of the last glaciation, some of which have been dated by radiocarbon enrichment methods. It is likely, however, that these merely represent minimum ages. A feature of some importance is the uncertain characterization both in time and space of the early last glaciation ice advance. It corresponds to Stage 2 of the ^{18}O deep-sea record but because of dating uncertainties is unclear on the continents. In North America, an early Wisconsin advance is only known in the Eerie lobe of the Laurentide ice sheet and in New England. In Europe, greater uncertainty exists on its timing and extent. In contrast, the extent and timing of the late last glaciation advance, between *c*.30 and 10 ka, is relatively well known (e.g. Clayton & Moran 1982), and the overall synchronous nature of intercontinental correlation clearly established (Fig. 2). In more detail, however, regional and local factors combine to produce marked lack of synchroneity between different lobes of the same ice-sheet and between regions (Fig. 2). The major disagreement which obtains on the origins dimensions, extent and timing of Late Wisconsin-Weichselian ice sheets arises in part because of the extreme scarcity of radiocarbon dates on the northern margin of the ice sheets. Thus conflict between a maximum (Denton & Hughes 1981) and minimum reconstruction (Boulton 1979; Andrews 1982) has arisen.

Of considerable local significance is the renewed ice growth after the Allerød interstadial in northern Europe. In the British Isles the growth of the Loch Lomond ice sheet took place between 11 and 10 ka, after the Windermere interstadial, and is related to a sudden rapid southerly movement of the ocean Polar Front (Sissons 1981; Ruddiman & McIntyre 1981).

The Early and Middle Pleistocene

Redundancy of the classical classification models created greatest uncertainty in the Middle Pleistocene of Europe and North America. Of work in progress, the location of the Matuyama–Brunhes boundary has proved crucial because between it and the position of the last interglacial (5e) should occur seven major cycles (Fig. 2). Potentially the most complete results have been obtained from the loess sequences of the Moravian Corridor (Czechoslovakia) which connects the Alpine and North European glaciated areas. The Matuyama–Brunhes boundary is found at several localities, notably the palaeontologically important site of Stranska Skala (Musil *et al.* 1955) and at Cerveny Kopec (Red Hill), Brno. Seven Loess Cycles defined by marklines (boundaries between thick loess units and overlying palaeosols) were described around Brno and related to local terraces (Kukla 1975, 1977). Further work revealed the Jaramillo event (Bucha *et al.* 1979). Matuyama–Brunhes, Jaramillo and Olduvai events were found at Krems, Austria, where nine major depositional cycles occurred between the top of the Olduvai event and the base of Brunhes (Fink & Kukla 1977). Thus, together with the eight cycles of middle and late Pleistocene time, some seventeen major ones occurred during the entire Pleistocene Epoch as defined from 1.7 Ma.

At Voigstedt, near Halle (East Germany), the Matuyama–Brunhes boundary lies within fluvioglacial gravels of the Helme Stage (Cepek *et al.* 1977) and is overlain by a sequence including the Voigstedt interglacial. This includes an Upper Biharian veterbrate fauna which is correlated with the Cromerian of West Runton, Norfolk (Stuart 1975). The latter is younger than the Cromerian Complex of the Netherlands (van der Hammen *et al.* 1971) (Fig. 2). Together with the palaeosol sequence at Mahlis, north-west of Dresden (Fuhrmann *et al.* 1977; Wiegank 1979), it is possible to show that three interglacials occur between the Matuyama–Brunhes boundary and the basal till of the Elster glaciation at Voigstedt (Fig. 2). The boundary lies in the first cold stage (Helme) above the Artern interglacial which also corresponds

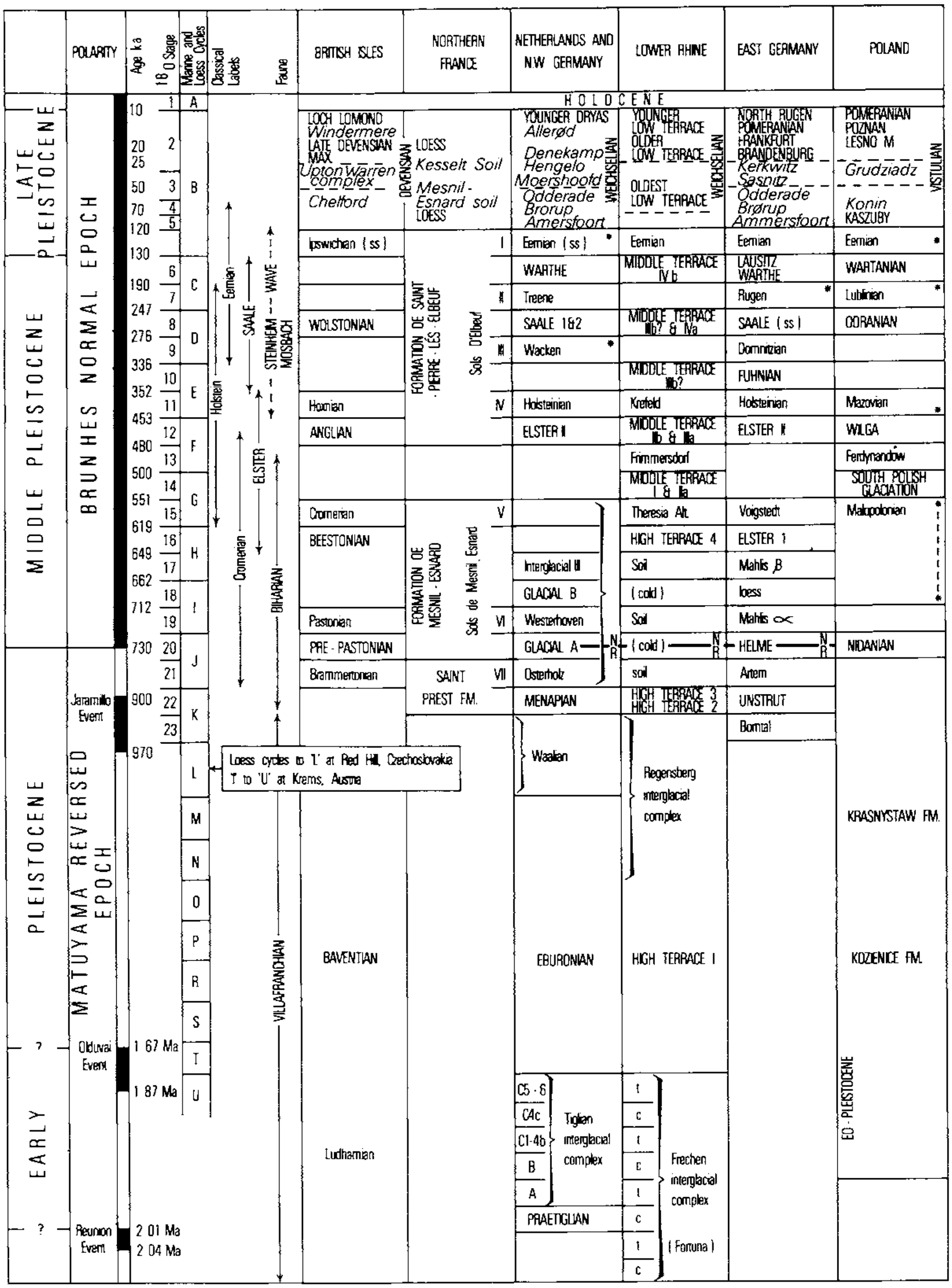

FIG. 2. European continental Pleistocene.

to its position in the Netherlands (Glacial A) (Zagwijn *et al.* 1971) and in the Ville interglacial complex of the Lower Rhine (Brunnacker 1979). Further point datings (Fig. 2) are provided by $^{230}Th/^{234}U$ (Glazek *et al.* 1980; Schwarcz 1980), and TL dates (Arkhipov *et al.* 1982; Lindner 1982; Zubakov & Pisarevsky 1982).

Palaeomagnetic work in the Netherlands shows that the upper part of the Waalian interglacial probably corresponds to the Jaramillo event (Van Montfrans 1971). The Menapian cold stage probably corresponds with High Terrace III of the Rhine and the Eburonian cold stage with High Terrace I (Brunnacker & Tillmans 1982). Long pollen sequences from Colombia (extending to Waalian), Macedonia (extending to before the Cromerian complex) and from Seneze, Massif

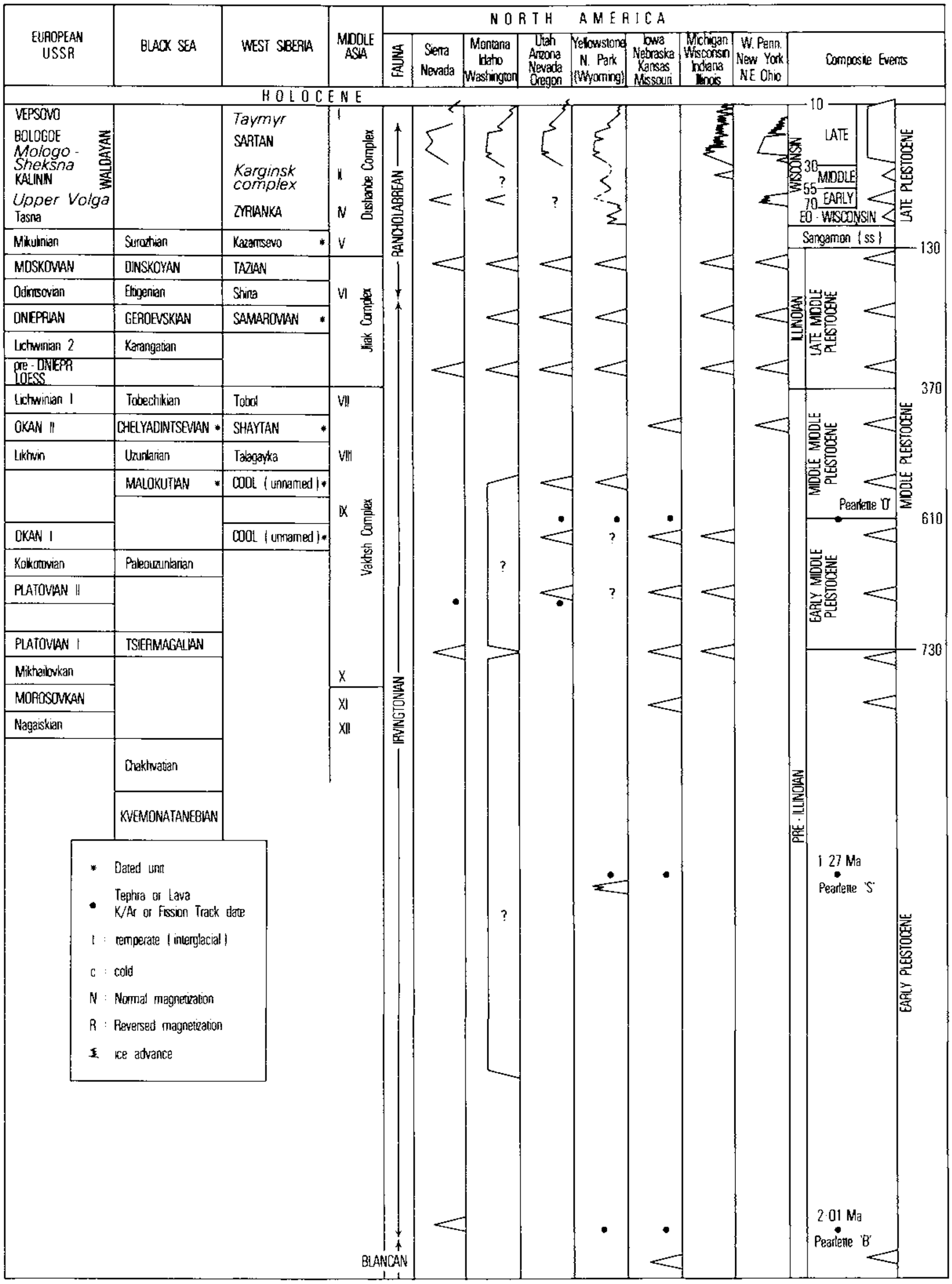

FIG. 2 continued; with North American continental Pleistocene (After Richmond 1983.)

Centrale, France (extending to well below Tiglian), have been correlated with the Netherlands sequence (van der Hammen *et al.* 1971). Farther east in Central Asia palaeomagnetic data and thermoluminescence dates have enabled subdivision of long loess sequences which contain nine palaeosol complexes younger than the Matuyama–Brunhes boundary (Fig. 2). Complex V is ascribed to the last interglacial (5e) by TL dating and because it is overlain by a reversal thought to be the Blake event (Dodonov *et al.* 1977). Loess sequences in China may also be correlated with the general sequence postdating the Matuyama–Brunhes boundary (Sun Dian-quing *et al.* 1981).

In North America the classical labels, Nebraskan and Kansan, are officially redundant (Richmond 1982). The Middle and early Pleistocene sequence of glaciations have been dated with reference to the palaeomagnetic time-scale and dated tephra deposits of the Pearlette group (Fig. 2). When the collective record of different regions is collated the scale of continental glaciation is readily matched with the ^{18}O ice-volume scale. Some discrepancies in the timing of earlier events are probably due to estimates of the dates of the deep-sea scale which may require further revision (e.g. cf Shackleton & Opdyke 1973 with Kominz *et al.* 1979 and Pisias & Moore 1981). The Middle Pleistocene of North America is subdivided and continental glacial events placed within that framework (Fig. 2). At least three major glaciations occurred within the late Middle Pleistocene (Illinoian); one Middle Pleistocene glaciation younger than the Pearlette type O tephra occurred; two early Middle Pleistocene glaciations older than the Pearlette type O tephra but younger than the Brunhes-Matuyama reversal occurred; and two early Pleistocene glaciations respectively older than Brunhes-Matuyama and the Pearlette type S tephra (1.27 Ma) occurred. When the stratigraphic record of Mountain glaciation is added to that of the Continental area a composite scale can be constructed (Fig. 2).

D.Q.B.

The Letter Classification and the Neogene:

The Letter Classification of Tertiary strata was introduced by van der Vlerk & Umbgrove (1927) for use in the East Indies at a time when it seemed impossible to apply the conventional stratigraphic terminology of Europe to this area. It was further developed and/or modified by Leupold & van der Vlerk (1931), Rutten (*In:* van Bemmelen 1949), van der Vlerk (1950, 1955), and was revised and applied to the entire Indo-West Pacific region by Adams (1970). The main changes made to the mid-Tertiary part of this classification during the last 50 years are summarized in Fig. 3. The classification is used mainly for shallow-water carbonates, in which sediments larger foraminifera are especially common.

The East Indies Letter Stages are not stages in the usual stratigraphic sense of the term, but assemblage zones based on associations of certain larger foraminiferal genera. However, the term letter 'stage' is now so firmly entrenched in the literature that it has to be maintained.

Changes to the Letter Classification since the last major revision have resulted both from new discoveries and from an improved knowledge of the ranges of individual taxa through their occasional association with age-diagnostic planktonic foraminifera. The most important changes concern the positions of the e/f and f/g boundaries. Figure 4 shows the relationship between the stage boundaries and Blow's planktonic zonation. For further information see Adams (1984).

One of the reasons for the present controversy (see Cati 1981) over the position and recognition of the Paleogene–Neogene boundary is that larger foraminifera, like their planktonic cousins, underwent no major changes at the end of Oligocene times (Drooger 1979; Adams 1981), and faunas regarded as early Miocene as recently as 20 years ago (e.g. by Eames *et al.* 1962) have now been recognized as late Oligocene (Chattian). Because the Paleogene–Neogene boundary is difficult to recognize, the Oligocene letter stages are included in Figs 3 & 4.

As far as the Oligocene is concerned, the outstanding biostratigraphic problems are palaeobiogeographic in origin. It is necessary to know whether *Lepidocyclina* (*Eulepidina*), the critical marker for the Tc/d boundary, appears everywhere at the base of Zone P19, and whether *Nummulites fichteli* (the principal Td/e boundary marker) became extinct across the whole region at the top of this zone. Only planktonic control over a number of first and last appearances in different parts of the region will solve these problems, and enable us to decide whether these boundaries are truly coincident with time planes or somewhat diachronous.

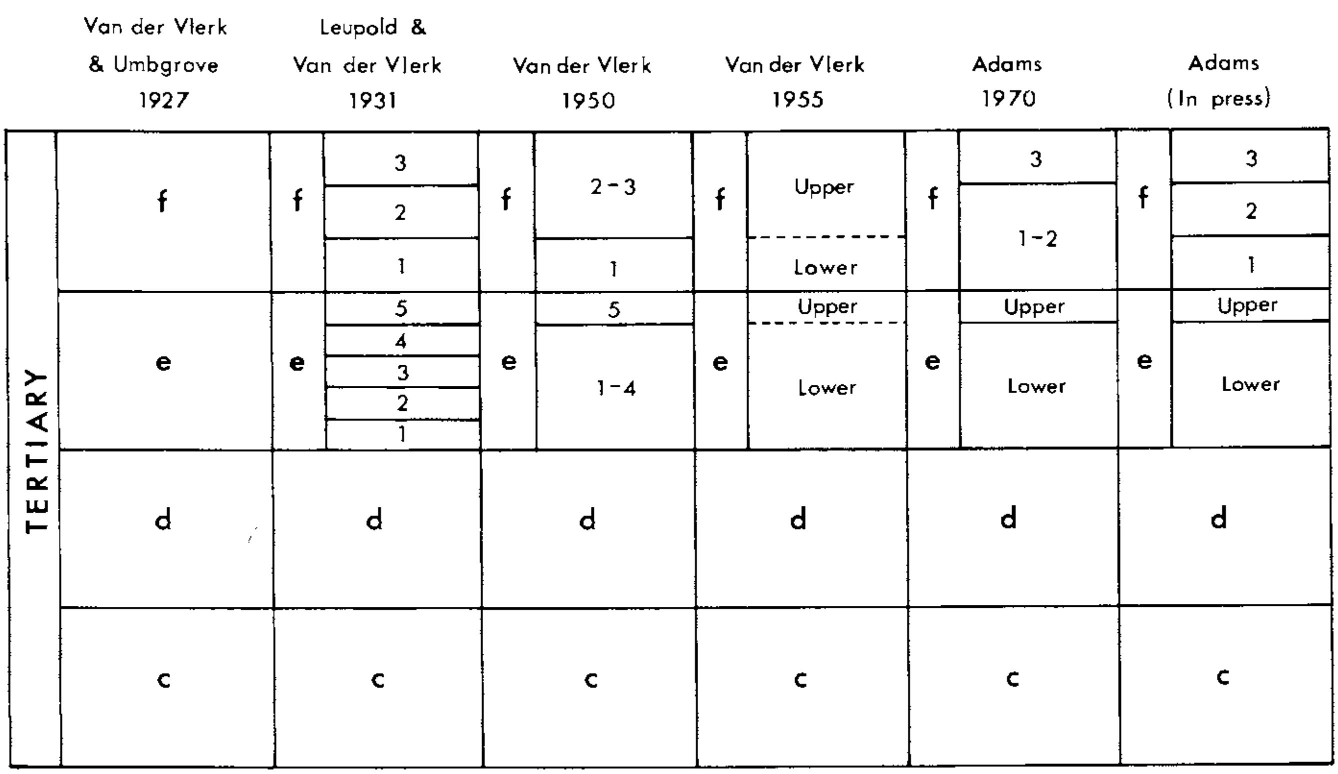

FIG. 3. Diagram showing important changes to the mid-Tertiary part of the East Indies Letter Classification since its introduction in 1927. The boundaries between lower and upper e and lower and upper f in the column headed van der Vlerk (1955) are shown by broken lines because he neither explained his use of these terms nor referred to van der Vlerk (1950). His meaning, therefore, has to be inferred.

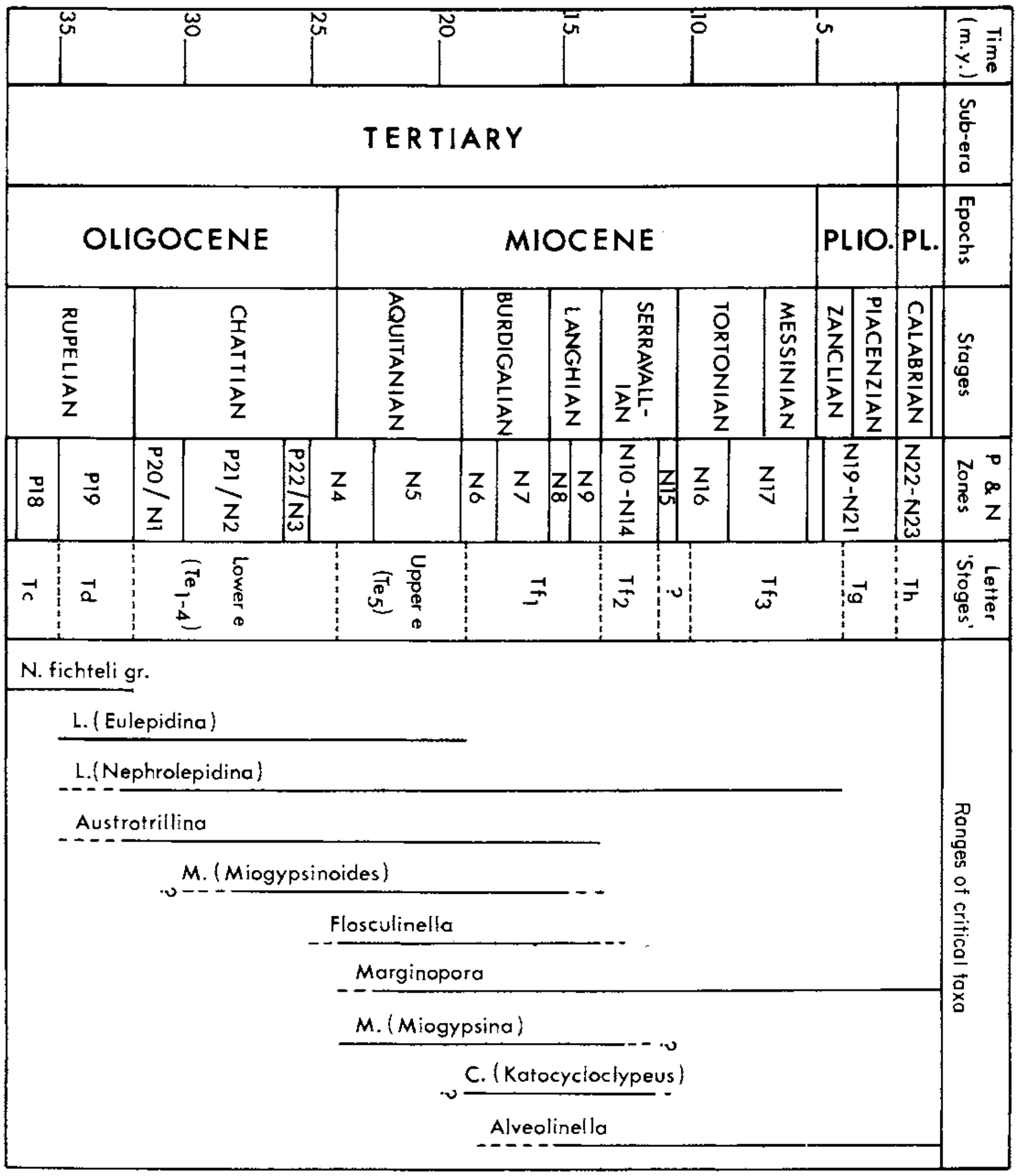

FIG. 4. Diagram showing the approximate position of the Letter Stage boundaries (Palaeocene and Eocene excluded) in relation to Blow's P and N zones and the European stages. Chronometric scale, epochs, stages and planktonic zones after Van Couvering & Berggren (1977) and Hardenbol & Berggren (1978). Letter stage boundaries and ranges of taxa mentioned in the text after Adams (1984).

The lower e/upper e boundary is defined by the first appearance of *Miogypsina* (*Miogypsina*). It would, perhaps, be better defined by tbe first appearance of *M.* (*M.*) *gunteri* Cole, but species of *Miogypsina* can be identified only from accurately orientated thin sections of individual specimens, and these are not usually obtainable from hard limestones.

The Tertiary e/f boundary, once thought to be more or less concident with the Lower–Middle Miocene boundary, is now known to fall well within the Lower Miocene (Haak & Postuma 1975; Chapronière 1981), and many rock units previously regarded as Middle Miocene are now known to be slightly older.

Tertiary f has a chequered history (Fig. 3). The three divisions recognized by Leupold & van der Vlerk (1931) were not maintained for long, and although recent work suggests that a tripartite division may still be possible, authors may prefer to use two divisions only. Adams (1984) has indicated that f_1 can be distinguished by the presence of *Austrotrillina howchini* (Schlumberger). *M.* (*Miogypsina*) spp and *M.* (*Miogypsinoides*) spp, *Flosculinella bontangensis* (Rutten) and *F. borneensis* (Tan). Tertiary f_2 lacks *Austrotrillina, Flosculinella,* and *Miogypsinoides*, but contains *Cycloclypeus* (*Katacycloclypeus*) *annulatus* Martin, while f_3 lacks *C.* (*Katacycloclypeus*) and yields only *Lepidocyclina* and other commonly occurring Miocene species such as *Marginopora vertebralis* Blainville and *Alveolinella quoyi* d'Orbigny. Since the last representatives of *Lepidocyclina* are known from Zone N19 (see Adams, 1984, for list of records), the Tf/g boundary must fall just within the Pliocene.

Tertiary g, a stage not properly defined by its original authors, represents the greater part of Pliocene time, and differs from f_3 only in lacking *Lepidocyclina*. Th is an unnecessary division, indistinguishable from Tg.

Although the Letter Classification is still in use — recent applications include those of Chapronière (1981) and Matsumaru & Barcelona (1982) — its importance will decline as the ranges of larger foraminifera become better known in terms of the planktonic zonation and therefore of the European stages. The letter stage boundaries of the Paleocene and Eocene are currently under revision, but it is hoped that a zonation of shallow-water carbonates by larger foraminifera, now in preparation by the writer, will serve as a replacement for the Letter Classification which is scarcely susceptible to further significant improvement.

C.G.A.

Neogene oxygen and carbon isotope stratigraphy

Cesare Emiliani initiated oxygen isotope studies in Neogene sediments. After first attempting to work on a Californian Pliocene section (Emiliani & Epstein, 1953) he appreciated that there was a better chance of obtaining useful data by studying deep-sea material away from the complications of local salinity variations in the nearshore environment. His first papers dealing with deep-sea deposits remain very valuable sources of basic data as well as being of historical importance (see especially Emiliani 1954a, 1954b, 1955, 1956).

Emiliani's important findings may be summarized as follows. First, through the Tertiary, the $^{18}O/^{16}O$ ratio in benthonic foraminifera became progressively greater. Emiliani interpreted this in terms of the effect of temperature on the equilibrium oxygen isotopic fractionation between water and calcite, and so inferred a cooling of ocean deep waters. This he judged to be the result of a cooling of surface waters at high latitudes. Much of the subsequent work has been to refine the picture painted by Emiliani; although considerable effort has gone into determining the timing of the contribution which must derive from the ocean isotopic change resulting from the accumulation of the isotopically light ice on Antarctica, the conclusion remains uncertain.

Emiliani's second major finding was the quasi-cyclic isotopic variability that occurred during the Pleistocene. His first venture in this field (Emiliani 1955) is a true masterpiece that has set the standards for the following decades. It is interesting to reflect that Emiliani established the principal that every isotopic measurement should be printed, followed by most workers in the field, which is certainly one reason for the widepread use of isotopic data by Quaternary geologists who are not themselves isotope specialists. Again, subsequent work has refined and extended Emiliani's early work, and much discussion has revolved, and will continue to revolve, about the relative contributions of temperature change, and isotopic changes resulting from the accumulation of isotopically light continental ice sheets, to the overall picture.

Knowledge of the stratigraphic record of oceanic carbon isotope changes has grown more slowly and with fewer outstanding scientific contributions. Although the majority of workers have made ^{13}C measurements whenever they analysed foraminifera for ^{18}O, for many years the data were either published without comment, or remained in laboratory files.

In the following sections the current state of understanding of the Miocene oxygen isotope record, of the Pliocene oxygen isotope record and of the Pleistocene oxygen isotope record are discussed separately, since there are logical breaks at these positions.

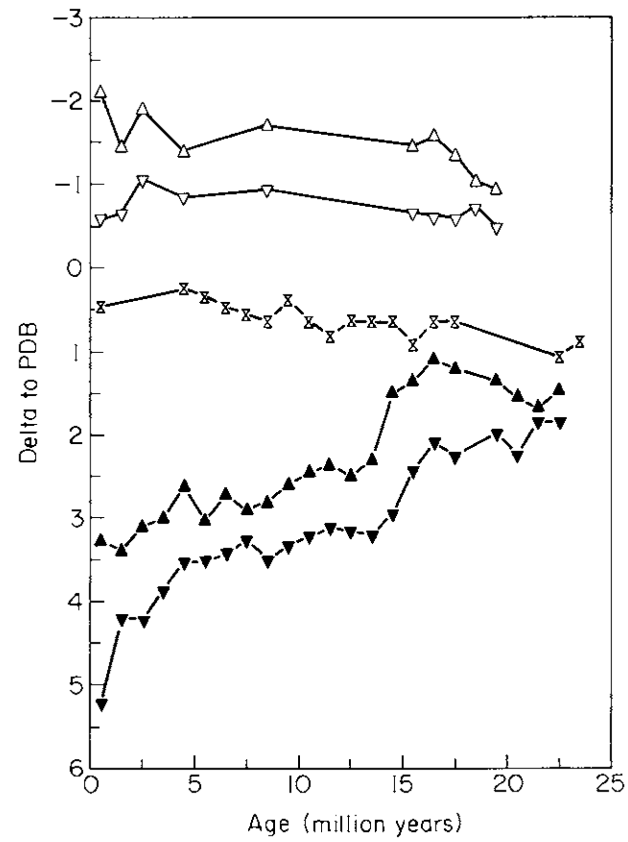

FIG. 5. Neogene oxygen isotope records from the tropics (West equatorial Pacific), the mid-latitudes (Walvis Ridge, South Atlantic) and deep water (Pacific). For the equatorial Pacific, the isotopically lightest and heaviest value in surface-dwelling planktonic foraminifera is plotted for each million-year increment in which we have detailed data. For the mid-latitudes the mean of all values for surface-dwelling planktonic foraminifera is plotted. For the deep water the isotopically lightest and heaviest value for calibrated monospecific benthonic foraminifera in each million-year increment is plotted. Compiled from data in Shackleton & Opdyke (1973, 1976); Shackleton (1982); Shackleton, Hall & Boersma (1984) and unpublished data.

△ ▽ } West equatorial Pacific

⧗ } Mid-latitudes

▲ ▼ } Deep water (Pacific)

Miocene

Figure 5 shows the oxygen isotope record of benthonic and planktonic foraminifera from the Western equatorial Pacific (mainly from DSDP Site 289 (Woodruff *et al.* 1981, Savin *et al.* 1981, Shackleton 1982, supplemented in the Pleistocene by piston core data), together with a composite planktonic foraminiferal record from the Walvis Ridege in the South Atlantic (Shackleton *et al.* 1984). The Site 289 data set is the most detailed so far available, and shows for the first time the dramatic fluctuations associated, it is thought, with the growth of the Antarctic ice sheet shortly after 15 Ma. For Fig. 5 all the published data have been re-assigned ages using the time-scale of Berggren *et al.*, this volume. The most striking aspect of Fig. 5, and one that is unaffected by uncertainties in the timing of the ocean isotopic change that resulted from the accumulation of Antarctic ice, is the dramatic increase in the vertical and latitudinal temperature gradients in the ocean that have occurred during the Neogene. Figure 5 shows more clearly than has hitherto been possible, that the major change has been in high latitudes. The temperature difference between the Western equatorial Pacific (one of the warmest spots on the ocean surface) and the Atlantic at 30° South, has not changed very much, but the temperature difference between the South Atlantic surface, and the ocean deep waters, has increased enormously. Presumably the surface temperature gradient between 30°S and the Antarctic coast has likewise increased dramatically.

In these particular records it would be difficult to treat the variations mathematically because of drilling disturbance. However, the existence of these easily-measured variations will in the future certainly attract the attention of analysts interested in detecting frequencies associated with changes in the earth's orbital geometry. On the one hand, this highlights the importance of developing a good time-scale for the Miocene; on the other, it indicates the longer-term possibility of obtaining an astronomically calibrated time-scale for this part of the Neogene.

The data of Shackleton (1982) in the Lower Miocene in Site 289 show quasi-cyclic changes on a similar time-scale (although smaller amplitude) to the dramatic fluctuations documented by Woodruff *et al.* (1981) in the top of the Middle Miocene. Unlike Woodruff *et al.* (1981) he was able to demonstrate synchronous variability in benthic and planktonic foraminifera, which is interesting in that it would be consistent with the existence of quasi-cyclic variations in oceanic isotopic composition, and hence of Antarctic ice volume, prior to the mid-Miocene. Recently Denton *et al.* (in MS) have provided field evidence which they consider suggests the existence of an ice sheet prior to the Middle Miocene. It now seems possible that the argument put forward by Shackleton & Kennett (1975a) was not correct, and that the well-documented mid-Miocene event was not the first major Antarctic glacial. Whether or not that is the case, it is clear from the data in Fig. 5, and from abundant other data (Douglas & Savin 1973; Shackleton & Kennett 1975a; Boersma & Shackleton 1977) that *either* (a) there was much more ice on Antarctica in the late Miocene and since, than in the early Miocene and before, or (b) the oceanic deep waters were significantly cooler after the mid-Miocene than before. The argument put forward by Shackleton & Kennett (1975a) was that high-latitude, and deep-water, cooling must have preceded ice accumulation. However, it may well be the case that the final cooling of oceanic deep waters was controlled not only by the cooling of high latitude surface waters, but also by reduction in the heat input into the deep water from the Tethys, a contribution which has clearly varied very considerably during the interval under discussion.

The Pliocene record; Northern hemisphere refrigeration

The most important contributions to the understanding of the Pliocene, and the onset of small fluctuations at about 3 Ma (Shackleton & Kennett 1975b) and the palaeomagnetic and oxygen isotope analysis of core V28-179 (Shackleton & Opdyke 1977). This showed very clearly the contrast between the rather minor climatic fluctuations of much of the Pliocene, and the onset of small fluctuations at about 3 Ma followed by larger fluctuations about 2.4 Ma. Previous workers had shown evidence for glaciation on Iceland as far back as 3 my ago (McDougall & Wensink 1966), for a glacio-eustatic sea-level lowering about 2.5 my ago (Stipp *et al.* 1967), and for Pliocene icerafting in the North Atlantic (Berggren 1972). Figure 6 shows the record for core V28-179 (Shackleton & Opdyke 1977, and unpublished additional data). It is becoming clear that the glacial event at about 2.4 Ma was the major event and that if any move were to be made in the future to select a new stratotype for the Pliocene–Pleistocene boundary that reflects a major environmental break, rather than adherence to historical precedent, then a boundary near this date should be seriously considered (cf., Zagwijn 1975). It was established by Emiliani *et al.* (1961) that in the Le Castella section the Pliocene–Pleistocene boundary is in the midst of a typical series of isotopic fluctuations of Lower Pleistocene character rather than at the beginning of such a series, and quantitative biostratigraphic studies (Backman *et al.* 1983) across the boundary in the proposed Vrica Section (Selli *et al.* 1977) again establish beyond doubt that the boundary is well within the interval of Pleistocene-like global isotopic fluctuations at an age of about 1.5 Ma.

Pleistocene oxygen isotope variations

Despite the enormous importance of the work of Emiliani (1955 and subsequently), many workers have adopted the data published by Shackleton & Opdyke (1973, 1976) as representing the standard sequences for the Pleistocene. The reasons for this could only be discovered by lengthy discussion, but it is important to appreciate that at least the following factors were significant:

(a) The very extensive discussions of chronology (Broecker

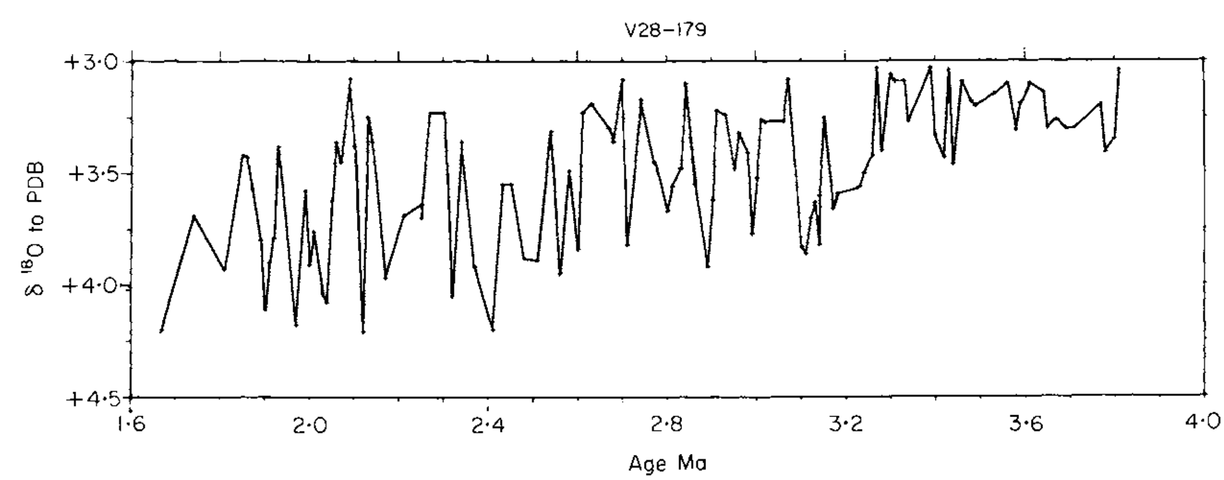

Fig. 6. Oxygen isotope record of the late Pliocene in core V28-179.

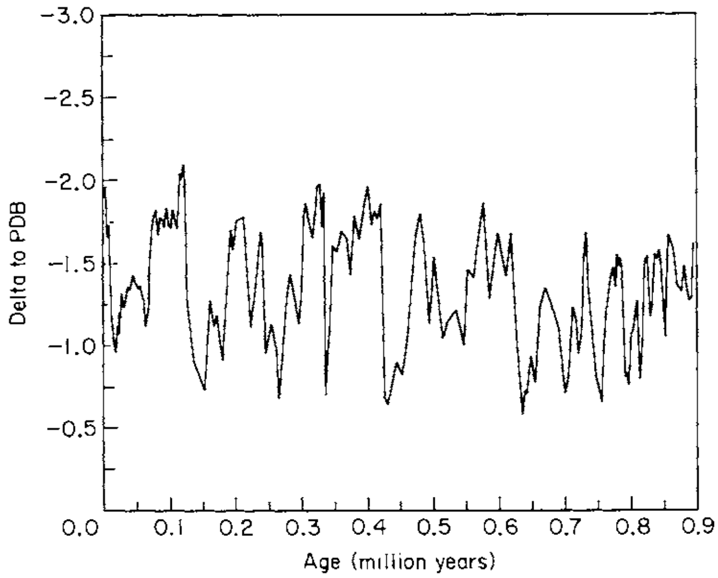

FIG. 7. Late Pleistocene oxygen isotope data of core V28-238 after Shackleton 8 Opdyke (1973), redrawn to an astronomically calibrated time-scale (Imbrie *et al.* 1984).

& Ku 1969; Rona & Emiliani 1969, Mo *et al.* 1971; Shackleton 1971) which must have confused many potential users of the oxygen isotope stratigraphy were rapidly dispelled by the rapid publication of several records which extended past the Brunhes-Matuyama reversal boundary (Parkin & Shackleton 1973; Shackleton & Opdyke 1973, 1976; van Donk 1976).

(b) Shackleton & Opdyke (1973) put considerable stress on the approach that the major value of the Pleistocene oxygen isotope record was as a stratigraphic correlation tool (which function it has admirably filled), at a time when more exciting methods were being developed for palaeo-climatic and palaeo-oceanographic reconstruction (Imbrie & Kipp 1971; CLIMAP 1976).

(c) The publications of Emiliani & Shackleton (1974), of Shackleton & Matthews (1977) and others, rapidly broke down the widespread misunderstandings about the marine oxygen isotope records that had grown up during many years of dispute (Ericson & Wollin 1956; Emiliani 1964; Evans 1971; Broecker & van Donk 1970).

Figure 7 illustrates the well-known oxygen isotope record of core V28–238 (Shackleton & Opdyke 1973), redrawn according to an astronomically calibrated timescale (Imbrie, *et al.* 1984). There are still important discussions as to precisely how much of the variability is due to changes in oceanic isotopic composition (and thus roughly proportional to sea-level) and how much to temperature. At least some of the dispute is based on inappropriate data. For example, the estimate by Yapp & Epstein (1977) for the isotopic composition of the North American ice sheet on the basis of the D/H ratios in trees that grew near its margin is analogous to estimating the isotopic composition of Antarctic ice (well known from surface and ice core study to be about −40 to −50 per mil; Dansgaard *et al.* 1973) from tbe isotopic composition of snow falling on the Southern Ocean (about −20 per mil; Craig, 1965). The true value for the mean isotopic composition of the Laurentide Ice Sheet must be close to −35 per mil (Olausson 1965; Dansgaard & Tauber 1969) implying a relationship between sea-level and δ (ocean) about 10 m per 0.1 per mil isotopic change. Less secure is the estimate of the amount of ice; informed estimates range over a factor of two. A value 160 m would imply negligible temperature drop in the deep Pacific, and less than 1 degree change in the deep Atlantic. A value 100 m would imply middepth (3000 m) temperatures in the equatorial Pacific about 0°C with similar values in the North Atlantic. There is scope for more ocean modelling studies to be applied to this problem. It should be noted that although the data in Fig. 7 give a reasonable portrayal of the broad pattern of ice-volume fluctuations during the past million years, better high-resolution records of the late Pleistocene had been obtained long before (e.g. core 280, in Emiliani 1955) while today the most characteristic detailed records are probably those obtained from the analysis of benthic foraminifera in the Eastern equatorial Pacific (e.g. V19–29, in Ninkovich & Shackleton 1975; V19–30, in Shackleton *et al.* 1983).

The carbon isotope record

Shackleton & Kennett (1975a) first drew attention to the fact that ocean $\delta^{13}C$ was considerably more isotopically positive in the Middle Miocene than it is today. This is emphasized by the data of Woodruff *et al.* (1981) and more recently by the detailed study of Shackleton & Hall (in press) who demonstrate by analysis of the ^{13}C content of bulk sediment that this is truly a phenomenon of changing ocean carbon budget (in the manner exemplified by Scholle & Arthur (1980) for the Cretaceous) rather than being only a result of changes in internal carbon cycling within the ocean. Since the Middle Miocene there has been a steady net loss of stored organic matter (coal, oil, peat etc.) on the globe and this reservoir of isotopically light carbon has gradually isotopically lightened the carbon entering the limestone pool.

In the Pleistocene, Shackleton (1977) argued that large carbon isotopic variations which he discovered in benthic foraminifera from a core in the North Atlantic were a measure of changes in the amount of carbon stored up on the globe, primarily in the living biomass of the tropical rainforests. Broecker (1982) put forward an alternative hypothesis based on Shackleton's data in which the organic matter on the continental shelves played a more important role, but in the meantime Shackleton *et al.* (1983) have shown by obtaining a carbon isotope record for the deep Pacific that in reality about one half the signal originally documented by Shackleton (1977) is attributable to changes in ocean deep circulation. Further work will be needed to establish the relative merits of the hypotheses of Shackleton (1977) and of Broecker (1982) for explaining the remaining half of the carbon isotope record which certainly has to he explained by one or other (or a combination) of the two factors discussed.

Conclusions

Despite the enormous amount of data that have been generated since Emiliani started work on the stable isotope stratigraphy of the Neogene, this remains a very fertile area of research. As our understanding of global Neogene stratigraphy and chronology improves, so the value of this field will increase.

N.J.S.

Organic geochemical stratigraphy

At present stratigraphic studies using organic geochemical methods are proceeding in two directions. First, fluctuations in the occurrence/abundance of specific molecules in sediment cores are being evaluated and related to climatic signals. Second, the systematic changes in the composition of organic matter effected by diagenesis can provide an accurate measure of maturity in sedimentary sequences and thence can be related to the sediment age.

Climatic information

The potential of organic compounds as climatic indicators stems from the relationships that exist between an organism's lipid composition and its growth temperature. In particular a greater degree of lipid unsaturation (e.g. in triglycerides and carboxylic acids) is biosynthesized under colder conditions (e.g. Holton *et al.* 1964) as part of the biochemical control of membrane flexibility. The first geochemical application of this principle was the discovery that core sections from Lake Biwa containing greater amounts of polyunsaturated carboxylic acids correlated with colder climatic episodes over the past 20000 years as suggested by palynological evidence (Kawamura & Ishiwatari 1981). Another, as yet unproven, indicator of water column temperatures is the proportion of di- to tri-unsaturated C_{37} to C_{39} ketones, compounds thought to derive from coccolithophores (Volkman *et al.* 1980), in various Neogene DSDP sediments (Eglinton *et al.*, 1983). Here sediments deposited beneath warmer waters, such as in the Middle America Trench, contain a markedly smaller proportion of the triunsaturated ketones than do sediments from beneath colder waters, such as those from the Japan Trench or Walvis Ridge. A further significant feature of these ketones is their occurrence in sediments deposited below the calcite compensation depth (Brassell *et al.* 1980), suggesting that organic geochemical data may perhaps provide palaeoenvironmental information in instances where micropalaeontological studies and techniques requiring calcareous microfossils, such as $\delta^{18}O$ measurements, cannot.

Diagenetic effects

During sediment diagenesis the organic constitutents inherited from biological sources are gradually transformed by various processes that modify their structures. Such processes are systematic rather than random (Mackenzie *et al.* 1982) and, following an initial phase of diagenesis where microbial activity plays a significant role, are largely determined by increases in temperature aided by the catalytic effect of clays. The detailed diagenetic pathways that link precursor biological lipids with their more stable geochemical products and their direct relationship to sediment maturity are becoming increasingly apparent. Hence, whereas specific fossil assemblages occur and characterize sediments of different ages, molecular fossil distributions are diagnostic of sediment maturity. For example, the original biological sterols found in recent and immature sediments are transformed, via various intermediate sterenes, into steranes which initially retain the structural configuration of their sterol precursors until this feature is replaced by adoption of the thermodynamically most stable configuration with a further increase in maturity (Mackenzie *et al.* 1982). Downhole trends in the changes in composition of these molecular fossils are therefore seen to occur more readily at sites of higher geothermal gradient (Fig. 8). More substantial and more rapid changes in the steroidal compounds are therefore observed at DSDP Site 467 in the San Miguel Gap than at Site 440 in the Japan Trench. The application of this molecular approach to stratigraphy is in its infancy, but its potential is evident from consideration of the great diversity of molecular species in sediments.

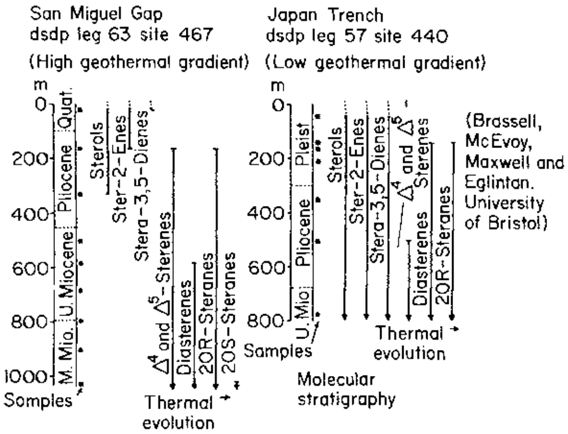

FIG. 8. Molecular stratigraphy of two DSDP sites: San Miguel Gap (467) and the Japan Trench (440).

S.C.B.

Conclusions

A comparison of the various dates for the Oligocene–Miocene boundary is given in Fig. 9 and although there have been wild fluctuations in dates ranging from 22.5–25.0 Ma, there seems to be a recent agreement between the glauconite ages of 23.0–23±1 (Odin *et al.* 1978; Kreuzer *et al.* 1980) and that of 23.7 by Berggren *et al.* in Part 2 based on palaeomagnetic stratigraphy and radiometric ages.

A similar pattern of fluctuation followed later by closer agreement can also be seen for the age of the world-wide *Orbulina* datum-plane in the lower part of the Middle Miocene (Fig. 10). The first evolutionary appearance of this well documented genus of planktonic foraminifera now seems to be between 15–16 Ma and Berggren *et al.* (in Part 2) have it at 15.0 Ma.

As more data become available for boundaries and datum-planes from various parts of the world, there is a tendency for the durations of the boundaries to be reduced to about 1 Ma in the Miocene with progressively shorter durations towards the Pliocene–Holocene. Eventually, these boundary dates will have to be related to the stage stratotypes or if this is not possible then extra boundary stratotypes will have to be chosen. It is concluded that if progress is to be maintained into the calibration of the Neogene then the next steps will have to determine the exact position and ages of the major boundaries in the stratotypes by boundary commissions of the IUGS; at present we only know the approximate ages of these boundaries in the stratotype sections.

D.G.J.

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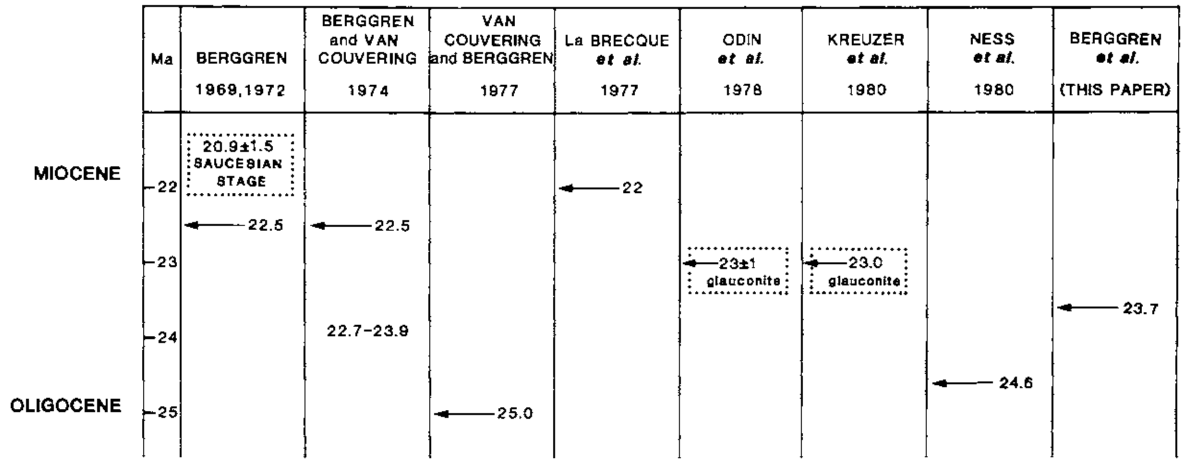

FIG. 9. Various ages for the Oligo–Miocene boundary; radiometric ages are shown in boxes; others are based on estimates from magnetostratigraphy and radiometric dates.

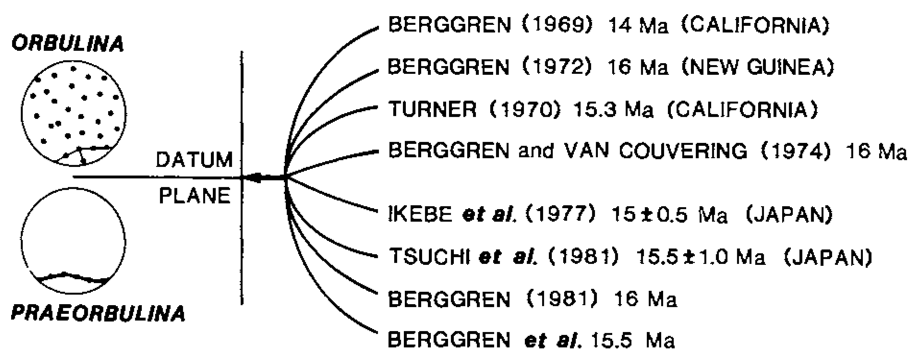

FIG. 10. Radiometric ages for the *Orbulina* datum-plane (first 6) plus two other estimates by Berggren (1981) and Berggren *et al.* in Part II (Editor's note: the age of 15.5 Ma for Berggren *et al.*, Part II, should read 15 Ma).

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The Neogene: Part 2
Neogene geochronology and chronostratigraphy

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SUMMARY: We present a revised Neogene 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: Oligocene/Miocene (Chron C6CN): 23.7 Ma; Miocene/Pliocene (slightly younger than Gilbert/Chron 5 boundary): 5.3 Ma; Pliocene/Pleistocene (slightly younger than Olduvai Subchron): 1.6 Ma.

Changes to the marine time-scale are relatively minor in terms of recent and current usage except in the interval of the middle Miocene where new DSDP data reveal that previous correlations of magnetic anomalies 5 and 5A to magnetic polarity Chrons 9 and 11, respectively, are incorrect. Our revized magnetobiostratigraphic correlations result in a 1.5–2 m.y. shift towards younger magnetobiochronologic age estimate in the middle Miocene. Radiometric dates correlated to bio- and magnetostratigraphy in continental section generally support the revized marine magnetobiochronology presented here. Major changes, however, are made in marine–non-marine correlations in the Miocene in Eurasia which indicate African–Eurasian migrations through the Persian Gulf as early as 20 Ma. The 12.5 Ma estimate of the *Hipparion* datum is supported by recent taxonomic revisions of the hipparions and magnetobiostratigraphic correlations which show that primitive hipparions first arrived in Eurasia and North Africa at *c.*12.5 Ma and a second wave in the tropics (i.e. Indian and central Africa) at *c.*10 Ma.

Neogene geochronology and chronostratigraphy:

The large issues in Neogene biostratigraphy and geochronology have been the subject of a number of papers during the past decade (*i. al.*, Berggren 1971, 1972, 1973, 1981; Berggren & Van Couvering 1974, 1978; Van Couvering & Berggren 1977; Ryan *et al.* 1974; Steininger 1977; Steininger *et al.* 1976; Steininger & Papp 1979; Ikehe *et al.* 1972, 1977; Tsuchi, (ed.) 1981). The common thread connecting these papers is the effort to continually improve upon Neogene chronology through an evaluation and integration of new data, predominantly from the fields of planktonic biostratigraphy, magnetostratigraphy and radiochronology. Recently, in a paper prepared for the International Workshop on Pacific Neogene Biostratigraphy by IGCP Project 114 (Biostratigraphic Datum Planes of the Pacific Neogene), one of the authors (Berggren 1982) reviewed the status of Neogene geochronology, placing particular emphasis on the various methodologies used in constructing magnetostratigraphic chronologies and the constraints placed upon alternate schemes by palaeobiologically controlled radiochronology.

Even in the short time since that review was completed, a considerable amount of new data has become available from magneto-biostratigraphic studies on Hydraulic Piston Cores (HPCs) taken by the Glomar Challenger, primarily in the North and South Atlantic; and also from integrated radiochronologic and magnetobiostratigraphic studies on mid-Tertiary terrestrial deposits. The magnetostratigraphic identification of several radiometrically dated mid-Tertiary terrestrial beds has provided us with much needed calibration points for the formulation of an improved Cenozoic time-scale (Prothero *et al.* 1982).

In this paper we shall outline the methodology used in devising a new, improved Cenozoic time-scale and then present a general discussion of the magnetobiochronology of Neogene chronostratigraphy. A detailed discussion of Neogene chronostratigraphic units will not be presented here; interested readers may find tbis information presented in adequate detail in Berggren (1971), Berggren & Van Couvering (1974), Van Couvering & Berggren (1977), and Ryan *et al.* (1974).

Neogene magnetochronology

As for many other geological phenomena, our knowledge of geomagnetic polarity reversals is better known for the most recent times, in this case encompassing the Neogene. After confirmation of geomagnetic reversals and the development of a magnetic polarity time-scale for about the past 4 Ma using measurements on discrete lava flows, it has been possible to extend the record of geomagnetic reversals further back in time by extrapolation using various methods and assumptions. Magnetostratigraphic investigation of sedimentary and volcanic sequences and analysis of marine magnetic anomalies over active sea-floor spreading ridge systems have provided abundant, mutually supportive evidence for geomagnetic polarity reversal history. In the process, the chronostratigraphy of the Cenozoic, particularly the Neogene, has been intricately tied to the development and use of the polarity time-scale.

The first precise knowledge of geomagnetic reversal history was obtained from the radiometric-age distribution of basaltic lavas of normal (in the same sense as the present day field) and reversed polarity (Cox *et al.* 1963; 1964). It was soon established that geomagnetic reversals do not occur regularly but define polarity intervals with durations ranging from about 50 ky to over 1000 ky (Cox 1969). The irregularity in polarity duration seems to be a characteristic feature of geomagnetic reversals, providing a signature which allows a reversal sequence to be correlated and identified. Normal and reversed polarity intervals have no intrinsic properties that allow them to be distinguished from intervals of like polarity; only the relative lengths of polarity intervals are diagnostic for their identification. Thus reference to the time-scale of geomagnetic reversals should be considered in an ordinal sense. Only when the geomagnetic reversal sequence is tied to the cardinal geological time-scale does it achieve its great potential in chronostratigraphy.

The current status of the geomagnetic reversal time-scale derived from radiometrically-dated lavas has been reviewed recently by Mankinen & Dalrymple (1979). In the perspective of geological time, the transition from one polarity state to another occupies an instant, on the order of 5 ky, and is globally synchronous. Because of inherent analytical errors in radiometric age determinations, typically a few percent in this time range, there are inconsistencies in the age-polarity distribution around what is supposed to reflect a simple transition from one polarity state to another. For example, a basalt with normal polarity may have a radiometric date that falls within an age population having predominantly reversed polarity. A statistic has been devised (Cox & Dalrymple 1967) to determine the best age estimate of a polarity reversal, wherein these inconsistencies are minimized by assuming that they are due to dating errors and not to additional reversals within a relatively small time window. This technique has been applied to estimate the ages of major subdivisions, now referred to as chrons (Anonymous 1979), in the youngest part of the polarity time-scale. According to Mankinen & Dalrymple (1979) the best estimates (using revised K-Ar decay constants) for chrons in the interval 0–4 Ma are:

Brunhes/Matuyama:	0.73Ma
Matuyama/Gauss:	2.47 Ma
Gauss/Gilbert:	3.40 Ma

The presence of shorter polarity intervals or subchrons within these chrons is also detected in the radiometric-polarity data set. However, the data are usually neither sufficiently dense nor are the age determinations precise enough to allow good estimates of the age of the reversals that bound the subchrons. An exception reported recently is for the most recent subchron, the Jaramillo, for which radiometric age estimates of 0.91 and 0.98 Ma have been obtained (Mankinen *et al.* 1980).

The inability to resolve subchron units becomes more acute as the age increases beyond about 5 Ma to the extent that even the longer chrons are difficult to resolve. This limitation results from the fact that analytical errors stay relatively constant as a percentage of calculated age and thus represent an increasingly large error in absolute terms farther back in time. Dalrymple *et al.* (1967) have shown that given even an infinite number of radiometric date-polarity data, where the dates have an uncertainty of 3%, the polarity structure beyond about 4 or 5 Ma cannot be constructed with any degree of confidence if the polarity intervals are about 1 Ma and less. This severe limitation stems largely from the total dependence on the age dates to determine both absolute age and relative age since the basalts are not otherwise physically related or ordered in some sequence.

The refinement of the 0–4 Ma magnetic reversal time-scale and its extension back in time are thus dependent on palaeomagnetic study of rock units in continuous sequence, either stratigraphic (magnetostratigraphy) or lateral (sea-floor spreading marine magnetic anomalies). The rate of the recording mechanism (i.e. sedimentation or sea-floor spreading) can be determined by correlatioin to the radiometrically-determined ages of the youngest chrons; ages of other polarity reversals can then be estimated by interpolation or extrapolation.

For several reasons, marine magnetic anomaly data provide the best, most reproducible and extended record of geomagnetic polarity history. A large number of track lines exist which allow detailed cross-correlation of the anomaly signature on a global basis. The continuity and completeness of the marine magnetic anomaly record of reversals is demonstrated by the observation that over all spreading systems, essentially the same anomaly pattern is found, differing primarily by a proportionality constant that reflects formation of oceanic crust at different spreading rates. If segments of reversal history are missing, then global hiatuses in sea-floor spreading are required, interruptions that are also not evident in the sea-floor depth vs. age relationships accounted for by simple cooling models (Parsons & Sclater, 1977).

In contrast, there are relatively few long magnetostratigraphic sections available to allow detailed checks in their completeness and both changes in rates of deposition and hiatuses are not infrequent in sedimentary sequences. The prime use of magnetostratigraphy has been to provide corroborating evidence for reversals and a framework for correlation of biostratigraphic events into the geomagnetic polarity sequence derived from marine magnetic anomaly data. For the Neogene in particular, the magnetic reversal time-scale provides critical information for chronostratigraphy by such correlations.

Magnetostratigraphy

The magnetostratigraphy of late Miocene to Pleistocene* deep-sea sediments has been well documented and is by now well known (Hays *et al.* 1969; Foster & Opdyke 1970; Opdyke 1972; Saito *et al.* 1975). Since this interval overlaps the age range of the radiometric reversal time-scale, age estimates for biostratigraphic and palaeooceanographic events recorded in the sediments can be determined by interpolation rather precisely after correlation of magnetozones to the radiometrically dated magnetochrons. For example, the time-scale for Pliocene–Pleistocene oxygen isotope stratigraphy (Shackleton & Opdyke 1973, 1976, 1977), and the best current age estimate for the Pliocene–Pleistocene boundary (Haq *et al.* 1977) are based largely on magnetochronology. Age estimates of the Jaramillo Subchron (Opdyke 1969) and the Olduvai Subchron (Opdyke & Foster 1970) have also been obtained using the sedimentary record.

Extension of the magnetostratigraphic record near to the base of the Neogene has been accomplished in conventional piston cores of deep-sea sediments (Foster & Opdyke 1970; Theyer & Hammond 1974a, b; Opdyke *et al.* 1974), providing direct biostratigraphic correlation to magnetozones. In order

* We do not use the term Holocene in the formal sense here, preferring to consider it rather as merely an interglacial period.

to place calcareous plankton biostratigraphy for the entire Neogene into a magnetostratigraphic framework, Ryan *et al.* (1974) used a complex correlation network involving magnetobiostratigraphic data from DSDP sites, conventional piston cores and exposed sedimentary sections. In the process, the European Neogene stages could be correlated to the marine magnetic anomaly sequence. They also recognized that the geomagnetic polarity time-scale, as developed from analysis of the magnetic anomaly signature of reversals, could be used to estimate ages for stage and system boundaries. In effect, they assumed that over appreciable time intervals, sea-floor spreading rates are more apt to be constant than sedimentation rates. Consequently the geomagnetic polarity time-scale should provide an improved basis for linear interpolation between levels of known or inferred age. The validity of these age estimates of course depends upon both the correlation to the geomagnetic polarity time-scale and how well the polarity time-scale is calibrated.

A considerable quantity of additional magnetobiostratigraphic data from both marine and terrestrial sequences have become available since the work of Ryan *et al.* (1974). This information allows further refinements in correlations to the standard geomagnetic polarity sequence and an assessment of these data is presented later in this paper (see also Appendix). The geomagnetic polarity sequence has also received additional attention and several revised versions have appeared. Since the chronology of geomagnetic polarity reversals now has a very direct bearing on Neogene chronostratigraphy, we present a new, revised polarity time-scale that incorporates what we believe are the best available calibration age data. The major differences among various versions of the polarity time-scale occur in the Palaeogene and are discussed in the companion paper by Berggren *et al.* (this volume). An outline of our methodology is presented here with emphasis on the Neogene.

Regarding polarity interval nomenclature, it should be noted that at least three schemes are in effect for the Neogene. Although the four most recent chrons are named after eminent geomagnetic researchers (Brunhes, Matuyama, Gauss, and Gilbert), this system of nomenclature was impractical for earlier chrons due to their large number. Hays & Opdyke (1967) introduced an identification scheme for magnetostratigraphy in which about 1 Ma intervals of predominantly normal or reversed polarity are numbered sequentially from 1 (the Brunhes, although the first four chrons usually retain their familiar names); odd numbers refer to predominantly normal intervals and even numbers reversed polarity intervals. Subchrons are then identified by letter suffixes added to the chron number. Opdyke *et al.* (1974) and Theyer & Hammond (1974a, b) extended this scheme to Chron 23, near the base of the Neogene. This sequential numbering scheme for chrons has not been further extended into the Paleogene. Instead, it has been found most useful to adapt the well-established numbering scheme of marine magnetic anomalies originally introduced in Pitman & Heirtzler (1966) and Heirtzler *et al.* (1968). For example, in the system proposed by LaBrecque *et al.* (1983), a chron is defined as extending from the youngest reversal boundary of a numbered anomaly to the youngest boundary of the next older numbered anomaly; the chron is named for the correlative magnetic anomaly number, with the letter 'C' (for chron) prefixed to avoid confusion with the pre-existing Neogene chron numbering nomenclature. A similar scheme has been also suggested by Cox (1982). Thus the interval of predominantly normal polarity corresponding to the Gauss Chron (Cox *et al.* 1964) (correlated to marine magnetic anomaly 2A) can be also referred to as Chron 3 (Hays & Opdyke 1967) or as Chron C2AN (LaBrecque *et al.* 1983), where the suffix 'N' refers to the normal polarity interval(s) associated with the magnetic anomaly.

Difficulties arose in attempts to correlate the chron subdivisions (*sensu* Hays & Opdyke 1967) with the standard marine magnetic anomaly sequence, particularly around the interval of Chron 9 and older. Foster & Opdyke (1970) originally correlated Chron 11 with Chron C5N (normal polarity interval represented by magnetic anomaly 5). Theyer & Hammond (1974a) discussed the problem of correlation of the deep-sea core magnetostratigraphy to the marine magnetic reversal sequence. On the basis of a variety of biostratigraphic arguments, they proposed that Chron 9 is instead correlative to chron C5N and this correlation scheme was used by Ryan *et al.* (1974) and in most subsequent Neogene time-scales (Berggren 1974; Van Couvering & Berggren 1977; Berggren & Van Couvering 1978). It has now become apparent from recently available magnetobiostratigraphy that this correlation (Chron 9 = Chron C5N) cannot be easily accommodated and in fact the original correlation (Chron 11 = Chron C5N) is preferable. The magnetobiostratigraphic arguments supporting this change are presented below under the Miocene.

Figure 2 shows our revised magnetostratigraphic chron correlations to the marine magnetic sequence, down only to Chron 11. We have not found it possible to recorrelate the magnetostratigraphic chrons older than Chron 11 to the standard marine magnetic reversal sequence in any convincing manner. An exception is the correlation of anomaly 5B (= C5BN) and Chron 15 owing to the fact that the position of Chron 15 is the same in both of the alternative correlations (anomaly 5 = Chron 9 vs. anomaly 5 = Chron 11) discussed by Theyer & Hammond (1974a: 316, 317). One obstacle is that the magnetostratigraphic chrons were originally correlated (Foster & Opdyke 1970; Theyer & Hammond 1974a; Ryan *et al.* 1974) to the marine magnetic reversal sequence of Heirtzler *et al.* (1968). The magnetic reversal pattern between anomalies 5 and 6 had been subsequently revised substantially (Blakeley 1974) and it is in part the differences between the Heirtzler *et al.* and Blakeley versions in this interval that make the magnetostratigraphic chrons difficult to realign. In any case, the marine magnetic reversal sequence is much better defined than the magnetostratigraphic record and we suggest that future magnetobiostratigraphic correlations go directly to the marine magnetic expression of geomagnetic reversals. Accordingly, we show chron labels for the anomaly 5 correlative and older portion of the Neogene that are based on the marine magnetic nomenclature, following at this level of detail the scheme proposal by Cox (1982). The magnetostratigraphic chrons (including the labels for the four most recent chrons) are retained to Chron 11 for mostly historical reasons and to emphasize the proposed change in correlation to the sea-floor magnetic anomalies. The chron terminology derived from magnetic anomaly nomenclature, however, can be easily obtained because the polarity intervals corresponding to the numbered marine magnetic anomalies are indicated. Note that because of the aperiodic nature of polarity reversals and the different ways in which the polarity intervals were subdivided (i.e. key, easily identifiable magnetic anomalies of positive polarity vs. intervals of about 1 m.y. of predominantly normal or reversed polarity), the chron boundaries according to the different schemes usually do not coincide. Ramifications and implications of the proposed change in correlations are discussed in the Miocene section of the manuscript.

Revised geomagnetic polarity time-scale

The first extended geomagnetic reversal time-scale was presented by Heirtzler *et al.* (1968) who selected a magnetic profile from the South Atlantic Ocean as representative of geomagnetic reversal history for about the past 80 Ma. Their chronology of reversals, hereafter referred to as HDHPL68, was derived by correlation of the axial anomalies to the radiometrically-dated magnetic reversal time-scale and then by extrapolation to the oldest then recognized anomaly (anomaly 32). This twenty-plus fold extrapolation assumed that the rate of sea-floor spreading in the South Atlantic was constant over 1400 km or 80 Ma, at the rate calculated from the ridge axis (time zero) to anomaly 2A (then dated at 3.35 Ma). Despite this large extrapolation, it is now apparent that HDHPL68 was within about 10% of current age estimates for this reversal sequence, an indication that the assumption of a constant rate of sea-floor spreading over prolonged time intervals is not an unreasonable approximation.

Virtually all subsequent polarity time-scales rely to some degree on the constant spreading rate assumption; differences arise mainly from the assignment and use of additional age-calibration data for interpolation and extrapolation. The apparent continuity of the marine magnetic anomaly record also requires that the entire sequence be considered at the same time in revision of the chronology. This is because changes in one portion of the sequence will tend to propogate to other portions, or else artificial discontinuities may be introduced. Moreover, geologic stage and epoch subdivision do not have, as far as we know, any particular significance with respect to sea-floor spreading history so that it would be both difficult to justify and out of context to attempt a revision only, for example, of the Neogene portion of the Polarity time-scale. However, the magnetic anomaly sequence extending from the present spreading ridge axes is bounded at the older end by the Cretaceous quiet zone, corresponding to a prolonged interval of constant normal polarity containing few if any correlatable reversals. Consequently it is possible and convenient to analyse the late Cretaceous to Recent sequence of magnetic anomalies as a unit.

As a representative sequence of geomagnetic polarity for the late Cretaceous to Pleistocene, we use a slightly modified version of the LaBrecque *et al.* (1977) time-scale (referred to here as LKC77). Several refinements of HDHPL68 are incorporated in LKC77, most importantly for the Neogene, revisions of the polarity reversal pattern between anomalies 5 and 6 (Blakeley 1974) and from the central anomaly to anomaly 3A (Klitgord *et al.* 1975). Virtually the entire Neogene portion of the anomaly signature of geomagnetic reversals has therefore been re-examined in detail, and adjusted accordingly to reflect the best available data. In contrast, the Paleogene portion is still largely based on the relative spacings of anomalies originally determined by Heirtzler *et al.* (1968). A minor modification made to LKC77 is that the polarity interval lengths described by Blakeley (1974) are recalculated according to the original age estimate in HDHPL68 for the younger end of anomaly 5. The resulting overall sequence is thus assembled from essentially the same data as in a recent compilation by Ness *et al.* (1980).

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 obtained by averaging 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. (In fact, even LKC77 does not require sea-floor spreading in the South Atlantic at a single constant rate for the entire late Cretaceous to Recent interval.) 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 a common denominator is required to express the best estimate of a geomagnetic reversal sequence synthesized from varied sources. Because of the ordinal nature of the magnetic reversal time-scale, the units they are given in can he referred to 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 (Chron C2A) 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 radiometrically-dated reversal time-scale and a traditional calibration point in virtually all Late Cretaceous to Recent geomagnetic reversal time-scales.

(b) 8.87 Ma—Anomaly 5y (younger end of Chron C5 or Chron 11). Based on stratigraphically-controlled distribution of radiometric date-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 (younger end of Chron C12). 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 Chron C12 (Prothero *et al.* 1982).

(d) 34.6 Ma—Anomaly 13y (younger end of Chron C13). Same source as item c; radiometric date (K-Ar) on biotite in volcanic ash stratigraphically overlying normal magnetozone correlated to Chron C13 (Prothero *et al.* 1982).

(e) 49.5 Ma—Anomaly 21y (younger end of Chron C21). 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 polarity magnetozone correlated to C21 (Flynn 1981).

(f) 84.0 Ma—Anomaly 34y (Chron C34 or the end of Cretaceous Long Normal). Age estimate 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 Chron C34, in Italian limestones (Lowrie & Alvarez 1977).

The radiometric age estimate for the younger end of Chron

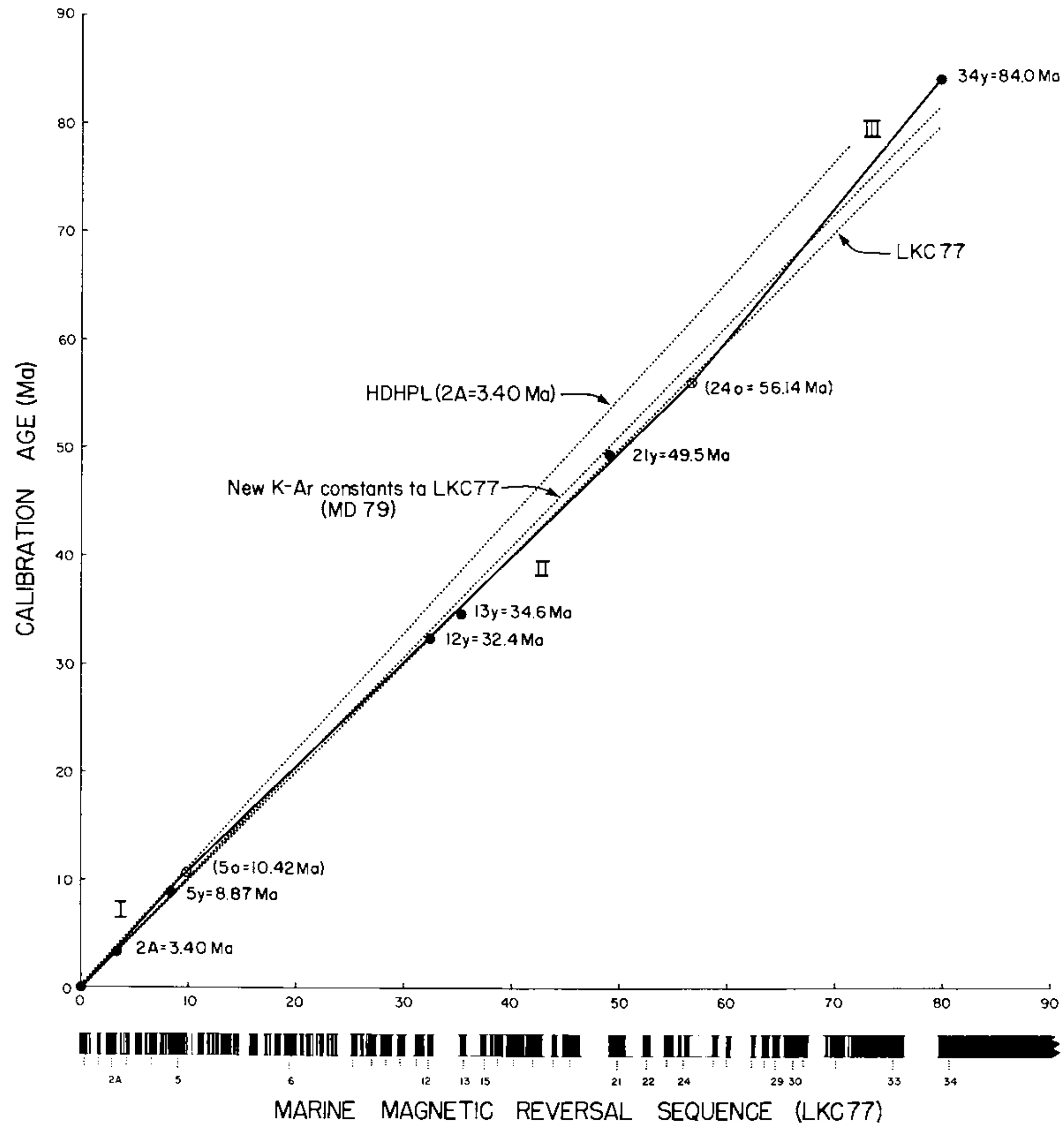

FIG. 1. Revised age calibration of marine magnetic reversal sequence from LaBrecque *et al.*, 1977 (LKC77). Solid lines are three linear apparent age-calibration age segments (I, II, and III) which satisfy calibration tie-points indicated by solid circles (Table 1). The two open circles with X's at anomalies 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 reversal 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 and Dalrymple, 1979). Anomaly numbers are indicated below bar graph of geomagnetic reversal sequence (filled for normal, open for reversed polarity).

C5 (8.87 Ma, item b, above) is very near to the age extrapolated for this anomaly in HDHPL68 (8.92 My, using the revised 3.40 Ma date instead of 3.35 Ma from Chron C2A). This is a strong indication that the original HDHPL68 timescale provides a good chronologic framework for polarity reversals at least as far back as this time. Beyond Chron C5, 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 Chron C5 (item b) and the top of Chron C12 (item c). This new trend, however, cannot also accommodate the calibration tie-point for Chron C34 (item f) and a change to another relationship must therefore occur somewhere between Chron C21 (item e) and Chron C34 (item f).

A minimum of two changes in the relationship between calibration age and apparent age in modified LKC77 appear to be required to satisfy this set of data. Consequently, we divide the geomagnetic reversal sequence into three linear calibration age-apparent age segments with the inflection points placed at anomaly 5 and at anomaly 24 for reasons discussed in Berggren *et al.* (this volume). 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 plus the origin. This trend is extrapolated to derive an estimated age of 10.42 Ma for the older end of anomaly 5. 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 derived age for the base of anomaly 5. Segment III is simply an interpolation between the estimated age of anomaly 24, derived by extrapolation of segment II, and the inferred age of anomaly 34 (item f).

Ages for magnetic polarity intervals or chrons are cal-

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	65.50–66.17	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

culated according to the linear equations of these three segments; a tabulation of these ages is presented in Table 1.

Direct bio-magnetostratigraphic correlations are the basis for comparison of the revised magnetochronology and biochronology. These correlations are discussed for the Paleogene in Berggren *et al.* (this volume) and for the Neogene below, and the resulting magnetobiochronology is shown in Fig. 3. In these correlations, the base of the Neogene is placed within Chron C6CN, at an estimated magnetochronological age of 23.7 Ma. This age estimate is not totally independent of biochronology because calibration data in c, d, and e, which define segment II, rely in part on biostratigraphic correlation. Calibration data (origin and items a and b) which define the initial segment (I), however, are totally independent of biostratigraphic control: the 3.40 Ma radiometric age for the base of the Gauss Chron and the 8.87 Ma radiometric age for the top of Chron C5 (Chron 11) are based on lavas and are correlated to the marine magnetic polarity sequence primarily by the palaeomagnetic reversal signature. It is ironic that in the Paleogene and older periods, biochronological data tend to be used for age calibration of the magnetic reversal chronology (e.g., Lowrie & Alvarez 1981) whereas at least in the younger half of the Neogene, magnetochronology is often the basis for biochronological estimates. The potential for circular reasoning in the construction of an internally consistent geological time-scale should be kept in mind.

The Oligocene–Miocene boundary

'The Oligocene–Miocene boundary is one of the most difficult and controversial boundaries in the Tertiary' (Jenkins, 1966: 11). Seventeen years later little has changed if one considers the continuing polemics which take place at the quadrennial meetings of the Committee on Mediterranean Neogene Stratigraphy, those of the IGCP 114 (Pacific Neogene Datum Events), and those of the IUGS Working Group on the Paleogene–Neogene Boundary (Cati, (ed.) 1981) as well as in the biostratigraphic literature in general.

The problem lies, primarily, in the fact that no major, global regression event(s), such as those that occur at the Cretaceous–Tertiary, Eocene–Oligocene, or Miocene–Pliocene boundaries is recognized at the Oligocene–Miocene boundary, which coincides with a period of moderate to high sea-level stand. Thus, dramatic faunal, climatic and palaeo-oceanographic changes do not attend the base of the Miocene, and for this reason it has been difficult to determine definitive biostratigraphic criteria in recognizing and correlating the base of the Miocene and, by definition, the base of the Neogene, on a global scale.

Zonation by calcareous plankton of upper Oligocene to lower Miocene marine sedimentary sections has been the subject of considerable controversy since the basic zonations were developed by Bolli (1957) and Martini (1971) on planktonic foraminifera and calcareous plankton, respectively. We shall discuss the problems associated with recognizing this boundary in terms of both groups of microfossils.

The zonal scheme proposed by Blow (1969), not unlike that proposed earlier by Bolli (1957) and Bolli & Bermudez (1966), is based upon the following sequence of biostratigraphic events/ranges (in stratigraphic order):

4. FAD (first appearance datum) of *Globigerinoides primordius*.
3. FAD of *Globorotalia kugleri*.
2. continued (partial) range of *Globigerina ciperoensis* or *G. angulisuturalis*.
1. LAD (last appearance datum) of *Globorotalia opima opima*.

Our own observations at DSDP Site 516 (Rio Grande Rise, South Atlantic) indicate the following succession of events (in stratigraphic order):

4. FAD *Globorotalia kugleri*
3. continued (partial) range of *Globigerina angulisuturalis*
2. FAD *Globigerinoides primordius*
1. LAD *Globorotalia opima*

Note the inversion between the FAD's of *G. primordius* and *G. kugleri*. Similar observations have been made by others on the relationship of the FAD of *Globigerinoides primordius* relative to other taxa.

Biostratigraphic data gathered over the past decade,

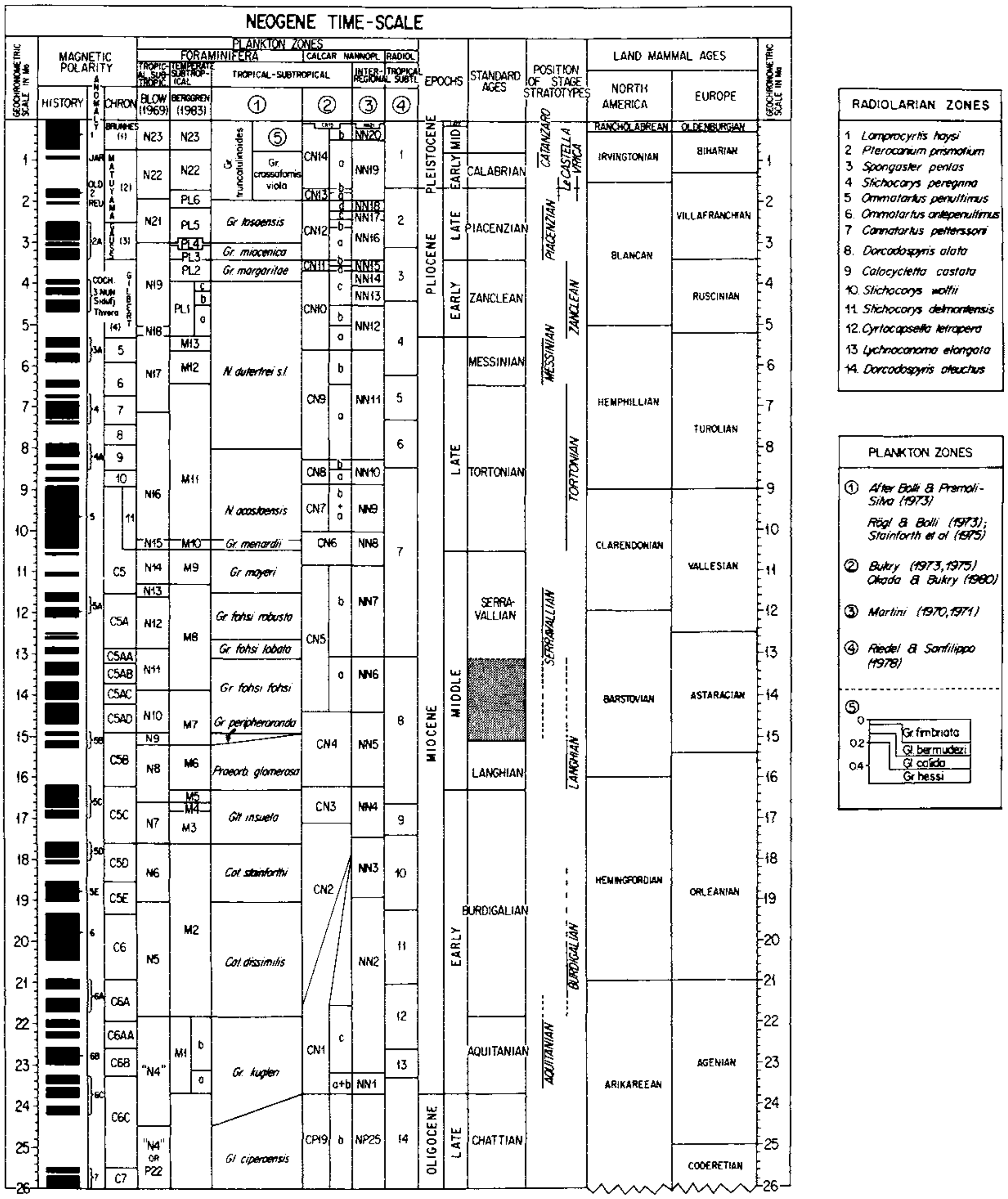

FIG. 2. Neogene geochronology. The geochronologic scale at the margins of the figure is derived from the magnetic polarity chronology which is in turn derived from palaeontologically and/or palaeomagnetically controlled radiometrically dated calibration points in the late Neogene, early Oligocene, middle 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 (see compilation in Vincent, 1981a, b; Appendix II, this paper). 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.

primarily in connection with the Deep Sea Drilling Project, has provided additional information, but often it appears to be of a conflicting nature. Whereas the FAD of the genus *Globigerinoides* (as *G. primordius*) was thought to be a reliable marker for the Oligocene–Miocene boundary by many workers, the apparent time-transgression of its FAD relative to other biostratigraphic markers led Lamb & Stainforth (1976) to suggest that it be abandoned as a precise index form for the Oligocene–Miocene boundary.

In his redefinition of Zone N4, Blow (1969: 222–225) replaced the concept of a *Globorotalia kugleri* Total-range Zone with a Concurrent-range Zone (with *Globigerinoides primordius*). He was of the opinion that the FAD of *G. kugleri* occurred within his Zone N3 (= P22), and his (revised) Zone N4 was based on the extension of the range of *G. kugleri* into the time following the FAD of *G. primordius*.

The recognition of *Globigerinoides primordius* in pre-Aquitanian levels in the Aquitaine Basin of SW France (Scott 1972) and the Rhone Valley (Anglada 1971) as well as in the Codrington College section of Barbados (in calcareous nannoplankton Zone NP25 and above the LAD of *Globorotalia opima*; Van Couvering & Berggren 1977) had the effect of reducing the span of Zone NP22. The presumed overlap of *G. primordius* and *G. opima* in the Ashmore Reef No. 1 well, north-west Australia (Chapronière *in*: Shafik & Chapronière 1978; Chapronière 1981), if validated elsewhere, would effectively eliminate Zone P22 from the planktonic foraminiferal hagiography. However, we remain sceptical of the suggested overlap in these taxa (the specimens of *G. opima* are not unequivocal) and recent magnetobiostratigraphic studies in the South Atlantic have indicated a distinct separation between these two taxa (LAD *G. opima* associated with Chron C9N; FAD *Globigerinoides* associated with Chron C6CR. The FAD of *G. primordius* earlier than that of *G. kugleri* and, in fact, virtually coincident with the LAD of *G. opima opima* and within the later part of the ranges of the calcareous nannoplankton taxa *Reticulofenestra bisecta*, *Cyclicargolithus abisectus* and *Zygrhablithus bijugatus* has been recorded by Báldi-Beke *et al.* (1978) in the Piedmont Basin of NW Italy. Nevertheless, adequate documentation (in the form of unequivocal illustrations) is still lacking.

Thus the fact that the genus *Globigerinoides* (*G. primordius*) appears before the FAD of *Globorotalia kugleri* effectively eliminates the utility of Zone N4 (as emended by Blow 1969: 223, who had the order of appearance of the two nominate taxa reversed). For if the primary definition of Zone N4 is the FAD of *G. primordius* (Blow 1969), N4 is no longer a Concurrent-range Zone and there is a considerable gap between the sequential FADs of *G. primordius* (within either Zone P22 of Blow 1969 – and perhaps even earlier – and NP25; Martini 1971) and *G. kugleri* (base Zone N4, Blow, 1969, and NN1; Martini 1971).

The absence of a clearly defined boundary stratotype section for the base of the Aquitanian coupled with the lack of distinctive faunal and floral elements useful in regional correlation is responsible for the continued polemics surrounding the biostratigraphic correlation of the Oligocene–Miocene boundary.

In order to determine which of the biostratigraphic criteria among the calcareous plankton are most suitable for recognition and correlation of the Oligocene–Miocene boundary we must consider first the biostratigraphic relationships that exist in the stratotype section(s) and in the deep sea where richer faunas and floras occur.

Studies of the planktonic foraminiferal faunas of the stratotype (and nearby area) Aquitanian-Burdigalian sections have spanned the past quarter of a century. An analysis of the various papers that have appeared during this time suggests that the stratotype limits of the Aquitanian Stage lie within the concurrent ranges of *Globorotalia kugleri*, *Globigerinoides primordius*, *Globigerina woodi* and *Globoquadrina dehiscens* (Jenkins 1966; Pujol 1970; Poignant & Pujol 1976; Müller & Pujol 1979). The stratotype limits of the Burdigalian Stage lie within the concurrent range of *Globigerinoides altiaperturus* and *Globoquadrina dehiscens* (Jenkins 1966; see also Anglada 1971; Poignant & Pujol 1976, 1979; Müller & Pujol 1979). *Globigerinoides altiaperturus* first appears in the Burdigalian Stage (Jenkins 1966) and in Zone N5, above the LAD of *Globorotalia kugleri* (Chapronière 1981).

A consideration of the above studies in the stratotype area suggests, then, that the Oligocene–Miocene boundary (base of the Aquitanian Stage) lies somewhere within the NP25–NN1 interval and within the range of *Globorotalia kugleri* (= N4 as a total range Zone).

Let us turn our attention now to a consideration of the way in which biostratigraphers have viewed the Oligocene–Miocene boundary particularly as a result of deep sea drilling. In the discussion ahead the reader is referred to Tables (4) and (5) (Appendix I) in which the relationships between various biostratigraphic datum events in the planktonic foraminifera and calcareous nannoplankton, respectively, are shown.

The many criteria used in deep-sea studies to delineate the Oligocene–Miocene boundary are expressed both in the calcareous nannoplankton and planktonic foraminifera.

1. *Calcareous nannoplankton*: the Oligocene–Miocene boundary has been placed according to each of the following criteria:

I. NP25/NN1 boundary;
II. A level within Zone NN1 (defined by the LAD of *Reticulofenestra* (*Dictyococcites*) *bisecta*); or
III. The NN1/NN2 boundary (defined by the FAD of *Discoaster druggii*)

In the first case, that of (I.) the NP25/NN1 (or its equivalent) boundary, this level in itself is determined by various criteria, as follows:

	Remarks
(a) LAD *Reticulofenestra bisecta* (+ *R. scissura*)	Wide geographic distribution near shore and open ocean.
(b) LAD *Sphenolithus ciperoensis*	Tropical-subtropical species.
(c) LAD *Helicosphaera recta*	Predominantly hemipelagic; usually rare.
(d) LAD *Reticulofenestra scrippsae* (+ *R. hesslandi*)	Predominantly high latitude; near shore and open ocean.
(e) LAD *Zygrhablithus bijugatus*	Predominantly high latitude, epicontinental — near shore.

The LADs of *S. ciperoensis*, *R. bisecta* and/or *H. recta* are generally separated by a distinct stratigraphic interval (sites 149, 151, 167, 238, 289, 357, 363, 445, 516F, 522); less frequently they are found to occur (virtually) simultaneously (as at sites 74, 236, 296). The LADs of *S. ciperoensis* and *C. abisectus* have been recorded simultaneously (as in Site 171) as well as those of *R. bisecta* and *C. abisectus* (Site 234). In one instance (Site 386) the virtually simultaneous extinction (LAD) of three of the nominate taxa for the NP25/NN1 boundary (*S. ciperoensis*, *R. bisecta* and *H. recta*) is recorded with the FAD of *D. druggii* (nominate taxa of Zone NN2) and the LAD of *C. abisectus* (nominate taxa of the lower subzone of Zone NN1). In this instance one may ask where is Zone NN1 or better, *what* is Zone NN1?

2. *Planktonic foraminifera*: in deep-sea studies, each of the following criteria have been used to identify the Oligocene–Miocene boundary:

(a) FAD *Globigerinoides primordius*
(b) FAD *Globorotalia kugleri*

(c) LAD *Globorotalia kugleri*
(d) FAD *Globoquadrina dehiscens**

Comparing the zonal boundary definitions, those of the planktonic foraminifera appear to be more consistent than within the calcareous nannoplankton. However, as we have seen above, the temporal uncertainty due to the variations in relative position of the biostratigraphic datum events near the boundary is essentially comparable in the two groups of calcareous microfossils (*c*. 2–3 m.y.). The different evaluations permitted by this uncertainty are legion.

The NP25–NN1 boundary (based on LAD of *R. bisecta*) is found to lie close to that of the FAD of *G. kugleri* in some instances (sites 74, 296, 516F). The NP25–NN1 boundary (based on the LAD of *S. ciperoensis*) is close to that of the FAD of *G. kugleri* at Site 214.

An Oligocene–Miocene boundary based on the FAD of *Globigerinoides* (*G. primordius*) has heen placed rather consistently in Zone NN1 (sites 149, 214, 238, 292, 296, 354, 357, 362, 363, 386), more rarely within Zone NN2 (Site 289). However, the FAD of *Globigerinoides primordius* is known to occur within Zone NP25 at some sites (e.g. 516) and has been documented to occur within this zone in various land sections and deep sea cores elsewhere.

The Oligocene–Miocene boundary based upon the FAD of *Globoquadrina dehiscens* has been observed either below the FAD of *D. druggii* (sites 296, 357, 522), above it (354, 356, 362, 363), or coincident with it (516F). The discrepancy here may well lie with the difficulty of determining in a consistent manner the FAD of *G. dehiscens*.

The relationship of the LAD of *C. abisectus* to the FAD of *D. druggii* is seen to be inconsistent also, occurring below in some instances (sites 296, 354, 357) as would be expected if *C. abisectus* is the nominate taxon of a lower subzone of Zone NN1, but above *D. druggii* in other instances (sites 356, 362, 363, 386, and 516F). Indeed, the taxon *C. abisectus* has been recorded up to the middle Miocene (Zone NN6).

It will be readily seen from the above that the inability to distinguish clearly between zones NP25 and NN1 may he responsible, in part, for the correlation of the FAD of *G. kugleri* to Zone NP25 in some, and to NN1 in other, deep sea drilling reports. Last occurrences (extinctions) are less than useful criteria in defining calcareous nannoplankton zones (unless done in a quantitative manner) owing to the problems of reworking. Alternatively the failure to distinguish between the components of the *Globorotalia kugleri* plexus (*G. kugleri, G. mendacis, G. pseudokugleri*) may also contribute to inconsistent biostratigraphic determinations of the 'FAD' of *G. kugleri* and the base of Zone N4.

Where then does the base of the Aquitanian (and, by extension, the Oligocene–Miocene boundary) lie with respect to the established calcareous plankton zones and/or datum levels discussed above? Inasmuch as we have shown that *Globigerinoides* appears at least as early as *Globorotalia kugleri* and that *Globoquadrina dehiscens* occurs *within* the range of *G. kugleri*, we would suggest that the base of the Aquitanian Stage should be relocated at a level immediately *above* the base of the present stratotype section (which is characterized at its base by an unconformity) and that it corresponds to the FAD of *Globorotalia kugleri* (= base N4 as a total range Zone) and to the NP25–NN1 Zone boundary (as defined by the LAD of *Reticulofenestra bisecta*). The FAD of *Globoquadrina dehiscens* appears to occur only slightly subsequent to that of *G. kugleri* and would serve as a secondary criterion in recognizing the approximate position of the Oligocene–Miocene boundary. Indeed Srinivasan & Kennett (1983) have chosen the FAD of *Globoquadrina dehiscens* as the definitive criterion in recognizing the Oligocene–Miocene. The boundary as recognized here corresponds to the lower part of Chron C6CN (*c*.23.7 Ma); the boundary as recognized by Srinivasan & Kennett is approximately correlative with upper part of Chron C6CN, *c*.23.2 Ma (see Table 6, Appendix II). The relationship is not unlike that which exists between *Globorotalia truncatulinoides, Globigerinoides obliquus extremus* and *Globigerinoides fistulosus* — taxa with sequential FAD's or LAD's that bracket the Pliocene–Pleistocene in different latitudes. The base of the Burdigalian Stage might then correspond closely to the N4/N5 boundary (Blow 1969) and the NN1/NN2 boundary (Martini 1971).

This conclusion appears to be consistent with those of some other authors. For instance, on the basis of a comparative study of the Soustons well (subsurface Aquitaine Basin) Vigneaux *et al.* (1970) suggested that the base of the Aquitanian coincides approximately with the appearance of *Globorotalia kugleri*. On the basis of a quantitative comparative study of the genus *Globigerinoides* Scott (1968, 1971) suggested that the stratotype Aquitanian straddled the *Globorotalia kugleri* and *Catapsydrax dissimilis (= N4–N5)* Zones.

Despite the cautionary note sounded by Blow (1969: 224) about the problem of confusion with *G. mendacis* and *G. pseudokugleri*, we believe that *G. kugleri* is basically a distinct taxon that is relatively easy to identify in a consistent manner (see in this context Chapronière 1981; p. 126, fig. 12).

We believe that part of the reason for the inconsistency in the data concerning the initial appearance of the genus *Globigerinoides* may be due to the effects of dissolution in deep sea cores and of geographic distribution in land sections. A further problem in the cross correlation of zones and datum levels based on calcareous plankton may lie in the difficulty in distinguishing consistently between calcareous nannoplankton Zones NP25 and NN1 as we have indicated above.

Magnetobiostratigraphic correlations (see above) suggest that the Oligocene–Miocene boundary (as limited by the FAD of *Globorotalia kugleri* and the LAD of *Reticulofenestra bisecta*) is associated with lower Chron C6CN and has an estimated magnetochronologic age of 23.7 Ma. This age estimate is seen to be remarkably consistent with published radiometric data, which suggests an age of 23 Ma (Kreuzer *et al.* 1980) to 24 Ma (Ritzkowski 1982) for the Oligocene–Miocene boundary (see additional discussion in Berggren *et al.*, this volume).

Corroborative evidence comes from California. Miller (1981), in study of the Zemorrian–Saucesian interval, recalculated the K-Ar dates reported by Turner (1970) using new decay and abundance constants (Steiger & Jäger, 1977). Accordingly, the Upper Zemorrian Iversen Basalt of Point Arena would now have an average age of 23.8 Ma; lower Saucesian volcanics from the San Emigdio Mountains, 22.5 Ma; and Upper Zemorrian or Lower Saucesian Santa Cruz Volcanics, 23.7 Ma. Taken together, these dates suggest that

* Widespread in the oceans; the other two are restricted to tropical-subtropical palaeoenvironments.

the Zemorrian–Saucesian boundary, correlated earlier to the N4–N5 boundary (Van Couvering & Berggren 1977), has an age of *c.*23 Ma.

This date is too old for the N4–N5 boundary, at 21.8 Ma in our present time-scale (Fig. 2), and in addition it is even more inconsistent with offshore correlation of the Zemorrian and Saucesian to calcareous nanoplankton zones, as recently published by Crouch & Bukry (1979) and Arnal (1980). Despite their other differences, these authors agreed that in dart core samples from the California shelf, Zemorrian and Saucesian benthic foraminiferal assemblages overlap throughout the *Sphenolithus belemnos* (CN2) Zone of Bukry (1973, 1975). According to these observations, and the principle of exclusion that 'base defines boundary' in overlapping sequences, the offshore Zemorrian–Saucesian boundary corresponds to the base of the Saucesian faunas near the lower boundary of Zone CN2, where it would be calibrated at 18 Ma (Bukry 1975; see also Fig. 2).

Evidence from standard sections on shore, however, indicate that in fact the Saucesian benthic foraminiferal assemblages range down to well below the CN2 level, as the dating has suggested. In the type section at Los Sauces Creek, beds of the uppermost Zemorrian (by exclusion), lying just below and conformable with basal Saucesian, contain coccoliths that are referable to uppermost Oligocene *S. ciperoensis* (CP19) Zone, and possibly the lowermost Miocene (*C. abisectus*, CN1a Zone) according to R. Z. Poore (written communication, 1981). In sediments associated with the 23.8-Ma Iversen Basalt of Point Arena, also of latest Zemorrian age, the base of the Miocene—as correlated by coccoliths (see above) – is identified by the LADs of *Reticulofenestra bisecta* and *R. scrippsae* (given as *Dictyococcites bisectus* and *D. scrippsae*) in a low-diversity but diagnostic nanoflora collected by Miller (1981). Thus, in California, the 23 Ma Zemorrian–Saucesian boundary in standard sections can now be placed more accurately below the N4–N5 boundary and just above the Oligocene–Miocene boundary. In addition, the Oligocene–Miocene boundary is linked directly to Iversen Basalt, convincingly dated to 23.8 Ma, in remarkably close agreement with the 23.7 Ma date obtained by linking this boundary biostratigraphically to Chron C6CN.

The Miocene

A more or less standard threefold Miocene chronostratigraphic subdivision is followed here, in which the Aquitanian and Burdigalian ages represent the early Miocene, the Langhian and Serravallian ages represent the middle Miocene, and the Tortonian and Messinian ages represent the late Miocene.

The base of the Miocene (= Aquitanian) is understood to be limited by the FAD of *Globorotalia kugleri* and the LAD of *Reticulofenestra bisecta*; to be approximately coincident with the NP25–NN1 (CP19–CN1) boundary; to be associated with the early part of Chron C6CN; and to have an age estimated by magnetochronology at 23.7 Ma (see above and Appendix II). The Burdigalian stratotype has a planktonic foraminiferal fauna indicative of Zone N5 (Anglada 1971) and NN2 (*druggi*, or CN1c; Müller *In:* Bizon & Müller 1979) or NN2–NN3 (*druggi-belemnos*; Schmidt *In:* Benda *et al.* 1977; see also Van Couvering Berggren 1977). In California, stratigraphic levels referable to the Saucesian Stage belong mainly to the *Sphenolithus belemnos* (NN3 or CN2) Zone, but according to Warren (1980) its lower part may include strata that can be assigned to the *druggi* (CN1c) Subzone and perhaps the *deflandrei* (CN1b) Subzone of the *Triquetrorhabdulus carinatus* (CN1) Zone of Bukry (1973). As we have noted above, R. Z. Poore (written communication, 1981) suggest that the youngest pre-Saucesian strata at Los Sauces Creek, California are of CP19 to CN1a age. Recalculation of radiometric dates (Turner 1970) from California suggests an age in this region of 23 Ma for the Saucesian–Zemorrian boundary and of 23.8 Ma for the CP19–CN1 boundary (Miller 1981).

In terms of magnetobiostratigraphy the following points are pertinent:

1. DSDP Site 15 was drilled on anomaly 6 in the South Atlantic and the oldest sediments above basement belong to Zone N5 (Blow 1970).
2. The FAD of *Calocycletta virginis* (radiolarian) occurs just below magnetic polarity Chron 20/21 (within C6A in terminology adopted in this paper) boundary and is closely correlative with the planktonic foraminiferal N4–N5 boundary (Theyer & Hammond 1974b).
3. The last appearance of *Globorotalia kugleri*, whose total range defines Zone N4, occurs in most cases between the base of Chron C6AN and the top of Chron C6AAN whereas the FAD of *Globigerinoides altiaperturus* is associated with the top of Chron C6AN (*c.*20.9 Ma), within Zone N5 (Appendix 1). The range of *G. kugleri* in the NW Pacific has been suggested to extend above the FAD of *G. altiaperturus* (Keller 1981) and even above the FAD of *Globigerinoides trilobus*, in Zone N6.
4. In agreement with point (3), we note that in the Burdigalian sequence the FAD of *Globigerinoides altiaperturus* also occurs in Zone N5 (Jenkins 1966), above the LAD of *Globorotalia kugleri* (Chapronière 1981), and also in Italian lower Miocene microfaunas (Borsetti *et al.* 1979).

Thus a comparison of calcareous plankton biostratigraphy in the stratotype Burdigalian with magnetobiostratigraphy of deep sea cores indicates that the Burdigalian is correlative with Chrons C6 and C6A at least.

Middle Miocene magnetobiochronology has been a controversial topic during the past decade since the first attempt at integrating magnetic stratigraphy and biostratigraphy by Theyer & Hammond (1974a, b) and Ryan *et al.* (1974) and the (predominantly) biochronologic study by Berggren & Van Couvering (1974). Direct and indirect magnetobiostratigraphic correlations led Ryan *et al.* (1974) and Theyer & Hammond (1974a) to equate magnetic anomalies 5 and 5A with polarity Epochs (Chrons) 9 and 11, respectively rather than with Chrons 11 and 13, as implied earlier by Foster & Opdyke (1970). These correlations have essentially formed the basis for middle Miocene bio- and magnetobiochronology during the past decade. New data from DSDP Sites 519 (Poore *et al.* 1983) and 558 and 563 (Miller *et al.* 1985) from the South and North Atlantic, respectively, have cast doubt upon these correlations and call into question the resulting magnetobiochronology of the various biostratigraphic zones, datum events, and chronostratigraphic boundaries. In the discussion below we shall review the basic biostratigraphic framework of marine middle Miocene chronostratigraphy and present new data which requires a major revision to the magnetobiochronology of this interval.

The middle Miocene (Langhian and Serravallian stage) is biostratigraphically bracketed by the FAD of *Praeorbulina glomerosa curva*, nominate taxon of redefined Zone N8

(Jenkins *et al.*, 1981) which is associated with the younger part of Chron C5CN (16.5 Ma; see Appendix II) and the FAD of *Neogloboquadrina acostaensis*, one of the nominate taxa of Zone N16. This latter 'datum' was identified by Ryan *et al.* (1974) in the lower part of the stratotype Tortonian at Rio Mazzapiedi at a level said to be equivalent to a normal event assigned to magnetic polarity Epoch (Chron) 10. The base of the Tortonian was correlated with a slightly older level, correlated to Chron C5AN and Chron 11, whereas the remainder of the Tortonian was said to span the interval represented by Chrons 10 to 7. We shall show below that the base of the Tortonian probably corresponds to the base of Chron C5N, corresponding to Chron 11 in the original correlation of Foster & Opdyke (1970). We note in passing that the FAD of *Orbulina suturalis* is associated with the base of Chron C5BN (*c.*15.2 Ma; Appendix II). Our magnetobiochronologic age estimate agrees precisely with the age assigned to this datum in Japan based on numerous radiometric age determinations bracketing this datum (Ikebe 1977) as well as with newly available dates on the Island of Martinique (Andreieff, pers. comm., February, 1984; see discussion below).

Calcareous nannoplankton studies suggest that the type Langhian belongs to Zones NN4, NN5 and NN6 (*partim*) (Martini 1968, 1971). The Cessole Formation, equivalent to a part of the Langhian, belongs to the *Sphenolithus heteromorphus* (NN5) Zone and the 'Transitional Unit' (between the Cessole Formation and the Serravalle Formation) in the Serravalle Scrivia section belongs to the *Discoaster exilis* (NN6) Zone (Müller 1975) and planktonic foraminiferal zones N9–N10 (Cita & Blow 1969) and the basal part of the overlying Serravalle Formation to the *Discoaster kugleri* (NN7) Zone (Müller 1975). The upper part of the Serravalle Formation at Serravalle Scrivia extends to the N13–N14 boundary (Ryan *et al.* 1974). In the Borbera section the 'Transitional Unit' belongs to Zone NN6 (Müller 1975) and corresponds to zones N11–N12? (Cita & Blow 1969).

Magnetostratigraphic studies, on the other hand, along the Bubbio-Casinasco Road have been subjected to alternative interpretations. Ryan *et al.* (1974: 647, 650) correlate the Langhian–Serravallian boundary with Chron C5BN at the base of Epoch (Chron) 15 on the basis of the occurrence of the '*Orbulina* Datum' just below the top of the Cessole Formation (which occurs within Chron 16). On the other hand, Nakagawa (1974) and Nakagawa *et al.* (1977) show the Langhian and Serravallian stages overlapping at the base of Chron 14 (which in the revised numeric system employed here would fall within the Chron C5AA-C5AB interval).

North Atlantic DSDP Site 396 was drilled on the eastern flank of the Mid-Atlantic Ridge and is situated between magnetic anomalies 5A and 5B (Schouten, pers. comm., January, 1984), *not* as originally reported, at anomaly 5 (Purdy *et al.* 1978: 119). The oldest sediments immediately above basement were placed in the *Globorotalia fohsi fohsi* Zone (Krasheninnikov 1978: 321) and the *Discoaster exilis* Zone (Bukry 1978: 309; note that Bukry, *loc. cit.*, states that Site 396 was drilled on the *western* flank of the Mid-Atlantic Ridge, on crust magnetized between magnetic anomalies 5A and 5B).

We have found (M-P. Aubry, pers. comm., January, 1984; Miller *et al.*, in press) that the NN5–NN6 (LAD *Sphenolithus heteromorphus*) and NN6–NN7 (FAD *Discoaster kugleri*) boundaries occur within Chron C5AD and within C5AA–C5AB interval, respectively, in DSDP Holes 558 and 563, i.e. between anomaly 5A and 5B correlatives. This latter correlation supports the studies on DSDP Site 396. We have, then, two options for drawing the Langhian–Serravallian boundary:

(a) base Chron C5BN which lies within Zone NN5 (Ryan *et al.* 1974: 665);

(b) at or near the NN6–NN7 boundary (Müller 1975) which lies within the Chrons C5AA–C5AB interval (Aubry *In*: Miller *et al.*, in press.)

There may be a stratigraphic hiatus between the Langhian and Serravallian stages but we see no evidence for this is in the published literature dealing with the litho-, bio-, or magnetostratigraphy. Accordingly we show an interval of stratigraphic overlap between Chron C5BN and the Chron C5AA–C5AB interval corresponding to the uncertainty in placing this boundary based on the magnetobiostratigraphic interpretations of Ryan *et al.* (1974) or Nakagawa (1974), Nakagawa *et al.* (1977; see also Müller 1975). However, we favour the latter interpretation based on the evaluation of data presented here, i.e. in the Chron C5AA/C5AB interval.

In the Rio Mazzapiedi section the type Tortonian overlies the Serravallian but the youngest definitive age assignment within the Serravallian is Zone NN7 (Müller 1975). It is not possible to determine the biostratigraphic position of the upper part of the Serravallian and/or the Serravallian–Tortonian boundary based upon extant/published data.

In DSDP Hole 563 the *Sphenolithus heteromorphus* (NN5)/ *Discoaster exilis* (NN6) boundary occurs in lower core 8 in the normal event (= Chron C5ADN) just above Chron C5BN (Miller *et al.*, in press) and within Zone N10. Zone NN7 spans the interval from the upper part of core 8 or lower part of core 7 to the lower part of core 4 (at least) (Chron C5AA/ C5AB interval and the early part of C5). Zone NN7 brackets Chron C5AN and the FAD of *Globigerina nepenthes* occurs in core 5, just above Chron C5AN, i.e., within Zone NN7 (cf. discussion in Berggren & Van Couvering 1978: 43, 44). The upper limit of *Globorotalia fohsi robusta* (= N13) and the initial appearance of *G. nepenthes* (= N13–N14) are virtually coincident here, calling into question the validity of Zone N13 (cf. Blow 1969: 243–246; Berggren & Van Couvering, 1978: 46). At the same time the LAD of *Globorotalia siakensis* and the FAD of *N. acostaensis* are virtually coincident at the Chron C5N in Hole 563, calling into question the validity of Zone N15, as well.

The type Tortonian has been assigned to calcareous nannoplankton zones NN9–NN11 (*partim*) (Martini 1971, 1975) and to the *Globorotalia mayeri* — *G. lenguaensis* through *G. menardii* — *Globigerina nepenthes* Zones (Cita *et al.* 1965) which correspond to Zones N15 through N17 (*partim*; Blow 1969; Cita & Blow 1969). The base of the stratotype Tortonian in the Rio Mazzapiedi section was assigned to the *Discoaster hamatus* (NN9) Zone based on the sole occurrence of the nominate taxon in sample 4 which was interpreted by Martini (1975) and accepted by Ryan *et al.* (1974) as the LAD of this taxon. This is the same level as the supposed FAD of *Neogloboquadrina acostaensis* in the Rio Mazzapiedi section (Cita *et al.* 1965). However, the lower 35–40 m of the Tortonian at Rio Mazzapiedi contain poor and undiagnostic microfossils and thus the occurrence of *N. acostaensis* and *D. hamatus* may, in fact, *not* be actual initial and terminal stratigraphic occurrences (see below). In fact, Ryan *et al.* (1974: 648, figure 4) indicate a stratigraphic range for *D. hamatus* extending over 100 m in the Rio Mazzapiedi section from about 70 m below the base of the stratotype Tortonian and through the lower 35 m of the Tortonian, up to sample 4. This range, according to their interpretation, spans the interval between the normal event in Chron 12 to the normal event in Chron 10. However, there is no evidence for this range, (see Martini 1975 and remarks above). If this were true, however, it would place the *Discoaster coalitus* (NN8) Zone well down within the Serravallian, within Chron 12, and close to the base of Zone N14. Indeed, this is the interpretation made by Ryan *et al.* (1974: 655) and supported by independent arguments by Berggren & Van Couvering (1978: 44).

The presence of *D. hamatus* in the lower part, and the apparent FAD of *D. quinqueramus* in the upper part, of the type Tortonian, indicates that Zone NN10 is restricted to the Tortonian (Martini 1971, 1975; Ryan *et al.* 1974). Ryan *et al.* (1974: 649) indicate that the upper range of *hamatus* in piston core RC12–65 is limited to below magnetic Epoch (= Chron) 10/11 by dissolution, but that by cross correlation with piston core V19–99 it can be shown that the LAD of *hamatus* occurs within the brief stratigraphic range of the diatom *Coscinodiscus vetustissimus* var. *javanicus* which spans the normal polarity event of Chron 10, which in turn is correlated with the (supposed) LAD of *hamatus* in sample 4 of the stratotype Tortonian.

At DSDP Hole 519 *Catinaster coalitus* and *C. calyculus* occur sequentially in, and are restricted to, sediments of Chron C5N age (Poore *et al.* 1983: 131). Von Salis (1984: 425) has confirmed the association of these two taxa in Hole 521A with sediments interpreted as belonging to Chron C5 (Heller *et al.* 1984). These, and other related, data were subjected to two interpretations by members of the shipboard party of DSDP Leg 73:

1. The appearances of *C. coalitus* and *C. calyculus* were considered to have been delayed relative to their appearance in equatorial Pacific cores; thus the entire interval spanning Chron C5N and older sediments was correlated with Zone NN10 (Percival 1983).
2. The appearances of *C. coalitus* and *C. calyculus* were interpreted as being true FAD's, the bulk of Chron C5N sediments was correlated with Zones NN8 and NN9 and the basal 5 m above the sediment basement contact in Hole 519 Chron C5 was correlated with Zone NN7 (equivalent; Hsü *et al.* 1984: 628). An essentially identical interpretation was reached by Hsü *et al.* (1984) for Hole 521A based on the study of Von Salis (1984).

This latter interpretation conflicts with that proposed by Ryan *et al.* (1974) for this interval as observed by Hsü, La Brecque *et al.* (1984: 47) as well as with a K/Ar date of 11.4 ± 0.6 Ma (Dymond 1966) on an ash in the experimental Mohole Site of the east Pacific associated with the FAD of *Catinaster coalitus*. The supposed FAD of *C. coalitus* within Chron C5N yielded a magnetostratigraphic age estimate of 9.6 Ma in the chronology of Hsü, La Brecque *et al.* (1984).

Our own correlation over this interval is based upon studies in DSDP Holes 558 and 563. In DSDP Hole 563 the FAD of *C. coalitus* and *Discoaster hamatus* occurs in the reversed interval just *below* Chron C5N and within the lower part of Chron C5N, respectively (M-P. Aubry, pers. comm., January, 1984). Our magnetobiostratigraphic age estimate for the FAD of *C. coalitus* is 10.8 Ma which is significantly closer to the radiometric date of Dymond (1966) and helps, to some extent, resolve some of the discrepancy noted by Hsü, La Brecque *et al.* (1984: 47). Our results then, while generally supporting the interpretation of Hsü *et al.* (1984) differ in a minor, yet significant way, with regard to their magnetostratigraphic correlation of the NN7–NN9 interval. In Hole 563 we have also found that the FAD of *Neogloboquadrina acostaensis* (Zone N16) occurs in core 4, at the base of Chron C5N, and with the range of *D. coalitus* (Zone NN8).

Our studies on the magnetobiostratigraphic correlations in DSDP Hole 563 (see Appendix II) have yielded additional data relevant to the late Miocene chronostratigraphy and magnetobiochronology (Miller *et al.*, in press). Thus the NN9–NN10 boundary (= LAD *D. hamatus*) is associated with the upper part of Chron C5N (8.85 Ma), the LAD of *D. bollii* (which occurs within Zone NN10) is associated with an interval of no palaeomagnetic data just below a normal event interpreted here as being the upper of the two normal events (8.3 Ma) associated with Chron C4AN. Forms transitional between *D. bellus* and *D. quinqueramus* (= NN10–NN11 boundary) occur in the normal event identified here as the upper normal event of Chron C4AN (8.2 Ma), consistent with the record of the FAD of *D. quinqueramus* of Haq *et al.* (1980: 430).

In summary our data indicate that:

(a) Zone NN8 spans the base of Chron C5N.
(b) Zone NN9 corresponds to the upper 3/4 of Chron C5N and extends into the reversed interval just above Chron C5N.
(c) Zone NN10 extends from just above Chron C5N to C4AN.

From the data summarized above we can draw several conclusions and suggest some revisions to currently accepted magnetobiostratigraphic correlations.

1. The correlation of anomaly 5 with magnetic polarity Epoch (= Chron) 9 and anomaly 5A with Epoch (= Chron) 11 by Ryan *et al.* (1974) and Theyer & Hammond (1974a) is probably incorrect because Zones NN8 (*partim*), and NN9 can be shown to occur in sediments deposited during anomaly 5 time.
2. The Serravallian–Tortonian boundary probably lies at, or near, the base of Chron C5N, or at any rate is probably not older than the first normal event below Chron C5N. It probably lies within Zone NN8 and we place it here at the base of Chron C5N (= 10.4 Ma).
3. The realignment of calcareous plankton zones (and corresponding correlations with siliceous plankton zones) to magnetic anomaly correlatives results in an approximately 1.5–2 my shift (toward younger age estimates) over part of the middle Miocene interval (see Appendix II for various datum levels recognized in this interval).

Late Miocene chronology of biostratigraphic and chronostratigraphic units remains essentially unchanged from earlier work. The Tortonian–Messinian boundary is correlated approximately with the FAD's of *Globorotalia conomiozea* and the *Amaurolithus delicatus-primus* group (see Appendix II) and antedates slightly the prominent carbon isotopic shift in the deep sea (Keigwin 1979; Haq *et al.* 1980; Vincent *et al.* 1980).

Integrated magnetobiostratigraphic studies on several upper Miocene sections on Crete have led to the correlation of several biostratigraphic events to a polarity sequence identified as Chrons 5 and 6 (Langereis *et al.* 1984). The initial appearance of *Globorotalia conomiozea* in a reversed interval between two normal events identified with Chron 5 (at 5.6 Ma) led to several conclusions:

(a) The FAD of *G. conomiozea* in the Mediterranean (5.6 Ma) was interpreted as a delayed (migration) event, following

by approximately 0.5 my the initial evolutionary appearance of this taxon at *c*.6.1 Ma in the SW Pacific.
(b) Since the Tortonian–Messinian boundary is defined in the boundary stratotype section in Sicily at a level which coincides essentially with the initial occurrence of *G. conomiozea*, the authors suggest that the Tortonian–Messinian boundary has an age of about 5.6 Ma.
(c) Since the initial appearance of *G. conomiozea* on Crete (near the top of Chron 5) is below the level of the sinistral to dextral coiling change in *Neogloboquadrina acostaensis*, the latter event must be younger than the base of the Gilbert Chron (i.e. younger than 5.3 Ma).
(d) Since the coiling change in *N. acostaensis* occurred *below* the base of the Messinian evaporites, they in turn must lie wholely within the Gilbert Chron.

We have reservations about these conclusions. Briefly stated they are:
1. The sinistral to dextral coiling change in *N. acostaensis* has been previously shown to occur in the lower normal polarity event of Chron 5 in the Pacific (Saito *et al.* 1975) and eastern Atlantic, DSDP Site 397 off Cape Bojador (Mazzei *et al.* 1979) with a palaeomagnetic age estimate of 5.8 Ma.
2. The FAD of *Ceratolithus acutus* has been associated with a level approximately 6 m above the base of the Zanclean Stage (lower Pliocene; Cita & Gartner 1974) and shown to occur in the earliest Gilbert Chron (Mazzei et al. 1979; Berggren *et al.* 1983) with an estimated palaeomagnetic age of 5.0 Ma.
3. Keigwin & Shackleton (1980) identified a distinct carbon shift of about 0.5‰ in Chron 6 (*c*.6.3 Ma) in piston core RC12-66 just below a dextral to sinistral coiling change in *Neogloboquadrina acostaensis* in the younger part of Chron 6 (6.1 Ma) and the sinistral to dextral coiling change in *N. acostaensis* in mid Chron 5 (= Chron C3AN) (5.8 Ma) which is the same biostratigraphic event observed in the Mediterranean just below the evaporite series. This carbon shift has been subsequently observed in several DSDP sites from the Pacific and Indian oceans (Haq *et al.* 1980) and gives a slightly younger age (5.90–6.10 Ma based on a different magnetic polarity chronology). This carbon shift has also been observed in the Kapitean Stage of the Blind River section of New Zealand within Chron 6 with an estimated magnetochronologic age of 6.2–6.3 Ma and virtually coincident with the FAD of *Globorotalia conomiozea* in that section (Loutit & Kennett 1979). This same carbon shift has also been found in the Carmona-Dos Hermanos (stratotype Andalusian Stage) section in the Guadalquiver Basin of south-western Spain (Loutit & Keigwin 1982) over a stratigraphic interval which lies between the FAD of *Amaurolithus primus* (recorded as *Ceratolithus tricorniculatus* in Van Couvering *et al.* 1976) and calibrated to Chron 6 at *c*,6.5 Ma here (see Appendix II, Table 7) and the base of the Caliza Tosca (correlated biostratigraphically with the base of the Messinian evaporites). Inasmuch as it is almost certain that the carbon shift in the Carmona-Dos Hermanos section is identical to, and correlative with, the late Miocene carbon shift seen in the global ocean(s), we can correlate the carbon shift at approximately 6.1–6.2 Ma with a level at, or in close temporal proximity to, the base of the Messinian.

These data are consistent with the age estimate of about 5.3 Ma for the Miocene–Pliocene boundary and support our earlier estimates (Van Couvering *et al.* 1976) that the Messinian evaporites are approximately time correlative with the younger part of Chrons 5 and basal Gilbert. Burckle (1978) has reached the same conclusion in correlating the Tripoli beds (unit 2) below the evaporites with the top of the reversed interval in Chron 5 in Capodarso, Sicily. Burckle & Opdyke (in press) have also correlated the base of the Caliza Tosca in the Andalusian Stage stratotype section with the same magnetostratigraphic interval. The age of the Tortonian–Messinian boundary remains somewhat controversial. It would appear to lie biostratigraphically close to the FAD's of *Globorotalia conomiozea* and *Amaurolithus tricorniculataus* over the range of 6.1–6.3 Ma. Magnetobiostratigraphic correlations by Ryan *et al.* (1974) suggested that the Tortonian–Messinian boundary was closely associated with the polarity Chron 6/7 boundary and the FAD of '*A. tricorniculatus*' (*c*.6.5 Ma). In this paper we retain this age estimate and place the boundary within the mid-part of Chron 6 coincident with the appearance of the earliest amaurolithids, *A. primus* (= including earlier references to *A. tricorniculatus*).

The regional magnetobiostratigraphic correlations discussed above suggest that the Cretan magnetic polarity sequence (Langereis *et al.* 1984) may have been misinterpreted. The Cretan correlations would be more consistent with published data if the normal polarity magnetozone above the FAD of *G. conomiozea* were identified with the lower normal of Chron 5. The remaining magnetic reversals below may then represent Chrons 6 and 7 (*partim*).

The revised middle and late Miocene magnetobiochronologic age estimates derived here are seen to be in remarkably close agreement with high temperature radiometric dates in Japan of 11.6 ± 0.4 Ma on the *Globigerina nepenthes* FAD (= N13–N14 boundary), 14.5 ± 0.4 Ma on the *Globorotalia peripheroacuta* FAD (= N9–N10 boundary), and interpolated biochronologic age estimates of *c*.11 Ma on the *Globorotalia siakensis* LAD (= N14–N15 boundary) and 10.0–10.5 Ma on the FAD of *Neogoboquadrina acostaensis* (= N15–N16 boundary) (Ikebe *et al.* 1981; cf. Tsuchi, (ed.) 1981).

Additional data comes from the Island of Martinique (Andreieff *et al.* 1976; P. Andreieff, pers. comm., February, 1984) where a series of intercalated sediments and basalt flows dated by the potassium argon method have yielded the following results (from older to younger):
(a) The *calcaires de Sainte-Anne* (*Globigerinatella insueta* Zone = Zones N7–N8) are bracketed by two basalt flows dated 17.8 Ma ± 1.8 (at the base) and 15.9 ± 0.5 Ma (at the top).
(b) Les calcaires du Francois and the *'tufs' de Bassignac* (*G. peripheroronda* Zone = Zone N9) are separated from the *Tuffites du Marin* s.s. (*G. fohsi — G. fohsi lobata robusta* Zones = Zone N10–N12) by volcanic deposits dated at 15.0 ± 0.3 Ma.
(c) The top of the *Tuffites du Marin* s.s (= lower part of *G. fohsi robusta* Zone = Zone N12) are overlain by a lava flow dated at 12.8 ± 0.6 Ma.
(d) The *Tuffites de Fort de France* (upper part of *Neogloboquadrina humerosa* Zone = top Zone N17, = Zone M12 and Zone M13 with *Globorotalia cibaoensis — G. margaritae* of Berggren *et al.* 1983) lie above an andesite dated at 6.4 ± 0.6 Ma.

These dates are seen to be in remarkably close agreement with the derived magnetobiochronology in this paper.

The Miocene–Pliocene boundary

The Miocene–Pliocene boundary was discussed in detail by

Cita (1975), who proposed a boundary stratotype section coincident with the base of the stratotype Zanclean stage at Capo Rosello, Realmonte, Province of Agrigento, Sicily. Cross-correlation between the biostratigraphically dated section at Capo Rosello and biostratigraphically and palaeomagnetically dated DSDP sites (125 and 132) in the Mediterranean led Cita (1975, 1976) to suggest that the Miocene–Pliocene boundary was biostratigraphically linked via the base of the *Sphaeroidinellopsis* Acme-Zone, with the upper part of palaeomagnetic polarity Chron 5 (= youngest part of Chron C3A) estimated to have an age of 5.3–5.4 Ma. However, the palaeomagnetic studies conducted on these cores are inadequate for a definitive identification of a polarity reversal history (Kennett & Watkins 1974) and studies by Berggren (1973) and Saito *et al.* (1975), among others, have suggested that the Miocene–Pliocene boundary is biostratigraphically linked via the LAD of *Globoquadrina dehiscens* and the FAD of *Globorotalia tumida* with the base of the Gilbert Chron, estimated to have an age of 5.3 Ma (Appendix I). Essentially similar conclusions have been drawn by Thunell (1981) and Van Gorsel & Troelstra (1981) who have made studies of low latitude planktonic foraminiferal successions in equatorial cores from the Atlantic, Pacific, Indian Ocean, and Indonesia, respectively.

The Pliocene

We follow previous usage (Berggren 1973; Berggren & Van Couvering 1974) in retaining a bipartite subdivision for the Pliocene of the Zanclean (lower) and Piacenzian (upper) stages. The relationship of these units to standard calcareous plankton biostratigraphy has been shown by Cita & Gartner (1974) and calcareous plankton biochronology of the Pliocene has been summarized by Berggren (1973). Over 30 planktonic foraminiferal and nearly 25 calcareous nannoplankton datum events have been palaeomagnetically correlated resulting in a high degree of biostratigraphic and biochronologic resolution within this geologically short (approximately 3.7 m.y.) epoch (Appendix II).

A recent study on the Lower Pliocene Suva Marl of Fiji (Rodda *et al.*, in press) is of particular interest to the Pliocene magnetobiochronologic framework presented here. By combining data on nearly 40 K/Ar (biotite) dates on volcaniclastic tuffs (ranging sequentially in age from *c.*4.8 Ma–3.8 Ma) in an approximately 180 m thick sequence of predominantly reversed polarity (but with three normal magnetozones identified as the Thvera, Sidufjall and Nunivak events of the Gilbert Chron) and calcareous plankton biostratigraphy, it has been possible to estimate ages for the normal polarity subchron boundaries as well as several biostratigraphic zonal boundaries and events. An age-stratigraphic height (relative to O-level marker bed) regression was used to estimate ages of the polarity boundaries which are seen to be in close agreement with the polarity scale of McDougall (1979) based predominantly on age and polarity data on subaerial volcanic rocks. Age estimates (in Ma) for the lower three normal events of the Gilbert Chron are compared below with those derived in this paper presented in parenthesis:

Nunivak: 4.09–4.22 (4.10–4.24);
Sidufjall: 4.52–4.67 (4.40–4.47);
Thvera: 4.87–4.93 (4.57–4.77).

The differences in chronology of the lower two polarity subchrons is puzzling both in regards to age (the estimate of Rodda *et al.*, in press, are approximately 0.15 to 0.3 m.y. older than ours) and duration. In the Suva Marl of Fiji the Sidufjall and Thvera Subchrons are shown to have durations of 150000 years and 60000 years, respectively, which is in inverse proportion to their respective durations as measured in sea floor anomaly profiles (Sidufjall: 70000 years; Thvera: 200000 years; Klitgord *et al.* 1975) and deep sea sediments (Sidufjall: 70000 years; Thvera: 170000 years; Opdyke 1972).

We note further that there are no constraints (by way of K/Ar dates) on the age of the lower normal magnetozone (= Thvera) in the Suva Marl inasmuch as the oldest K/Ar date of 4.8 Ma is from a stratigraphic level *above* the Thvera Subchron so that direct extrapolation from stratigraphically higher levels may not be yielding a reliable numerical value for this subchron.

Finally we note the close correspondence between the radiochronologic framework provided for early Pliocene calcareous plankton biostratigraphy on Fiji (Rodda *et al.*, in press) and the magnetobiochronology derived here (see Fig. 2 and Appendix II):

1. 33 K/Ar dates over the stratigraphic interval representing the combined Zones NN13 and NN14 range from 4.9 to 4.0 Ma with estimated boundary ages of 4.82 Ma (NN12/NN13) and 3.86 Ma (NN14/NN15) based on a best fit regression through the dated values. By comparison our magnetobiochronologic estimates for the NN12–NN13 and NN14–NN15 boundaries are 4.6 and 3.7 Ma, respectively.
2. 6 K/Ar dates from the same stratigraphic level within Zone NN15 yield average dates (on replicate samples) of 3.78 Ma, 3.74 Ma, and 3.65 Ma with an estimated age for the NN15–NN16 boundary of 3.52 Ma based on best fit regression through the dated values. By comparison our magnetobiochronologic estimate for the NN15–NN16 boundary is 3.5 Ma.

The Pliocene–Pleistocene boundary

The Pleistocene, and particularly its lower boundary, remains as controversial as ever. Recent summaries (Haq *et al.* 1977; Pelosio *et al.* 1980; Colalongo *et al.* 1981; Pasini & Colalongo 1982) have clarified the problems associated with an adequate definition in the Mediterranean of a Pliocene–Pleistocene boundary stratotype and its recognition elsewhere. It is now apparent that the over 300 m thick, mid- to upper bathyal, Vrica section, south of Crotone (Calabria) is more suitable as a boundary stratotype section (Selli *et al.* 1977) than the earlier nominees at Santa Maria di Catanzaro and Le Castella.

Integrated bio- and magnetostratigraphic studies suggested that the 'marker bed' at Le Castella is closely associated with a level coincident with, or slightly younger than, the top of the Olduvai Event, *c.*1.6 Ma (Haq *et al.* 1977). Investigations at the Vrica section are essentially complete. Biostratigraphic studies (Colalongo *et al.* 1981) indicate that the Pliocene–Pleistocene boundary, on palaeontological grounds, should be located within an approximately 36 m thick interval (the *e–m* interval) which extends from a distinct sapropel layer (*e*) to the level of occurrence of a distinct volcanic ash bed (*m*). This latter bed has been the subject of some considerable attention lately (Boelstorff 1977; Savelli & Mezzetti 1977; Selli *et al.* 1977). Obradovich *et al.* (1982) have recently shown that the previous age determinations attributed to this poorly-preserved ash, indicating an age over 2.5 Ma, have been mistaken. Samples from a second and much thicker ash, tens of metres downsection in Pliocene strata, were apparently the source of the published ages. The 'Pleistocene ash' (level *m*) can only be said to be less than 1.9 Ma.

The (*m*) ash lies approximately 9 m above the FAD of *Cytheropteron testudo* and is essentially coterminous with the FAD of *Gephyrocapsa oceanica* (Colalongo *et al.* 1981). These authors have recommended that the Pliocene–Pleistocene boundary be placed at a level immediately preceding the FAD of *C. testudo*, the first of the so-called 'cold guests', or 'northern guests' (Suess 1983), whose appearance in the Mediterranean has been considered to denote the beginning of the Pleistocene Epoch.

Several planktonic foraminiferal events have been shown to span the *e–m* interval which may aid in recognition of the boundary (as formally proposed) in extra-Mediterranean regions. Calcareous nannoplankton (Backman *et al.* 1983) and palaeomagnetic stratigraphy (Tauxe *et al.* 1983) have recently been re-investigated in this section in an attempt to resolve previous discrepancies in interpretation. Two intervals of normal polarity separated by a short interval of reversed polarity have been identified as N1 and N2, respectively, in the lower part of the Vrica Section (Tauxe *et al.* 1983). A third normal polarity interval (N3) has been identified in the upper 40 m of the exposed section. The LAD's of *Discoaster brouweri* at the base of the lower (N1) normal interval and of *Cycloccolithus mcintyrei* about 56 m above the top of the second (N2) of the lower two normal events supports the identification of these two normal polarity intervals as the Olduvai Subchron. The physical horizon immediately below the initial appearance of *Cytheropteron testudo* is located about 9 and 10 m, respectively, above sapropel bed *e* and the top of N1 + N2 (= Olduvai Subchron) at about 1.6 Ma.

As this paper goes to press, members of the INQUA Subcommission on the Pliocene–Pleistocene Boundary and of International Geological Correlation Project 41 (Pliocene–Pleistocene Boundary) are formulating a resolution regarding the (revised) definition of the Pliocene–Pleistocene boundary in the Vrica Section of Calabria, Southern Italy, to be submitted to the IUGS International Committee on Stratigraphy. This resolution will suggest locating the Pliocene–Pleistocene boundary stratotype at the top of marker bed *e* about 3–6 m above the top of the Olduvai normal polarity event. It will be seen that this recommendation results in continuity with some previous studies which placed the boundary at the top of the Olduvai Event, at *c.*1.6 Ma (Haq *et al.* 1977; Berggren *et al.* 1980).

The Pleistocene

Attempts at formulating a unified Pleistocene chronostratigraphy have suffered from the climatic overprint which characterizes the late Neogene part of geologic time. The classic units have been recognized in northern Europe and have been correlated with the continuous oxygen isotope records from the oceans (Shackleton & Opdyke 1973, 1976) by means of the intermediate link of loess and terrace stratigraphy (Kukla 1970, 1975, 1977). Problems in classic stratigraphic subdivision of the Pleistocene have been compounded by the fact that it is now apparent that while the terraces representing the four classic Alpine 'glacial' ages cover the last 0.8 m.y., they correspond to both glacial and interglacial climatic intervals. Furthermore at least 17 glacials and interglacials have been recognized over the past 1.7 Ma in Europe, eight of which belong to the Brunhes Chron (last 0.73 m.y.). Some 23 oxygen isotope stages representing glacial-interglacial cycles have been recognized in the deep sea since the late Jaramillo Subchron with nine of these cycles being confined to the Brunhes Chron (and corresponding to the eight glacial-interglacial cycles represented in the loess sequence of Europe). Thus the stratigraphic record upon which the classic North European Pleistocene chronostratigraphy is based represents but fragmentary portions of the total Pleistocene Series. This has led to both misunderstanding of the nature of the stratigraphic record and miscorrelation between the marine and continental record, and we would agree with Kukla (1977) in his call for elimination of classic climatostratigraphic units from the Pleistocene chronostratigraphic hagiography and substitution of the $\delta^{18}O$ record in deep sea sediments in their place.

It will suffice to point out here that the degree of chronologic resolution increases considerably as we approach the present time. In particular the oxygen isotope scale (stages 1–23), extending from the present day to the Jaramillo Subchron (t = *c.*0.9 Ma), can discriminate essentially isochronous climatic cycles with a cyclicity of approximately 0.09 m.y. The use of Pleistocene biostratigraphic datum levels within a framework of oxygen isotope and magnetic-reversal scales leads to a further chronologic resolution on the order of 3 000–5 000 years (Berggren *et al.* 1980) in some instances. The relationship of Pleistocene calcareous plankton datum events to palaeomagnetic stratigraphy is shown in Table 7 (Appendix II). Spectral analysis of the $\delta^{18}O$ records in five deep sea cores by members of the SPECMAP project (Imbrie *et al.*, in press) has provided the basis for a revised chronology of the marine $\delta^{18}O$ record for the past 0.78 m.y. and provided strong evidence that orbital variations are the main external cause of the succession of middle and late Pleistocene ice ages.

At the same time, larger-scale perturbations of the climatic base-line are recorded in less sensitive marine systems, which show four principal climatic 'lows' (periods of maximum winter cold and maximum seasonality) at roughly 0.5 m.y. intervals during the Pleistocene, centred at 1.6, 1.0, 0.4, and 0.1 Ma (Briskin & Berggren 1975). These are, despite the imperfections of the continental record, clearly equivalent to distinct cold-climate intervals evident in mammalian biochronology (summarized by Berggren & Van Couvering 1974, 1979) and deserve to be recognized, at least in geohistorical scenarios, as the traditional 'Ice Ages' of Donau/Nebraskan, Günz/Kansan, Mindel/Illinoian, and Riss-Würm/Wisconsin.

The Marine Neogene time-scale: summary

Our assessment of palaeontologically correlated magnetochronologic age estimates for the various Neogene boundaries, based on the revised magnetochronology presented here, are: Pliocene–Pleistocene (just above top Olduvai): 1.6 Ma; Miocene–Pliocene (basal Gilbert): 5.3 Ma; Oligocene–Miocene (mid Chron C6N): 23.7 Ma (see Fig. 2).

Features of interest in this revised Neogene time-scale include the following:

1. 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 just below Chron C6CN). It thus retains its usefulness as a guide to the approximate position of the Oligocene–Miocene boundary.

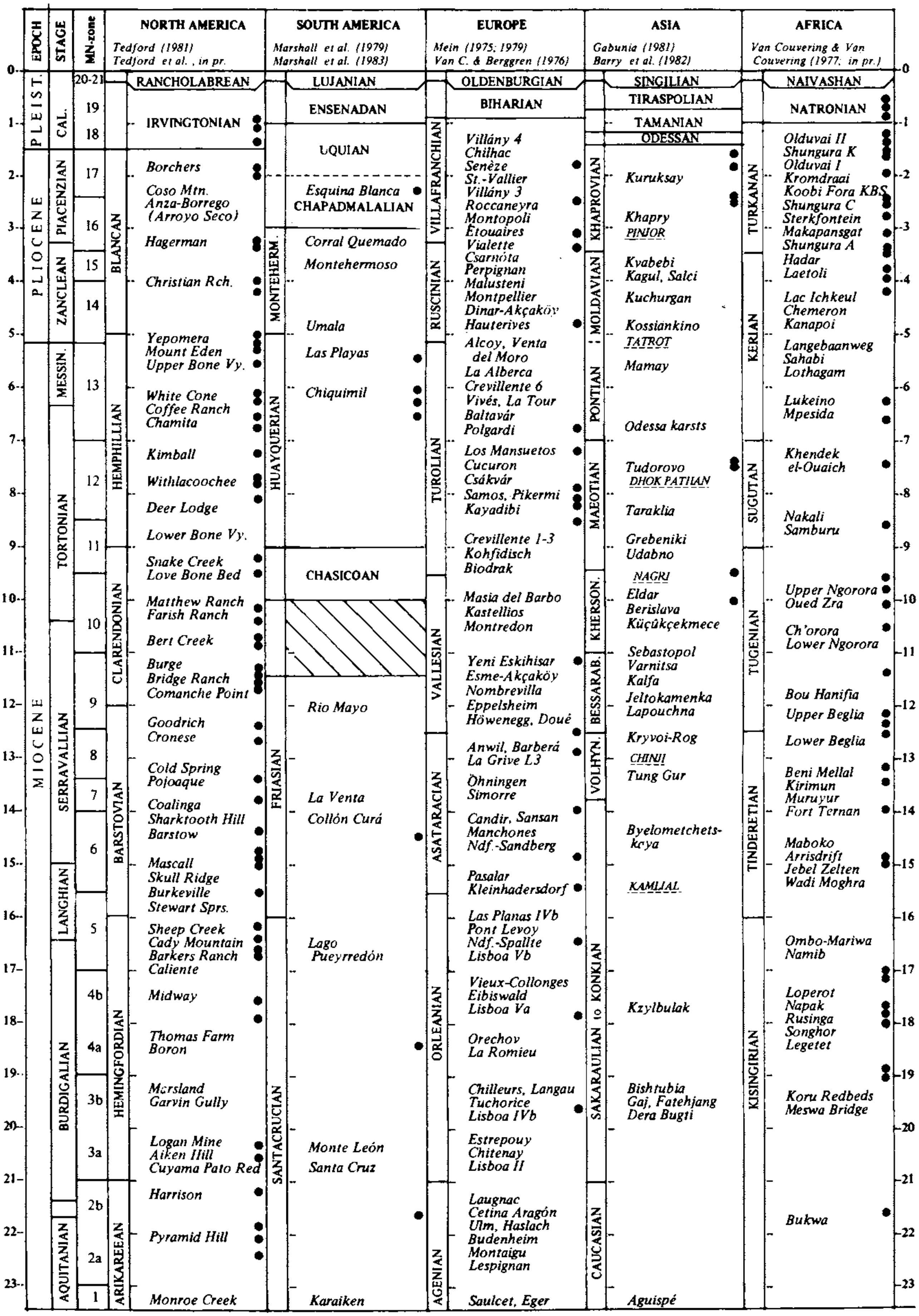

FIG. 3. Neogene continental time-scale. Principal mammal faunas from the Neogene of the indicated regions are shown in relative biochronological order, and are positioned, within the limits imposed by the medium, according to best estimates of absolute age. The biochron boundaries are dated to the nearest 0.5 Ma, except where better control is available, according to the estimated ages of the included faunas. 'MN-zone' (column 3) gives the suggested calibration for the divisions of European 'Mammal Neogene' biochronology defined by Mein (1975, 1979). Note that many major collections are from composite sections with a number of stratigraphically superposed faunal levels (e.g. Barstow; Santa Cruz; Samos; Maragheh; Lothagam). In these cases it is the 'main level,' or conceptual middle of the collection, the position of which is indicated. The Siwaliks 'mammal stages' (e.g. Nagri, Dhok Pathan), however, are shown according to the probable position of their respective bases. *Black dots* at the right side of each column denote radiometric age determinations in the regional mammal records, but not (necessarily) for any mammal fauna whose name is listed nearby. The dates are indicated mainly to give an idea of the level of radiometric control in our calibration.

2. Numerous biostratigraphic criteria have been suggested to correlate the position of the Oligocene–Miocene boundary from the Aquitanian stratotype. We have chosen the FAD of *Globorotalia kugleri* and the LAD of *Reticulofenestra bisecta*, both of which are associated with mid Chron 6CN, as the 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 (see item 14 in the Palaeogene chapter, and the discussion of California Zemorrian-Saucesian geochronology above).
3. The Aquitanian–Burdigalian boundary is correlated with the regional LAD of *Globorotalia kugleri* in Chron C6A time with an estimated magnetochronologic age of 21.8 Ma.
4. The lower Miocene stratigraphic interval encompassed by magnetic polarity Chrons C5D-C6AA contains few calcareous plankton datum events. However, at least 15 planktonic foraminiferal and five calcareous nannoplankton datum events occur in Chron C5B-C5C. The LAD of *Catapsydrax dissimilis* (= N6/N7 boundary) and the rapid sequential appearance of *Globorotalia praescitula* and *G. zealandica* occur near the Chron C5C-C5D boundary (*c.*17.5 Ma).
5. The lower–middle Miocene boundary (= base of the Langhian Stage) is biostratigraphically correlated with the FAD of *Praeorbulina glomerosa curva* which is associated with Chron C5CNI (the younger part of anomaly, correlative 5C) and which has an estimated age of 16.5 Ma.
6. The FAD of *Orbulina* (*O. suturalis*) occurs at the base of Chron C5BN with an estimated magnetochronologic age of 15.0 Ma, in precise agreement with recent assessment of radiometric dates that place the same numerical value on this datum level.
7. The middle–upper Miocene (= Serravallian–Tortonian) boundary is linked biostratigraphically with a level only slightly older than the FAD of *Neogloboquadrina acostaensis* near the base of anomaly 5 correlative (mid Chron C5 time) and has an estimated age of 10.4 Ma.
8. The base of the Messinian Stage is linked biostratigraphically with the FAD of *Amaurolithus primus* in mid-Chron 6 and has an estimated age of 6.5 Ma.
9. The Miocene–Pliocene (Messinian–Zanclean) boundary is linked biostratigraphically with the LAD of *Globoquadrina dehiscens* and the FAD of *Globorotalia tumida*, only slightly above the Gilbert/Chron 5 (Chron C3A) boundary at *c.*5.3 Ma.
10. Boundary age estimates and duration of informal divisions (in parenthesis) of the Neogene are as follows: early Miocene: 23.7 Ma–16.5 Ma (7.2 m.y.); middle Miocene: 16.5 Ma–10.4 Ma (6.1 m.y.); late Miocene: 10.4 Ma–5.3 Ma (5.1 m.y.); early Pliocene: 5.3 Ma–3.4 Ma (1.9 m.y.); late Pliocene: 3.4 Ma–1.6 Ma (1.8 m.y.); early Pleistocene: 1.6 Ma–0.73 Ma (0.87 m.y.); middle Pleistocene: 0.730 Ma–0.128 Ma (0.602 m.y.); late Pleistocene: 0.128 Ma–present (0.128 m.y.)

Continental biochronology and the palaeomagnetic time-scale

In the following section, the continental time-scale of the pre-Pleistocene Neogene, summarized in five regional syntheses of mammalian biochronology (see Fig. 3) is compared to the revised geomagnetic polarity time-scale, in order to test (as far as possible) our present assumptions about the internal accuracy of the GPTS. We note that the continental time-scale is primarily a framework of 'hard' K-Ar and zircon age determinations, which are much more available in mammal-bearing deposits than in the deep-sea sequences.

The continental (mammalian) and marine (planktonic microfossil) biochronologies, each with their own time-scale, can be compared in two ways. The classical approach is by direct stratigraphic correlation of marine and nonmarine biostratigraphy in sequences where the fossils are, or can be made to be, relevant to regional calibrated biochronologies. The second, recently-developed approach, is to find (and accurately identify) geomagnetic polarity reversals in mammal-bearing strata. Although these are still as rare as good radiometric dates in marine pelagic sections, palaeomagnetic tie points provide more precise correlations than any other in common use. (In future, stable-isotope curves may be introduced into continental correlations, if the cyclic variations can be successfully indexed.)

All comparison points, or 'tie points' that we have selected are annotated below, together with brief discussions of regional stratigraphic syntheses from which the 'tie points' obtain their control. We have also included separate discussions of two of the most significant datum events in Neogene mammalian biochronology, the immigration of two horse genera into Eurasia, respectively the '*Hipparion* Datum' and the '*Equus* Datum', both of which have been related to the GPTS and numerous radiometric dates.

With some exceptions the marine and continental calibrations brought together at the 'tie points' agree to within 0.5 m.y. Most events in mammalian history, at least before the Pleistocene, cannot be correlated to marine history from biostratigraphic-radiometric information closer than this limit, while systematic deviations in comparing some of the superpositional sequences indicate the possibility of fundamental miscorrelations in these regions (e.g. central California and Vienna Basin mid-Miocene).

Regional mammal biochronology.

Current subdivisions of the mammalian history of North America, South America, Europe, non-oriental Asia, and Africa are presented in Figure 3. The radiometric calibration of these regional biochronologies are taken in large part from published reviews but we have also relied on original interpretations of the dating literature, especially with regard to western Europe and Africa. Faunal characterizations of the biochronological units are beyond the scope of this paper, and the reader should consult the indicated references. The North American land-mammal ages are the traditional 'Wood Committee' units, currently under review by Tedford *et al.* (in press). South American land-mammal ages were first outlined by Patterson & Pascual (1972). European mammalian biochrons were characterized in the so-called MN-zones by Mein (1975; see also Mein 1979), and incorporated into regional land-mammal 'stages' subsequently (see Fahlbusch 1976). The Asian biochrons are discussed by Gabunya (1981), with reference to central Asia and the Ukraine, and by Barry *et al.* (1982) with reference to the Indo-Pakistani sequence. The African biochrons given here are from a work in progress by J. A. Van Couvering & J. A. Harris (Van Couvering).

Hipparion datum.

The immigration of hipparionine equids into the Mediterranean and central European areas is the defining event for MN-9, the base of the Vallesian biochron (Mein 1979). Although the totally mistaken Pliocene correlation of the '*Hipparion* Datum' is no longer debated, the geochronometric age of this event continues to be argued between advocates of 10 Ma and 12.5 Ma. This debates has not been greatly advanced by a concurrent upheaval in the systematics of hipparions, which has cast into doubt the identity of 'the first hipparion' in many sequences. Resolution of the systematics, however, now suggests that the age debate concerns two different datum events: first arrivals of two different hipparions, in two different areas.

Forstén (1982) and MacFadden & Skinner (1982) recapitulated the rapidly-growing literature on the systematics controversy. The conservative position, upheld by Forstén, is that most, if not all, of Eurasian hipparions (including the type species, *H. prostylum* Gervais) evolved from *Hipparion primigenium* (von Meyer), a single and locally variable species found throughout Eurasia and North Africa at the beginning (by definition) of the Vallesian. The argument focuses on statistics of tooth and limb bone morphology. MacFadden & Skinner, who recognize 4 North American hipparion genera according to cladistic analysis based mainly on facial morphology (mainly the pre-orbital fossa shape and location), hold that hipparion dental and limb bone morphology is too conservative and repetitive for generic-level phylogenetic decisions (see also Woodburne & Bernor 1980; MacFadden & Woodhurne 1982). In this view, a hipparion close to *Cormohipparion sphenodus* (Barstovian-Clarendonian, 13–11 Ma) reached the Old World and there gave rise to various hipparions with the 'Type 1' facial morphology (Woodburne & Bernor 1980), that characterizes all Vallesian hipparions. This was closely followed by the immigration, in early Turolian, of a species of true *Hipparion* closely similar to *H. tehonense* (early Clarendonian, 10–11 Ma). Many additional Old World species and genera evolved from these pioneers by later Miocene time (see MacFadden & Woodburne 1982). More to the point of the *Hipparion* – or hipparion – datum, however, is the fact that both parties agree that *the Vallesian hipparions are monophyletic*, whether belonging to the 'primigenium' morphogroup or to the 'Type 1' facial fossa group.

Berggren & Van Couvering (1978) cited a number of marine-nonmarine correlations and radiometric dates in the Middle Miocene of the Mediterranean-central European area, to narrow down the hipparion datum to a level equivalent to late Serravallian and to calibrate it to 12.5 Ma. Some of these limiting cases have been reinvestigated and their evidence confirmed, for instance Bou Hanifia and Kastellios (see below), which document medial Vallesian hipparion faunas in direct association with earliest Tortonian microfaunas. Further corroboration of this calibration comes from the Eastern Paratethys, where the Vallesian assemblages are equated with the Bessarabian Stage, the base of which is correlated to upper Chron C5A, or 12.5 Ma in the present GPTS (Andreescu 1981, Table 2).

Geochronological arguments in favour of a 10 Ma hipparion datum come from Turkey and Pakistan. In Turkey, continental biostratigraphy is based originally on pollen-zones (*pollenbilder*) that were identified during a lignite resource study (Sickenberg *et al.* 1975). The Yeni-Eskihisar pollen zone comes from paludal sediments with sparse mid-Miocene small mammals and no hipparions among the few large-mammal remains; it is dated as young as 11.1 Ma. The next-youngest pollen zone, Kizilhisar, contains Turolian (not middle-Vallesian as initially stated) mammals at Kayadibi, which is bracketed by tuffs dated 9.1 and 7.95 Ma. Early Vallesian (MN-9) large-mammal faunas, with an hipparion and *Anchitherium*, are found, at Esme-Akcaköy, in fluvial sediments with no pollen and no radiometric ages. If one assumes that the Esme-Akcaköy early Vallesian levels must lie *between* the Yeni-Eskihisar and Kizilhisar pollen zones, a 12.5 Ma hipparion datum is hard to justify. Fortunately, Benda & Meulenkamp (1979) have found both Yeni-Eskihisar and Kizilhisar pollen spectra in lowermost Tortonian pelagic sections of the Aegean. The Yeni-Eskihisar zone occurs on the island of Zakynthos in the lower Tortonian (overlap of lower Zone N16, and Zone CN7; see Fig. 2), while the Kizilhisar zone is found in a number of other lower N16 associations, including that of Kastellios with its mid-Vallesian mammals. The authors concluded that the two pollen zones meet in lower N16, a level which we show (Fig. 2) at *c*.9 Ma, and are consistent with the Turkish ages. Evidently, the mid-Vallesian (Kastellios) is equivalent to early Kizilhisar pollen assemblages, and early Vallesian (Esme-Akcaköy) is time-equivalent to the upper Yeni-Eskihisar pollen zone. The absence of hipparion remains from 11 Ma swamp deposits in Turkey with Yeni-Eskihisar pollen could be due to taphonomical or ecological bias.

In the richly-fossiliferous and well-correlated Siwaliks Series of Pakistan and India (see below), the oldest hipparions are found in normal-polarity beds that clearly belong to lower Chron C5N (Anomaly 5 equivalent), at about 10 Ma (Barndt *et al.* 1978; Tauxe 1979; Barry *et al.* 1982, corrected for the present GPTS). This very well-documented palaeomagnetic 'call' (Johnson *et al.* 1982; Tauxe & Opdyke 1982) has impressed some colleagues as a better standard for calibrating the Old World hipparion datum than the motley assortment of radiometric ages and stratigraphic relationships alluded to in support of the 12.5 Ma datum. The Siwaliks have a fossil mammal faunal that is more equatorial in its relationships than temperate, however, and it is not possible at present to identify a definitive Vallesian assemblage in association with the Siwaliks datum. Instead, MacFadden & Woodburne (1982) have concluded that the earliest Siwaliks hipparion is a *Cormohipparion* without the derived characters of the Vallesian '*primigenium*/Type 1' taxon, and which descended directly from *C. sphenodus*-like ancestors in an independent, south-Eurasian lineage. Its first appearance at *c*.10 Ma in the Nagri faunas would therefore be irrelevant to the time of arrival of other hipparions north and west of the Himalayas.

We note, in passing, that the Siwaliks may have the only *Cormohipparion* in the Old World. If genoholotype *Hipparion prostylum* is set on a separate phylogenetic line from the '*primigenium*/Type 1' taxon (MacFadden & Skinner 1982), the genus *Hippotherium* Kaup 1836, genoholotype *Equus primigenius* von Meyer 1829, is thereby restored, and takes priority over *Cormohipparion* as the name for some, at least, and perhaps all early Vallesian hipparions in the temperate regions of the Old World.

Equus datum

The magnetostratigraphy in superposed faunal sequences of Anza-Borrego (California) and San Pedro Valley (Arizona) document the evolution of *Equus* in middle Blancan faunas in the short normal-polarity zone of the middle Gauss Chron (Johnson *et al.* 1975; p. 325; Opdyke *et al.* 1977) at about 3.1 Ma in the time-scale of this paper, and in good agreement with North American dating of Middle Blancan (Fig. 3).

The immigration of *Equus* in Europe occurs within the MN16b interval of Mein (1979), typified in the Montopoli level of the Val d'Arno (Azzaroli 1977) and bounded by the Etouaires and Roccaneyra local faunas (southern France) near its beginning and end, respectively. Dating of these and other French continental Pliocene sites (Bout 1975; Bandet *et al.* 1978) suggests that MN-16b spans 3.0–2.5 Ma, in round numbers. Lindsay *et al.* (1980) cite evidence that *Equus* did not enter Eurasia until *c.*2.6 Ma, and so far it has been conclusively identified only in the Roccaneyra collection.

MN-16b assemblages, with earliest *Equus*, are found at the base of the Khaprovian 'faunal complex' of the eastern Parathethys (Black Sea–Caspian-Aral basins) according to Gabunya (1981). In the USSR, Lower Khaprovian correlates to upper Akchagylian Stage (Gabunya 1981), and in Romania the Khaprovian fauna is identified with the upper Romanian Stage (Andreescu 1981). Both of these stages begin in the upper Gauss (cf. Andreescu 1981), and the *Equus* datum is therefore somewhat younger, e.g. near the upper limit of the Gauss.

MN-16b assemblages cannot be identified in the Siwaliks, which have an endemic fauna, but the first *Equus* in this region is generally associated with the Tatrot-Pinjor 'stage' boundary. In their study of the Pinjor type section in India, Azzaroli & Napoleone (1981) place this boundary at the base of the Kansal Formation, and locate it in the uppermost Gauss (*c.*2.6 Ma in the present GPTS). Barry *et al.* (1982), working in Pakistan on the Tatrot type area, identified the earliest *Equus* slightly lower in the Upper Gauss normal subzone (Opdyke *et al.* 1982) (*c.*2.8 Ma in the present GPTS), and stratigraphically below tuffs in the lowermost part of the Matuyama magnetozone dated at 2.53 Ma (Johnson *et al.* 1982).

In Africa, the arrival of *Equus* is apparently influenced by ecological barriers. In the equatorial drylands of the Rift Valley, the evidence is unequivocal that *Equus* is not found at levels older than 1.9 Ma. The first *Equus* remains occur synchronously at the base of Bed I Olduvai, in collecting unit 3 below the KBS Tuff at Koobi Fora, and in level G at Shungura (Churcher & Richardson 1978), all of which are dated *c.*1.9 Ma and are correlated to the early Matuyama just below the Olduvai Subchron (Drake *et al.* 1980). The undated 'Equus Tuff' in the Awash Valley sequence of Ethopia is apparently another example of the datum, but Kalb *et al.* (1982) have correlated it to East African faunal levels of about 1.0 Ma.

North America

The standard North American regional land-mammal ages, comprehensively dated by Evernden *et al.* (1964), have recently been reviewed by Tedford *et al.* (1983) to take account of the new information, and new analytical standards, of the 20 years that have passed since that virtually indestructible study. In this context we have included marine-nonmarine 'tie points' in central Florida (Webb *et al.* 1978; MacFadden & Webb 1982) and the Florida Gulf Coast (Hunter & Huddlestun 1982), which support and amplify the 'Barstovian-*Orbulina*' correlation in the Texas coastal plain proposed by R. H. Tedford years ago (see Van Couvering & Berggren 1974).

Direct palaeomagnetic profiling in North American (and South American) mammal-bearing sequences generally has been used to refine K-Ar age calibration estimates, not the other way around. The synthesis of North American mammal biochronology, however, gives a more accurate control for the palaeomagnetic stratigraphy than direct K-Ar age determinations on the local stratigraphy. In the Rio Grande graben, for instance, MacFadden (1977) used local fission-track dates to place the Upper Chamita l.f. within the lower reversed-polarity subchron of the Gilbert, at *c.*5.0 Ma. Tedford (1981) pointed out that this was too young for late Hemphillian (Coffee Ranch equivalent) mammals, and suggested that the correct 'call' would be to the lower part of Chron 6, at *c.*6.5 Ma, in view of the more extended magnetostratigraphy of Rio Grande strata, subsequently obtained by

TABLE 2 Calibration of California benthic zones. Sources: (1) Obradovich & Naeser 1981. (2) Poore *et al.* 1981. (3) Rowell 1981. (4) Keller & Barron 1981. (5) Poore, McDougall *et al.* 1981. (6) Crouch & Bukry 1979; Arnal 1980. (7) This paper. For the correlation of *Stichocorys peregrina* FAD see Weaver *et al.* 1981. For the palaeomagnetic correlation of the diatom zones see Barron 1981. Note:[a] This date refers to the correlation by Armentrout & Echols 1981, Fig. 9, of lower Saltos Shale (near base Relizian in ref. 5 above) to Tranquillon Volcanics. *indicate direct K-Ar dating, (=) is GPTS calibration.

Base of:	*Dated (1)*	*Correlated to:*
Bolivina obliqua zone	close to *6.9 (Palos Verdes)	*S. peregrina* FAD, = 7.0 (1) *N. reinholdii* FAD, = 6.8 (2)
Middle Mohnian (mid *B. hughesi* zone)	*7.8 (stratotype)	*T. antiqua* FAD, = 8.8 (3)
Lower Mohnian	over *11.5 (stratotype) *14.6 (Caliente Rge)	mid-*C. miopelagicus* CN5a, = 14.5 (2) common *D. hustedtii*, = 14.0 (4)
Luisian	over *16.5 (Caliente Rge)	*S. hetermorphus* CN4, = 17.0 (5) lower *D. lauta* 'a', = 16.2 (5)
Relizian	*17.5[a] (Tranquillon)	?up. *S. belemnos* CN2, = 17.8 (5) base *H. ampliaperta* CN3, = 17.6 (6)
Saucesian	*23, avge.	lwr. *C. abisectus* CN1a, = 23.4 (2, 7)

Barghoorn (1981). Similarly, May & Repenning (1982) used the biochronological synthesis to control the palaeomagnetic 'call' on the Warren, Mt. Eden, and Yepomera local faunas. At Anza-Borrego and in the San Pedro Valley, likewise, Opdyke *et al.* (1977) reasoned from biochronological considerations to identify the palaeomagnetic reversals observed in relation to Middle Blancan–Irvingtonian faunal sequences containing the *Equus* datum (see above), althought this again is hardly a test of the internal calibration of the GPTS.

In California coastal basins, a number of mammalian local faunas have been related to marine biostratigraphy in relatively intimate stratigraphic sequences, locally with radiometric control. Savage & Barnes (1972) summarized Miocene marine-nonmarine correlations in a general way, but without correlating the marine sequence to global planktonic microfossil zonation. In view of somewhat more detailed lithostratigraphic studies of the fossiliferous units, as presented by Armentrout & Echols (1981; see also Tedford *et al.*, in press, who adopt the same correlations), and recent revisions to dating and correlation of the California benthic zones to which the mammals are directly related, it is possible to show second-order tie-ins between North American land mammal biochronology and the standard marine microfossil zonation that appear to be relatively sturdy and reliable.

The sequence of California benthic foraminifera stages of Kleinpell, with which the mammal localities have been directly correlated, is now known to overlap and gap over major intervals. Dating by Obradovich & Naeser (1981) confirmed that the type Delmontian Stage, near San Francisco, is the same age as the lower part of the type Mohnian, in Los Angeles, and that the '*Bolivina obliqua*' beds overlying the type Mohnian are therefore not Delmontian but an unassigned, latest Miocene interval. Studies of the California benthic stage faunas associated with planktonic zones in the traditional onshore sections and in offshore cores also indicate considerable overlap, at least in some of the nominally characteristic taxa (Crouch & Bukry 1979; Arnal 1980). The planktonic zonal correlation of the definitive lower limits of each successive stage indicate a GPTS calibration of the sequence that is systematically older than the K-Ar calibration of Turner (1970), but more in line with that indicated by Obradovich & Naeser (1981). The mammal 'tie points' (Appendix III) refer to this new calibration.

South America

K-Ar calibration of the South American land mammal ages is based, so far, on too few dates to adequately test the direct GPTS correlations that have been made on mammal-bearing beds (Marshall *et al.* 1979: Marshall *et al.* 1982; Marshall *et al.* 1983). Also, there are no published correlations between the mammalian and marine microfossil biostratigraphy.

Europe (and North Africa)

In Western Europe and the Mediterranean basin, little has been added to the marine-nonmarine correlations cited in our earlier studies (Van Couvering & Berggren 1977; Berggren & Van Couvering 1978), with regard to Portugal, southern Spain, the Rhone Valley, Tuscany, and Tunisia; we must regard the Rierussa (Valles-Penedes, Barcelona) and Gebel Zelten (Libya) ties as unconfirmed awaiting further study. However, at Bou Hanifia (Ouda & Ameur 1978) and at Kastellios (Bruijn & Zachariasse 1979), follow-up studies have strengthened the marine-nonmarine correlations, and Benda & Meulenkamp (1979) have added a new dimension with trans-facies correlation of pollen zones in the Aegean.

With regard to Early Miocene marine-nonmarine correlations, the classical Aquitanian-Burdigalian-Helvetian equation in SW France has been superseded, perhaps only temporarily, by evidence from the Tajo Basin sequence in Portugal and from the central Paratethys sequence, where direct stratigraphic correlations have been adduced between MN-zones and marine microfossil zones. Current data on the calcareous nannoplankton of the Paratethys sequence suggests strongly that the calibration ages that we proposed earlier for early Miocene mammal history in western Eurasia (Van Couvering & Berggren 1977) are too young, and should be revised older in accordance with the GPTS calibration of the correlated marine microfossil zones.

The calibration of the early Miocene mammal biochronology of Europe and North Africa below MN-5 obviously depends more on correlation to the GPTS than on internal K-Ar dating, and for this reason the early Miocene 'tie points' in the correlation matrix do not prove anything.

The modification suggested for the continental time-scale is of considerable interest in terms of mammalian biogeography and biochronology. It shows that MN-3b, the time of African immigration and the closing of the Tethys in the Zagros zone, begins at 20 Ma or slightly earlier, not *c.*16 Ma (Van Couvering & Berggren, 1977). In the Tajo faunas, the marine correlation of Lisboa IVb is estimated from the FAD of *Gds. altiapertusus* in directly underlying beds, and the presence of *Praeorb. sicana* in directly overlying beds (Antunes *et al.* 1973), although this biostratigraphy is currently being resampled (L. Ginsburg, pers. comm. 1983). On the Paratethys side, the absence of *D. druggii* from upper Eggenburgian samples has been taken as evidence of 'NN3' (CN2) age (Horvath & Nagymarosy 1979), while planktonic foraminifera suggest that upper Eggenburgian is in Zone N5 (Papp 1981, *et auctt.*). Rögl *et al.* (1979) point out that both upper Eggenburgian and Ottnangian exhibit cool-regressive lignitiferous continental facies which are very hard to distinguish without biostratigraphic control. Within this lignitiferous sequence is the only well-dated unit in the European early Miocene, the 'Lower Rhyolite' of Hungary, at 19.6 Ma (Hamor *et al.* 1979; the 21.5 Ma given by Vass 1979, is a preliminary figure). Its conventional assignment to basal Ottnangian without faunal evidence is clearly open to question (Vass 1979), and it is here tentatively referred to as upper Eggenburgian instead, as a provocative possibility.

Asia

In Asia, two regions have a well-developed mammalian biochronology: the Black Sea–Caspian-Aral depression, or 'Eastern Paratethys' of Neogene palaeogeography, and the Siwaliks–Salt Range in the Himalayan foothills. For reasons of space we do not treat Siberia, Mongolia, and China, where Neogene mammal biochronology is (as yet) not well dated, nor Japan and South-east Asia, where the mammal record is mostly Pleistocene.

The Asian biochrons shown on Fig. 3 are those of the Eastern Paratethys, as characterized and correlated by Gabunya (1981); for full faunal lists of the sites in the Black Sea basin see Dubrovo & Kapelist (1979). For the Miocene part, terminology follows the present consensus for 'regional

TABLE 3 Early Miocene correlation matrix in Europe. Tajo basin correlations (Antunes *et al.* 1973; Antunes 1979) estimated according to current standards. MN-zone assignments *fide* Mein (1975; 1979). Paratethys correlations for mammals by Rabeder & Steininger (1975) and Rabeder (1978), for planktonic microfossils from data in Horvath & Nagymarosy (1979), Báldi-Beke & Nagymarosy (1979), Rögl *et al.* (1979) and Lehotayova & Molčikova (1978); Paratethys dating from Vass & Bagdasarjan (1978), Vass (1979), and Hámor *et al.* (1979). Dates with (*) are local K-Ar ages from the indicated stages; those with (=) are from GPTS calibrations. Abbreviations (b. m. u) for basal, and middle, and upper, respectively.

Correl.	*Tajo Basin mammal faunas*	*Mammal zone*	*Cent. Paratethys mammal faunas*	*Correl.*
—	—	MN-6	Neudorf-Sandberg (u. Baden. *15.0)	m-CN5a = 14.0
		MN-6	Kleinhadersdorf (m. Baden)	b-CN5a = 14.5
m-N.9? = 15.2	Qta. da Farinheira (Lisboa Vb)	MN-5	Neudorf-Spallte (b. Baden *16.5)	m-CN4, m-N.8 = 15.5
—	—	MN-4b	Eibiswald (Karpat *18, *c.*)	m-CN3, m-N.7 = 17.3
b-N.8? = 16.8	Qta. Pombeiro (Lisboa Va)	MN-4a	Orechov, Ipolytarnoc (Ottnangian)	?CN2, u-N.6 = 17.8
b-N.6? = 19.0	Qta. do Narigao (Lisboa IVb)	MN-3b	Langau, Tuchorice (u. Eggenb.) (*19.6?)	CN2, m-N.5 = 20.5, *c.*
b-N.4? = 23.5	Horta das Tripas (Lisboa II)	MN-3A	Eggenburg (b. Eggenburgian)	b-CN1c, m-N.4 = 23.0
—	—	MN-1?	Linz (u. Egerian)	m-CP19b? = 25.0

stages' (Papp 1981), which in principle represent both marine and non-marine sequences (Steininger 1977). For the Pliocene and Pleistocene, however, we adopt the 'mammal ages' or 'faunal complexes' recognized by Soviet palaeobiologists (cf. Nikiforova 1977; Gabunya 1981) in place of the Kimmerian, Akchagylian, Apsheronian, and 'Pleistocene' stages. Unfortunately, Miocene 'faunal complexes' have not yet been defined for this region.

The early Miocene marine sediments in the deeper basins of the Eastern Paratethys are well-correlated to those of the Central Paratethys, and to this extent to the open-ocean zonation (Papp 1981). However, the vast sheets of shallow-water and epicontinental strata that spread out over the Ukrainian and Central Asian platforms beyond the deep basins are zoned primarily by endemic shallow-water mollusc faunas (Steininger & Rögl 1979). From the mid-Miocene Volhynian (= early Sarmatian of Central Paratethys) onwards, after the retreat of the Paratethys oceans to the Black and Caspian seas, the endemic molluscs provide the principal zonation in European and Central Asia (Steininger 1977; Andreescu 1981; Papp 1981).

The mollusc-bearing paludal and limnic strata of the Eastern Paratethys Neogene, in turn, are well suited to close correlation with mammal faunas, and also to surface-outcrop palaeomagnetic traverses. Furthermore, several K/Ar dates of mid to late Neogene age have been related, somewhat loosely, to the mollusc zonation (Vass & Bagdasarjan 1978), and at Saro, Eastern Georgia, dates of 10.0 and 10.6 Ma have been obtained in sequence with early Maeotian (early Turolian) mammals (Gabunya & Rubinstein 1977).

Andreescu (1981) reviewed dating, palaeomagnetic analysis, and biostratigraphy in the western Black Sea (Dacian) basin. According to his Table 2, the first 'Néogène Supérieur–Mollusques' zone (NSM_1, *Abra reflexa*), in Volhynian (basal Sarmatian) strata, was most probably coincident with lower 'Chron 14' (the top of the major normal event of Chron C5AD, *c.*14 Ma in the present GPTS), and NSM_4, *Sarmatimactra bulgarica* Zone, in Khersonian (upper Sarmatian) strata, spans the earliest reversed part of Chron C5 (between 11.0 and 11.5 Ma). Gabunya (1981) placed late Volhynian mammals of the eastern Black Sea basin in MN-8, and Khersonian mammals in MN-10; these we have independently calibrated to 13.0 Ma and 10.5 Ma, respectively (Fig. 3), which compares fairly well with the apparent GPTS ages in the coeval molluscan zones. Another association of Khersonian molluscs and MN-10 mammals is described by Nicolas (1978) from the Küçükçekmece sites on the Sea of Marmora, near Istanbul.

Semenenko & Pevzner (1979) obtained palaeomagnetic measurements on mid- to late Miocene marine beds and on upslope shallow-water sections of the Black Sea basin that they interpreted as the record of Chron 7 to Chron 4. At the base of this section, polarities correlated the uppermost part of Chron 7 (*c.*7.0 Ma in age, in the present GPTS) are associated with an Upper Maeotian (Akmanaian) molluscan fauna characteristic of the NSM_6, *Congeria pantacapaea* Zone of Andreescu (1981). The mammal faunas of the Upper Maeotian in this region are independently assigned to MN-13 (late Turolian) age (Gabunya 1981), which we date at *c.*6.8 Ma (Fig. 3). The 'MN-10' *D. neohamatus* – *D. neoerectus* nannoflora reported from this level (cf. Semenenko & Pevzner 1979) is apparently too old to be *in situ*, whereas the higher marine beds have yielded NN-11 to NN-13 nannofloras associated with Chrons 6 to early 4, comparable to the standard deep-sea record.

According to Andreescu (1981) the base of the NSM_{12}, *Ebersiniania milcovensis* zone is within the late Gauss (Kaena) normal interval, and is well correlated to the Upper Romanian Stage of the Dacic Basin, and also to the Upper Akchagylian Stage of central Asia. Mammals found at these levels, both in the Slatina faunas of Romania (Feru *et al.* 1978) and the later Khaprovian complex of Transcaucasia (Gabunya 1981) correlate very closely to the mid-MN16b, Roccaneyra level. The GPTS age of the NSM_{12} zone in the Ponto-Caspian (2.6 Ma) is in exact agreement with dating at Roccaneyra (see '*Equus* Datum' above).

Siwaliks

Recent collecting in the classical Siwaliks sections of the Potwar Plateau has been carefully integrated with extensive palaeomagnetic profiling (Barry, *et al.* 1982; Johnson, *et al.* 1982; Tauxe & Opdyke 1982), such that four newly-established 'mammal zones' based on observed biostratigraphy can be tied directly to the GPTS (see Appendix III). These zones, in turn, are approximately equivalent in time to the Nagri, Dhok Pathan, Tatrot and Pinjor (Barry, *et al.* 1982), considered purely as biochrons divorced from formations.

The problem remains, as before, that the Siwaliks faunas are not easy to correlate to the dated mammalian biochronology in the temperate regions, because the region was part of the 'monsoonal' subtropical province extending from South-east Asia to Arabia and the African Sahel. In the discussion of the '*Hipparion* Datum', we have explained why we assign the early Nagri, represented in the rocks by the base of Barry *et al*'s (1982) '*Hipparion* s. l.' zone in lower Chron C5N, to latest Vallesian or earliest Turolian time. Also, in the discussion of the 'Equus Datum,' we have noted that the first occurrence of *Equus* in the lower part of the '*Elephas planifrons*' zone, in the late Gauss (Azzaroli & Napoleone 1981; Barry *et al.* 1982) identifies the beginning of the Tatrot biochron with mid-Villafranchian, middle MN-16b.

The beginning of the Dhok Pathan biochron can be defined by the immigration of *Selenoportax lydekkeri*, the nominative taxon for the second of the new Siwaliks zones (Barry *et al.* 1982). The lowest remains of *Selenoportax* appear in the middle of Chron 8 (present notation) at a level dated to *c.*7.6 Ma in the present GPTS. The extinction of *Deinotherium* and the appearance of large giraffids in the earliest part of this zone correlate to the middle of MN-12, just above the level of Samos (*c.*8.0 Ma, re-dated *file* N. Solounias), which has *Selenoportax, Deinotherium* and *Helladotherium* (Solounias 1981), but below Maragheh (7.4 Ma–7.0 Ma: Campbell, *et al.* 1980), as well as Los Mansuetos, from which no deinotheres have been reliably documented.

The Tatrot land-mammal age begins with the immigration into the Siwaliks area of the hippopotamus *Hexaprotodon sivalensis*, the nominate taxon for the third Siwaliks zone of Barry *et al.* (1982). The first remains of *Hexaprotodon* occur here in a zone of normally magnetized strata assigned to the upper normal interval of Chron 5 (5.4 Ma in the present GPTS). In SW Europe, hexaprotodonts similar to (or conspecific with) Siwaliks *Hexaprotodon* and those of early Kerian Wadi Natrun and Sahabi faunas (see Fig. 3) immigrated from Africa in latest MN-15 (e.g. at Casino, Alcoy and Venta del Moro) at the end of the Miocene, *c.*5.4 Ma. (cf. Coryndon 1978; Azzaroli & Guazzone 1979). We assume, for the purpose of calibration, that the appearance of the genus in SW Europe was synchronous in the Siwaliks.

Africa

The Neogene fossil mammal record of Africa has excellent radiometric control, particularly in East Africa, but marine-non-marine correlations are few and generally not of good quality (Van Couvering & Van Couvering 1976). The best is at Bou Hanifia, Algeria (Ouda & Ameur 1978), where marine intercalations and a dated tuff can be related to a diverse Vallesian mammal fauna (see Appendix III). At Laangebaanweg, in Saldanha Bay, South Africa, and also at Sahabi, in Libya, poorly-dated early Kerian mammal faunas with transient 'northern tourists' such as bears and peccaries are preserved in geological settings that suggest end-Miocene backfilling of channels incised during Messinian sea-level decline (Hendey 1981), just as in southern Spain (Azzaroli & Guazzone 1979); the Rhone Valley (Clauzon , *et al.* 1982); and Bone Valley, Florida (MacFadden *et al.* 1981). Direct correlation to microfossil zones is missing, however, in the African sites, and tying these mammals to the latest Miocene at 5.4 Ma is based on the assumption that the Messinian regression has been correctly identified. Similarly, coeval North African large-mammal faunas, of late Kisingirian (or early Tinderetian?) age, *c.*16–15 Ma, at Jebel Zelten, Wadi Moghra, and Yeroham (Israel) may have been huried in coastal drainages that were backfilled during the mid-Miocene transgression whose culmination is associated with the *Orbulina* Datum, at *c.*15.0 Ma.

Palaeomagnetic measurements in Africa, including the studies at richly fossiliferous sites of Olduvai, Shungura, Hadar, and Koobi Fora, in East Africa, and Makapansgat in South Africa, are confined to the Pliocene and Pleistocene (see 'Equus Datum' above). Except for Makapansgat, these sequences are also directly dated and the question of comparing time-scales does not arise.

Postscript

In a recent paper on the magnetostratigraphy and K-Ar geochronology of basalts in north-west Iceland, McDougall *et al.* (1984) report an estimate of 11.1 Ma for the age of the older boundary of marine magnetic anomaly 5. However, the study of Tauxe *et al.* (1984) on the magnetostratigraphy of the middle Miocene Ngorora Formation in the East African Rift Valley of Kenya, combined with K-Ar age dates on intraformational tuff beds, supports a much younger age for this boundary of about 10 Ma. The age of 10.42 Ma which we derive for the older boundary of anomaly 5 by extrapolation of sea-floor spreading rates therefore lies at about the average numerical value of these recent radiometrically derived estimates. As Tauxe *et al.* (1985) conclude from this disparity, isotopic ages can have errors far greater than quoted analytical uncertainties, thereby limiting temporal correlation using radiochronology alone to a resolution of perhaps only 10% of the age. Much better temporal resolution, if not accuracy, is obtained using a combination of magneto-bio- and radiochronology.

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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.

TABLE 4 Planktonic foraminiferal biostratigraphy near the Oligocene–Miocene boundary in some DSDP cores.

Site No.	LAD *opima*	FAD *kugleri*	FAD *primordius*	LAD *anguli-suturalis*	FAD *dehiscens*	LAD *kugleri*	O/M Boundary & criterion (a)	Author
55			Mid-14 (125 m)		13/2 (114 m)	12/2 (105 m)	125 m (FAD *primordius*)	Krasheninnikov (Keller)
71	upper 46 (410 m)	top 39 (347 m)	base 36 (326 m)	—	—	top 31 (272 m)	345 m (FAD *kugleri*)	Beckmann
74	basal 6 (54 m)	mid-5 (47 m)	upper 5 (44 m)	—	—	lower 3 (25 m)	47 m (FAD *kugleri*)	
77B	—	basal 34 (318 m)	—	—	31/5 (289 m)	30/5 (277 m)		Jenkins & Orr (Keller)
149	in 29 (251 m)	base 25 (223 m)	base 25 (223 m)	—	in 21 (180 m)	22/6 (195 m)	223 m (FAD *primordius*)	Bolli
151	FAUNAL ASSOCIATIONS ONLY						311 m (LAD cf. *kugleri*)	& Premoli-Silva
171	INADEQUATE DATA							—
173	INADEQUATE DATA							—
214	—	24/5 (225 m)	24/1 (219 m)	—	23/2 (211 m)	23/2 (211 m)	219 m (FAD *primordius*)	McGowran
214	—	24/2 (222 m)	—	—	—	23/2 (211 m)	219 m (near FAD *kugleri*)	Berggren *et al.*
237		21/2 (189 m)	basal 19 (178 m)	—	17/2 (151 m)	19/5 (175 m)	178 m (FAD *primordius*)	Vincent
238	—	—	mid-47 (436 m)	—	—	top 63 (405 m)	mid-47 (FAD *primordius*)	

Table 4 continued.

Site No.	LAD *opima*	FAD *kugleri*	FAD *primordius*	LAD *anguli-suturalis*	FAD *dehiscens*	LAD *kugleri*	O/M Boundary & criterion (a)	Author
289		76/4 (717 m)	69/5 (654 m)	—	—	basal 65 (617 m)	654 m (FAD *primordius*)	Saito
292	lower 21 (193 m)	lower 17 (158 m)	basal 18 (168 m)	basal 18 (168 m)	—	basal 13 (121 m)	168 m (FAD *primordius*)	Ujiie
296	—	—	36/3 (333 m)	—	mid-34 (313 m)	—	333 m (FAD *primordius*)	(Keller)
317B	mid-28 (853 m)	basal 25 (801 m)	basal 24 (782 m)	—	—	—	801 m (FAD *kugleri*)	Kaneps
354	9/10 (406–453m)	—	mid-9 (400 m)	8/9 (349–396 m)	7/8 (292 m)	—	400 m (FAD *primordius*)	
356	—	—	in 5 (90 m)	—	4/5 (66–85 m)	4/5 (66–85 m)	not determined (unconformity)	Boersma
357	18/3 (212 m)	14/15 (160–170 m)	14/1 (152 m)	16/1 (190 m)	13/1 (132 m)	11/2 (105 m)	160–170 m (conc. range *primordius* & *kugleri* in 14)	
362	—	42/43 (767–796 m)	42/43 (767–796 m)	—	40/41 (710–730 m)	42/43 (767–796 m)	767–786 m (FAD *primordius-kugleri*)	Jenkins
363		lower 2 (58 m)	1/4 (37 m)	2 (55 m)	1/4 (37 m)	1/4 (37 m)	55 m (within conc-range of *primordius* & *kugeri*)	Toumarkine
448		in 10 (90 m)	—	in 11 (100 m)	—	top 6 (43 m)	not determined but FAD *kugleri* = in 10 (90 m)	
516F	8/1–130 (237 m)	5/1–110 (208 m)	10/1–100 (256 m)	4/1–110 (198 m)	4/1–110 (198 m)	4/1–110 (198 m)	208 m (FAD *kugleri*)	Berggren
522	21/2 (83 m)	17/3 (68 m)	15/3 (59 m)	—	15/2 (57 m)	—	56–58 m (FAD *G. dehiscens*)	Poore

TABLE 5 Calcareous nannoplankton biostratigraphy near the Oligocene–Miocene boundary in some DSDP cores.

Site No.	LAD *H. recta*	LAD *S. ciperoensis*	LAD *Z. bijugatus*	LAD *R. bisecta*	LAD *R. hesslandii*	LAD *C. abisectus*	FAD *D. druggii*	NP25/NN1 or equivalent	Oligoncene-Miocene boundary	Author
55	INADEQUATE DATA									
71	20-cc (180 m)	39-cc (350 m)		41 cc (372 m)		36–6, 63 (325 m) (Bukry)	30-cc (270 m)	LAD *R. bisecta* (372 m)	LAD *R. bisecta* (372 m)	Haq & Lipps
74	10–1 (82 m)	6–1,140 (47.40 m)		6–1,140 (47.40 m)			3-cc (27 m)	LAD *R. bisecta & S. ciperoensis* (47.40 m)	LAD *R. bisecta* (47.40 m)	Haq & Lipps
77B		37–4,65 (343 m)		37–4,65 (343 m)		34–4,66 (313 m)	33–4,63 (304 m)	LAD *R. bisecta & S. ciperoensis* (343 m)	LAD *R. bisecta* (343 m)	Bukry

TABLE 5 continued

Site No.	LAD *H. recta*	LAD *S. ciperoensis*	LAD *Z. bijugatus*	LAD *R. bisecta*	LAD *R. hesslandii*	LAD *C. abisectus*	FAD *D. druggii*	NP25/NN1 or equivalent	Oligoncene-Miocene boundary	Author
149		27–3,73 (236 m)		27–3,73 (236 m)		27–2,66 (234 m)	23–4,59 (200 m)	LAD *S. ciperoensis & R. bisecta* (236 m)	LAD *R. bisecta & S. ciperoensis* (236 m)	Bukry
149	29–2,25 (252.75 m)	28–2,25 (242.75 m)		higher than 20–3,25 (171.75 m)			23–3,100 (199 m)	LAD *H. recta* (252.75 m)	LAD *H. recta* (252.75 m)	Hay & Beaudry
151		8–1, 81–82 (329.80 m)		4–2,25 (238.75 m)			4–2,141 (240 m)	LAD *S. ciperoensis* (329.81 m)	LAD *S. ciperoensis* (329.81 m)	Hay & Beaudry
161		3–5,38 (24.40 m)		5-cc (44 m)	3–5,38 (24.40 m)	3–2,120 (20.7 m)	2–5,114 (15 m)	LAD *S. ciperoensis R. scrippsae* (24.40 m)	between LAD *C. abisectus* & LAD *S. ciperoensis*	Bukry
165A		2–1,114 (15.14 m)		2–3,85 (17.35 m)		top of hole below LAD *R. abisecta*	top of hole below FAD *D. druggi*	LAD *S. ciperoensis* (15.14 m)	not recovered but at LAD *C. abisectus*	Roth
167		11–5,50 (303.50 m)		12–1,50 (334.5 m)		10–3,50 (263.5 m)		LAD *R. bisecta* (334.5 m)	LAD *C. abisectus* (263.5 m)	Roth
171		4–1,50 (64.5 m)		4–1,50 (64.5 m)	6–2,50 (104 m)	4–1,50 (64.5 m)	3–1,50 (46.50 m)	LAD *S. ciperoensis* (64.5 m)	LAD *C. abisectus, S. ciperoensis, R. bisecta* are simultaneous	Roth
173				35–1,115 (312 m)		34–1,143 (306 m)		LAD *R. bisecta* (312 m)	LAD *R. bisecta* (312 m)	Wise
214		24–2,2 (220.5 m)	24–2,2 (220.5 m)	25-cc (237 m)				LAD *S. ciperoensis* (220.5 m)	LAD *S. ciperoensis* (220.5 m)	Gartner
234				12–1,85 (181.3 m)		10–3,49 (165 m)	6–2,27 (77.75 m)	LAD *R. bisecta* (181.35 m)	LAD *R. bisecta* (165 m)	Roth
236		21–6,99 (195.5 m)		21–6,99 (195.5 m)	28–1,127 (254.77 m)	20–5,98 (184.5 m)	19–3,95 (172 m)	LAD *R. bisecta & S. ciperoensis* (195.5 m)	LAD *C. abisectus* (184.5 m)	Roth
237		INADEQUATE DATA								
238		50–4,114 (467.65 m)		48–1,116 (444.16 m)		44–1,122 (406.22 m)	42–4,44 (389 m)	LAD *S. ciperoensis* (467.65 m)	LAD *C. abisectus* (406.22 m)	Roth
289		79–1,60 (741.6 m)	90–3,126 (849.75 m)	92–1,120 (865.7 m)	91–1,120 (856.2 m)	57–2,30 (533.80 m)	72–3,30 (677.80 m)	LAD *S. ciperoensis* (741.6 m)	undetermined	Shafik
292		18–3,40 (162 m)		18–3,40 (162 m)		17–1,126 (150.25 m)	14–2,40 (122.40 m)	LAD *R. bisecta & S. ciperoensis* (162 m)	LAD *C. abisectus* (150.25 m)	Ellis

Table 5 continued.

Site No.	LAD *H. recta*	LAD *Z. ciperoensis*	LAD *Z. bijugatus*	LAD *R. bisecta*	LAD *R. hesslandii*	LAD *C. abisectus*	FAD *D. druggii*	NP25/NN1 or equivalent	Oligocene-Miocene boundary	Author
296		38–1,96 (349.45 m)		38–1,96 (349.45 m)	48–1,140 (445 m)	34–1,98 (311.45 m)	33–1,105 (302 m)	LAD *R. bisecta* & *S. ciperoensis* (349.45 m)	LAD *C. abisectus* (311.45 m)	Ellis
317B		26–2,95 (237 m)	34–1,30 (310.80 m)				20-cc (187 m)	LAD *S. ciperoensis* (237 m)	LAD *S. ciperoensis* (237 m)	Martini
354	10–3,85 (457.30 m)	9-cc (406 m)		10–1,73 (454.20 m)		9–1,73 (397.20 m)	8–1,61 (340 m)	LAD *S. ciperoensis* (406 m)	LAD *S. ciperoensis* (406 m)	Perch-Nielsen
356						3–2,72 (40.20 m)	4-cc (66.5 m)	unconformity	unconformity	Perch-Nielsen
357		15-cc (179.5 m)	14-cc (160.5 m)	14-cc (160.5 m)		10-cc (103.5 m)	10–1,65 (94.65 m)	LAD *S. ciperoensis* (179.5 m)	LAD *S. ciperoensis* (179.5 m)	Perch-Nielsen
362	41–3,10 (722.6 m)	44-cc (796 m)		41-cc (729.5 m)		33-cc (530.5 m)	40-cc (710.5 m)	LAD *S. ciperoensis* (796 m)	FAD *D. druggii* (710.5 m)	Proto-Decima
363	2–2,10 (51.60 m)	2-cc (59.5 m)	1–4,101 (36.5 m)	1-cc (40.5 m)		1–2,60 (33.10 m)	1-cc (40.5 m)	LAD *H. recta* (51.60 m)	unconformity NN1 missing	Proto-Decima
386	4–4,93 (152.64 m)	4–4,93 (152.64 m)	11-cc (261 m)	4–5,80 (154 m)	5–1,80 (157.60 m)	4–4,93 (152.64 m)	4–5,80 (154 m)	LAD *R. scrippsae* (157.60 m)	between LAD *C. abisectus* & FAD *D. druggii* but undetermined	Thierstein & Okada
406	28–1,35 (255.35 m)		31–3,112 (287.60 m)			16–1,100 (142 m)		LAD *H. recta* (255.35 m)	LAD *H. recta* (255.35 m)	Muller
407	31–1,52 (283.5 m)		28–3,90 (258.4 m)	31-cc (291 m)	31–1,52 (283.5 m)			LAD *H. recta* (283.5 m)	LAD *H. recta* (283.5 m)	Steinmetz
407	43-cc (399 m)		30–1,44 (275 m)			18–3,10 (169.10 m)	21-cc (202 m)	LAD *Z. bijugatus* (275 m)	LAD *Z. bijugatus* + *D. enormis* (275 m)	Martini
445		45-cc (426.50 m)	43-cc (407.50 m)	41–6,42 (386.90 m)	41–3,42 (382.42 m)	33-cc (312 m)	41–1,42 (379.42 m)	LAD *S. ciperoensis* (426.50 m)	FAD *D. druggii* (379.42 m)	Okada
448		10–2, 6 (83.1 m)	10–2,6 (83.1 m)			1-cc (5 m)	5-cc (24 m)	LAD *S. ciperoensis* (83.1 m)	LAD *S. ciperoensis* (83.1 m)	Martini
516F (516)	516-n 39–1,80 (136 m)	9–1,98 (246 m)	7–1,96 (220 m)	5–1,100 (208 m)		516–12–1 (47 m)	3–5,100 (198 m)	LAD *R. bisecta* (208 m)	LAD *R. bisecta* (208 m)	Aubry
522	—	15/16 (65 m)	—	15/2 (57 m)	15/3 (59 m)	15/2 (57 m)	14/2 (52 m)	LAD *R. scrippsae* + *R. bisecta* + *C. abisectus* (56–58 m)	LAD *R. scrippsae* + *R. bisecta* + *C. abisectus* (56–58 m)	Percival

Appendix II.

TABLE 6 Relationship of Neogene planktonic foraminiferal datum levels to observed magnetic polarity stratigraphy. Age estimates (Ma) derived from revised geomagnetic polarity time scale presented in this paper (Table 1). These data have provided the basic magnetobiochronologic framework for estimating the chronology of standard time-stratigraphic units and stage stratotypes.

Early Miocene

Datum	Chron	Magnetic polarity Chron and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. FAD *Praeorbulina sicana*	C5C	mid-part of anom. 5C	16.6	2,3,6	ranges down to at least top of Chron C5CN in Site 521 (ref. 4)
2. FAD *Globorotalia miozea*	C5C	lower of 3 normal events of anom. 5C	16.8	2,6	
3. LAD *Globorotalia zealandica*	C5C	"	16.8	2	
4. LAD *Globorotalia pseudomiozea*	C5C	"	16.8	2	
5. FAD *Globorotalia birnageae*	C5C	base anom. 5C	17.0	2	
6. LAD *Catapsydrax Stainforthi*	C5C		17.4	2	
7. FAD *Globorotalia zealandica*	C5D	top anom. 5D	17.6	2	
8. FAD *Globorotalia pseudomiozea*	C5D	top anom. 5D	17.6	2	
9. LAD *Globorotalia semivera*	C5D	"	17.6	2	
10. LAD *Globorotalia incognita*	C5D	"	17.6	2	
11. LAD *Catapsydrax dissimilis*	C5D	"	17.6	2	
12. FAD *Globorotalia praescitula*	C5D	top anom. 5D	17.7	2	
13. LAD *Globoquadrina dehiscens* (f. *spinosa*)	C5D	between anom. 5D and 5E	18.3	2,3	observed in South Atlantic only to date.
14. FAD *Globigerinoides altiaperturus.*	C6A	top anom. 6A	20.9	2,3	
15. LAD *Globorotalia kugleri.*	C6A	between base 6A and top unnumbered anomaly between 6A and 6B	21.8	2,3,6	
16. FAD *Globoquadrina dehiscens* (f. *spinosa*)	C6AA	anomaly between 6A and 6B	22.0	2,3	
17. LAD *Globoquadrina globularis*	C6B	anom. 6B	22.8	2	
18. FAD *Globoquadrina dehiscens*	C6B	between 6B and 6C	23.2	2–4,6	
19. FAD *Globorotalia incognita.*	C6B	"	23.2	2,3	
20. LAD *Globigerina angulisuturalis.*	C6B	"	23.2	2	
21. LAD *Globorotalia pseudokugleri*	C6B	"	23.2	2	
22. FAD *Globorotalia kugleri*	23 (=C6C)	mid–6C	23.7	2,6	recorded also between Chrons C6CN and C7 at Site 522 in South Atlantic (ref. 4) and Mediterranean region (ref. 5)
23. LAD *Globorotalia mendacis.*	23 (=C6C)	"	23.7	2	

Early Miocene

Datum	Chron	Magnetic polarity Event and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
24. FAD *Globigerinoides primordius* (common)	C6C	just below 6C in reversed interval	24.3	2,3	
25. FAD *Globigerinoides primordius* (rare)	C7	anom. 7	25.8	2,3	

Refs.: 1. Ryan *et al.* (1974)
2. Berggren *et al.* (1983)
3. Pujol (1983)
4. Poore *et al.* (1982, 1983)
5. Lowrie, Alvarez *et al.* (1982)
6. Miller *et al.* (in press)

Middle Miocene

Datum	Chron	Magnetic polarity Event and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Globorotalia siakensis*	C5	base anom. 5	10.4	1	
2. FAD *Globigerina nepenthes*	C5	just above 5A	11.3	1	identified between lower of two normal Subchrons interpreted as being in Chron 12 and top Chron 13 (ref. 2).
3. LAD *Globorotalia fohsi lobata* and *G. fohsi robusta*	C5	top 5A	11.5	1	
4. FAD *Globorotalia fohsi robusta*	C5A	normal Subchron, lower 5A	12.6	1	
5. FAD *Globorotalia fohsi lobata.*	C5AA	reversed Subchron, base 5AA	13.1	1	
6. FAD *Globorotalia praefohsi*	C5AC	normal Subchron, 5AC	13.9	1	
7. LAD *Globorotalia peripheroronda*	C5AD	lower part normal Subchron	14.6	1	
8. FAD *Globorotalia peripheroacuta*	C5B	normal Subchron, 5B	14.9	1.2	This is same taxon (i.e. datum level) as *G. fohsi* (no. 35) of ref. 2.
9. LAD *Praeorbulina sicana*	C5B	normal Subchron 5B	14.9	1	LAD *P. sicana* occurs in interval of no paleomagnetic data about 20 m above FAD *P. sicana* which occurs in anom. 5C correlative in Hole 516 (ref. 3).
10. LAD *Praeorbulina glomerosa*	C5B	normal Subchron 5B	14.9	1	Same remarks as above (9).
11. FAD *Orbulina suturalis*	C5B	anom. 5B	15.2	1,5	FAD *Orbulina* occurs in interval of no paleomagnetic data about 20 m above FAD *Praeorbulina sicana* which occurs in anom. 5C correlative in Hole 516F (ref. 3).
12. FAD *Praeorbulina circularis.*	C5C	top anom. 5C	16.2	1	sequential appearance of praeorbulinid taxa occurs in "normal" (DSDP Hole 563) and expanded (DSDP Hole 558) sequence of anomaly 5C correlative in North Atlantic (ref. 1)
13. FAD *Praeorbulina glomerosa*	C5C	near top 5C	16.3	1	occurs in Hole 516 about 15 m below FAD *Orbulina suturalis* in interval with no paleomagnetic data and about 5 m above FAD *P. sicana* in mid-part of Chron C5CN (ref. 3).
14. FAD *Globigerinoides diminutus*	(C5C)	(near top 5C)	16.3	3,4	see Remarks above (13).
15. FAD *Praeorbulina curva.*	C5C	mid-part anom. 5C	16.5	1	see Remarks under 12 above.
16. FAD *Praeorbulina sicana*	C5C	mid-part anom. 5C	16.6	1,3,4	ranges down to at least top of Chron C5C in Site 521 (ref. 5).

Refs.: 1. Miller *et al.* (in press)
2. Ryan *et al.* (1974)
3. Berggren *et al.* (1983)
4. Pujol (1983)
5. Poore *et al.* (1982, 1983)

* (C5C) = not directly correlated to polarity interval but certainly within this interval based on bracketing data.

Late Miocene

Datum	Chron	Magnetic polarity Event and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Globoquadrina dehiscens*	GIL-BERT (4)	lowermost Gilbert	5.3	1–4	
2. LAD *Globorotalia lenguaensis*		lowermost Gilbert	5.3	3,4	located in 'Epoch 7' in Site 519 (South Atlantic, ref. 8) but this interpretation is equivocal.
3. FAD *Globorotalia cibaoensis*	5	lowermost Gilbert	5.3	3,8	
4. FAD *Globorotalia margaritae*	5	mid-5 between events a and b	5.6	1,2,4,7	FAD usually recorded at or slightly above a level thought (by indirect correlation) to lie in basal Gilbert (4) Chron. Recorded by Saito (ref. 2) in RC-12–66 (Equatorial Pacific) and Pujol (ref. 4) in DSDP Hole 516A (Rio Grande Rise, South Atlantic) in Chron 5, and and Cita (ref. 7) in DSDP Site 397 (off Cape Bajador, W. Africa). Recorded in basal Gilbert in Hole 502 (ref. 10).
5. FAD *Pulleniatina primalis*	5	anom. 3A	5.8	2	Indo-Pacific only. Recorded in mid-Chron 5 in Hole 503 (ref. 10).
6. *Neogloboquadrina acostaensis* (S to D)	5	anom. 3A	5.8	2,7	Suggested to have occurred in Mediterranean at a level younger than Gilbert/Chron 5 boundary = 5.3 Ma (ref. 11) but this may be due to misinterpretation of local (Cretan) paleomagnetic stratigraphy.
7. FAD dextral menardine Globorotaliids (= *G. menardii* form 5)	6	younger reversed interval	6.0	11	Mediterranean region. However, this may, in fact, be within Chron 6R (early part) if Cretan magnetic stratigraphy has been misinterpreted (see text for further discussion).
8. *Neogloboquadrina acostaensis* (D to S)	6	just below chron 5/6 boundary	6.1	2	
9. FAD *Globorotalia conomiozea*	6	younger reversed interval	6.1	5,6	Recorded in Crete from reversed interval of Chron 5 at 5.6 Ma (ref. 11). However, this may be due to misinterpretation of Cretan magnetostratigraphy (see text for further discussion).
10. LAD sinistral menardine Globorotaliids (*G. menardii* form 4)	6	early reversed interval near base Chron 6	6.6	11	This group includes *G.* miozea s.l. of Berggren (1977) which showed correlative disappearance at Rio Grande Rise in late Miocene. The exit level in Crete may actually lie within Chron 7 (rather than 6) if the Cretan magnetostratigraphy has been incorrectly identified (ref. 11) (see text for further discussion).
11. FAD *Neogloboquadrina humerosa*	8	reversed interval between 4 and 4A	7.5	6	Chron 8 of ref. 6 is reinterpreted to lie between anomalies 4 and 4A accounting for younger age estimate. Recorded in Sidufjall (C2) Subchron of Gilbert Chron (*c*.4.7 Ma) in Pacific (ref. 2) and eastern Atlantic (ref. 7) areas.
12. FAD *Neogloboquadrina acostaensis*	11	lower anom. 5	10.2	9	identified in normal Subchron identified as occurring in mid-Chron 10 in stratotype Tortonian (ref. 6).

Refs.:
1. Berggren (1973)
2. Saito *et al.* (1975)
3. Berggren *et al.* (1983)
4. Pujol (1983)
5. Kennett & Watkins (1974)
6. Ryan *et al.* (1974)
7. Mazzei *et al.* (1979)
8. Poore *et al.* (1983)
9. Miller *et al.* (in press)
10. Keigwin (1982)
11. Langereis *et al.* (1984)

Early Pliocene

Datum	Chron	Magnetic polarity Chron and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Pulleniatina primalis*	G	reversed interval, latest Gilbert	3.5	9	Pacific Ocean only
2. *Pulleniatina* (S to D coiling change)	I L	just above Cochiti Subchron (a)	3.8	2,9	Recorded above Nunivak Subchron at level which should be equivalent to Cochiti Subchron (but not observed) on Fiji (ref. 10).
3. LAD *Globigerina nepenthes*	B E	top Cochiti Subchron (a)	3.9	1–4,8	Same as above.
4. LAD *Pulleniatina spectabilis*	R T (4)	top Cochiti Subchron (a)	3.9	1	Indo-Pacific only; FAD *P. spectabilis* recorded at Site 503 (eastern Pacific) at about 4 Ma but this is a lone occurrence and probably not a true FAD (ref. 9).
5. FAD *Globorotalia crassaformis* s.s.	G I	within Cochiti (a) and Nunivak (b) combined	4.1	3	
6. FAD *Globorotalia crassaformis* s.l.	L B	just above Sidufjall (C_1) Subchron	4.3	3–4,9	
7. LAD *Globigerinoides seiglei*	E R	just above Sidufjall (C_1) Subchron	4.3	3,4,8	LAD *G. seiglei* observed in Hole 502 (Caribbean) within Thvera Subchron (ref. 9).
8. FAD *Globorotalia puncticulata*	T	Sidufjall (C_1) Subchron	4.4	3–5,8	
9. LAD *Globorotalia cibaoensis*	(4)	Sidufjall (C_1) Subchron	4.4	3,6	
10. FAD *Sphaeroidinella dehiscens.*	G I	middle of lower reversed Gilbert	5.1	1,2	Recorded at various levels in Holes 502 and 503 in basal Gilbert depending on sample intervals (ref. 9).
11. FAD *Globorotalia tumida*	L	lower reversed Gilbert	5.2	2,9	Predominantly Indo-Pacific; recorded in Caribbean Hole 502 also (ref. 9).
12. FAD *Pulleniatina spectabilis*	B	lower reversed Gilbert	5.2	1,2	Indo-Pacific only
13. FAD *Globigerinoides conglobatus*	E	lowermost Gilbert	5.3	3,4,6	
14. FAD *Globorotalia cibaoensis*	R	lowermost Gilbert	5.3	3,6	
15. LAD *Globoquadrina dehiscens*	T (4)	lowermost Gilbert	5.3	1–4	Recorded near base (just below) Sidufjall Subchron on Fiji (ref. 10).
16. LAD *Globorotalia lenguaensis*		lowermost Gilbert	5.3	3,4	

Refs.:
1. Hays *et al.* (1969)
2. Saito *et al.* (1975)
3. Berggren *et al.* (1983)
4. Pujol (1983)
5. Mazzei *et al.* (1979)
6. Poore *et al.* (1983)
7. Ryan *et al.* (1974)
8. Leonard *et al.* (1983)
9. Keigwin (1982)
10. Rodda *et al.* (in press)

Late Pliocene

Datum	Chron	Magnetic polarity Chron and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. FAD *Globorotalia truncatulinoides.*	GAUSS	just below Olduvai event (= anom. 2)	1.9	1–6,9,10	
2. LAD *Globorotalia exilis*		Reunion event	2.1	3–6	Atlantic Ocean only
3. LAD *Globorotalia miocenica*		just below Reunion event	2.2	3–6,9	Atlantic Ocean only
4. (local) reappearance of *Pulleniatina*		just below Reunion event	2.2	2,10	Atlantic Ocean only
5. LAD *Globorotalia pertenuis*	(3)	top Gauss normal	2.5	10	Recorded in Caribbean (Site 502) and Pacific (Site 503) (ref. 10).
6. LAD *Globorotalia multicamerata*	GAUSS	between Kaena and Mammoth	2.9	3,4,9	
7. LAD *Globoquadrina altispira*		between Kaena and Mammoth	2.9	1–4,7,9	
8. FAD *Globigerinoides fistulosus*		between Kaena and Mammoth	2.9	1	
9. LAD *Sphaeroidinellopsis* spp.		top Mammoth event	3.0	1–4,7,9	
10. FAD *Sphaeroidinella dehiscens* s.s.		Mammoth event	3.0	1–4,9	
11. FAD *Globorotalia inflata*	(3)	Mammoth event	3.0	3,4	
12. LAD *Globorotalia conomiozea*		Mammoth event	3.0	3	
13. FAD *Globorotalia tosaensis*	MATUYAMA	just below Mammoth	3.1	1	
14. FAD *Globorotalia crassula*		just below Mammoth	3.1	3	
15. (local) LAD *Pulleniatina*		between Mammoth and Gauss/Gilbert boundary	3.3	2,10	Atlantic Ocean only
16. FAD *Globorotalia pertenuis*		between Mammoth and Gauss/Gilbert boundary	3.3	10	Recorded in both Caribbean (Site 502) and Pacific (Site 503) (ref. 10).
17. FAD *Globorotalia miocenica*	(2)	between Mammoth and Gauss/Gilbert boundary	3.4	3,4,7	Atlantic only
18. LAD *Globorotalia margaritae*	GIL-BERT (4)	Gauss(4)/Gilbert(5) boundary	3.4	2	LAD recorded in reversed interval, just below Gauss/Gilbert boundary in Hole 502 (Caribbean Sea) (ref. 10).

Refs.:
1. Hays *et al.* (1969)
2. Saito *et al.* (1975)
3. Berggren *et al.* (1983)
4. Pujol (1983)
5. Haq *et al.* (1977)
6. Berggren *et al.* (1980)
7. Mazzei *et al.* (1979)
8. Poore *et al.* (1983)
9. Leonard *et al.* (1983)
10. Keigwin (1982)

Pleistocene

Datum	Chron	Magnetic polarity Chron and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Globoquadrina pseudofoliata*	BRUNHES	δ^{18}O stage 7	0.22	1,2	
2. LAD *Globorotalia tosaensis*		lower Brunhes	0.6	5	
3. FAD *Pulleniatina finalis*		midway between Jaramillo and Olduvai Subchrons	1.3	3,4	
4. LAD *Globigerinoides fistulosus*	MATUYAMA	just above Olduvai Subchron	1.6	2–4	
5. LAD *Globigerinoides obliquus extremus*		Olduvai Subchron (anom. 2)	1.8	2,5,6	
6. FAD *Globorotalia truncatulinoides*		just below base Olduvai	1.9	2–4,6	

Refs.: 1. Thompson & Saito (1974)
2. Berggren *et al.* (1980)
3. Hays *et al.* (1969)
4. Saito *et al.* (1975)
5. Thompson & Sciarillo (1978)
6. Keigwin (1982)

Appendix II

TABLE 7 Relationship of Neogene calcareous nannoplankton datum levels to observed magnetic polarity stratigraphy. (Further explanation as in Table 6.)

Early Miocene

Datum	Chron	Magnetic polarity Chron and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Helicosphaera ampliaperta*	C5C	top anom. 5C	16.2	5	located at base Chron C5CN in Hole 516 (ref. 1).
2. FAD *Sphenolithus heteromorphus*	C5C	just below anom. 5C	17.1	1	
3. LAD *Sphenolithus belemnos*	C5C	reversed interval just above 5D	17.4	1	indirectly correlated to Chron C5D–C5E boundary elsewhere (see ref. 3).
4. FAD *Sphenolithus belemnos*	C6A	mid-part anom. 6A	21.5	5	observed in early part of Chron C5DN in Hole 516 (ref. 1).
5. FAD *Discoaster druggi*	C6C	between anom. 6B and 6C	23.2	1,2,5	
6. LAD *Reticulofenestra bisecta*	23 (=C6C)	anom. 6C	23.7	1,2	LAD *R. bisecta* recorded between Chron C6C and C7 in Mediterranean (ref. 4).
7. LAD *Reticulofenestra scrippsae* (= *hesslandi*)	23 (=C6C)	anom. 6C	23.7	2	
8. LAD *Reticulofenestra abisecta*	23 (=C6C)	anom. 6C	23.7	2	May be local LAD inasmuch as this taxon is reported elsewhere at levels as high as middle Miocene; LAD *abisecta* recorded just below Chron C6CN in Mediterranean (ref. 4).

Ref.: 1. Berggren *et al.* (1983)
2. Poore *et al.* (1982, 1983)
3. Ryan *et al.* (1974)
4. Lowrie, Alvarez *et al.* (1982)
5. Miller *et al.* (in press)

Middle Miocene

Datum	Chron	Magnetic polarity Chron and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. FAD *Catinaster coalitus*	11 (C5)	reversed interval below 5	10.8	1	range of *C. coalitus* restricted to Chron C5N in Hole 519 (initial occurrence several meters above contact with basement; ref. 2) and 521A (ref. 4).
2. LAD *Cyclococcolithus floridanus*	C5A	top anom. 5A	11.6	1	LAD of *C. floridanus* often used as surrogate criterion for NN6–NN7 boundary. However, *C. floridanus* occurs within Zone NN7 as defined by the FAD of *D. kugleri* in some areas, including Holes 558 and 563 (ref. 1).
3. FAD *Discoaster kugleri*	C5AB	within interval of C5AA-C5AB	13.1 ca.	1	FAD of *D. kugleri* not precisely located in 558 and 563. NN6–NN7 boundary arbitrarily drawn between C5AA and C5AB in figure 2.
4. FAD *Triquetrorhabdulus rugosus*	C5AC	base normal Subchron, 5AC	14.0	1	
5. LAD *Sphenolithus heteromorphus*	C5AD	normal event, 5AD	14.4	1	LAD occurs at essentially same level as FAD *Orbulina suturalis* in interval of no paleomagnetic data about 20 m *above* Chron C5CN in Hole 516 (ref. 3).

Ref.: 1. Miller *et al.* (in prep.)
2. Poore *et al.* (1983)
3. Berggren *et al.* (1983)
4. Salis (1983)

Late Miocene

Datum	Chron	Magnetic polarity Chron and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. LAD *Discoaster quinqueramus*	5	Event A	5.6	1,2	recorded in lower Gilbert (ref. 7).
2. LAD *Discoaster berggrenii*	5	Event A	5.6	2	
3. LAD *Amaurolithus amplificus*	5	Event A	5.6	3	The FAD & LAD of *A. amplificus* is recorded from a level equivalent to Chron 6 (in a palaeomagnetic gap) & Subchron B respectively in the Mediterranean (ref. 2; = *c*.6.6 Ma–5.8 Ma).
4. FAD *Amaurolithus amplificus*	5	Event B	5.9	3	
5. FAD *Amaurolithus tricorniculatus*	5	Event B	6.0	3	FAD *A. tricorniculatus* recorded in Subchron A (= *c*.5.5 Ma) in Mediterranean (ref. 2) & apparently at the same level (along with FAD *A. delicatus* & *A. primus*) in Hole 516 F (Rio Grande Rise) at *c*.5.5 Ma (ref. 4). (See items 6, 7 below.)
6. FAD *Amaurolithus primus*	6		6.5	2,3	See remarks under item 5, above. Recorded in upper Chron 5 at Site 577 (ref. 7). Recorded just above Thvera Subchron on Fiji (ref. 8).
7. FAD *Amaurolithus delicatus*	6		6.5	2,3	See remarks under item 5, above.
8. FAD *Discoaster quinqueramus*	8 (10 of this paper)	mid-anom. 4A	8.2	3,6	
9. LAD *Discoaster bollii*	8 (10 of this paper)	(lower 4A)	8.3	6	LAD of *D. bollii* found in interval of no palaeomagnetic data interpreted here as probably equivalent to lower part of anomaly 4A correlative (ref. 6).
10. LAD *Catinaster calyculus*	8 (10 of this paper)	top normal Subchron just above anom. 5	8.75	6,7	
11. LAD *Discoaster exilis*	8 (10 of this	base normal Subchron just above anom. 5	8.8	6	
12. LAD *Discoaster hamatus*	11 (C5)	reversed interval just above anom. 5	8.85	6	
13. LAD *Catinaster coalitus*	11 (C5)	uppermost anom. 5	9.0	6	
14. FAD *Discoaster hamatus*	11 (C5)	lower part anom. 5	10.0	6	
15. FAD *Catinaster calyculus*	11 (C5)	lower part anom. 5	10.0	6	FAD *C. calyculus* reported subsequent to LAD *C. coalitus* in Hole 519 (ref. 7); short overlap observed in Hole 563 (ref. 6). FAD of *C. calyculus* subsequent to that of *D. hamatus* used to differentiate upper subzone b of Zone CN7. In holes 558 and 563 these two taxa have simultaneous FADs which precludes subdivision of Zone CN7 (see fig. 2).

(lower 4A) = interval of no palaeomagnetic data; identification tentative.

Refs.: 1. Gartner (1973)
2. Mazzei *et al.* (1979)
3. Haq *et al.* (1980)
4. Berggren *et al.* (1983)
5. Ryan *et al.* (1974)
6. Miller *et al.* (in prep.)
7. Poore *et al.* (1983)
8. Rodda *et al.* (in press)

Early Pliocene

Datum	Chron	Magnetic polarity Chron and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. FAD *Pseudoemiliania lacunosa*	G I	top Gilbert	3.4	8,11	
2. LAD *Sphenolithus abies & S. neoabies*	L B E R T	late Gilbert, between top of chron and Subchron A	3.47	9,11	Found in 2 cores lacking magnetostratigraphy at same level (by correlation; ref. 10). Recorded above Nunivak and above level where Cochiti Subchron expected (but not recorded) on Fiji (ref. 12).
3. LAD *Reticulofenestra pseudoumbilica*	(4) G I L B E	between base Gauss and Gilbert Subchron A	3.5	3,8,9,10	References to younger (*c*.3.0 Ma) LAD of this taxon may be due to taxonomic problem of younger extension of small, atypical specimens; see discussion in refs 4, 10. Also cited just above Cochiti Subchron (ref. 9). LAD *R. pseudo-umbilica* recorded in interval above Nunivak and above level where Cochiti Subchron expected (but not recorded) and approximately equivalent to LAD *S. abies* (see remarks under Datum 2 above) on Fiji (ref. 12).
4. FAD 'small' *Gephyrocapsa*	R T (4)	upper part of Gilbert	3.6	11	Recorded in DSDP Site 503 and cross correlated to otber DSDP holes and to a level near the Zandian–Piacenzian boundary in Italy (ref. 11).
5. LAD *Amaurolithus tricorniculatus*	G I L B E	between Mammoth and Cochiti Subchrons	3.7	3–5,8	Associated with C_1 and C_2 (Sidufjall and Thvera) Subchrons (undifferentiated) in DSDP Site 297 (off Cape Bojador (*c*.4.5 Ma) (ref. 6) and between Nunivak and Sidufjall Subchrons (ref. 9) and just above Nunivak Subchron in Fiji (ref. 12).
6. LAD *Amaurolithus delicatus*	R T	"	3.7	3–6,8,10	
7. FAD *Discoaster tamalis.*		just above Cochiti Subchron	3.8	6,8	Recorded near Gauss–Gilbert boundary in eastern Pacific Site 503 (ref. 11). Recorded above Nunivak Subchron at a level equivalent to where Cochiti Subchron expected (but not recorded) on Fiji (ref. 12).
8. FAD *Discoaster asymmetricus*	(4)	between Cochiti and Nunivak Subchrons	4.1	3–6	
9. LAD *Amaurolithus primus*		near top of Sidufjall (C_1)	4.4	6	
10. FAD *Ceratolithus rugosus*		between C_1 & C_2 (Sidufjall Thvera) Subchrons	4.5	3–6,8,10, 11,12	Recorded in mid-Sidufjall (ref. 9). See discussion of redefinition of datum event in ref. 10. Recorded just above Thvera Subchron in Fiji (ref. 12).
11. LAD *Ceratolithus acutus.*	G I L B E R T	between C_1 & C_2 (Sidufjall & Thvera) Subchrons	4.57	5,6,9	*C. rugosus* evolved from *C. acutus*. The respective datum levels of items 9 & 10 are placed at the frequency crossover of these two taxa (ref. 10, 11). Recorded in basal Gilbert in Site 503 (ref. 11). LAD *C. acutus* recorded just above Thvera Subchron on Fiji (ref. 12).
12. FAD *Ceratolithus acutus*		early Gilbert	5.0	3–5,7	associated with gap in paleomagnetic record between Subchrons C_1 & C_2 (above) and Chron 5 (below) = early Gilbert = *ca.* 5.0 Ma at DSDP Site 397 (ref. 6).
13. LAD *Triquetrorhabdulus rugosus*	(4)	early Gilbert	5.0	6	see remarks under item 12 ahove.

Refs.:
1. Berggren *et al.* (1980)
2. Haq *et al.* (1977)
3. Gartner (1973)
4. Berggren (1973)
5. Berggren *et al.* (1983)
6. Mazzei, R. *et al.* (1979)
7. Haq *et al.* (1980)
8. Poore *et al.* (1983)
9. Monechi *et al.* (in press)
10. Backman & Shackleton (1983)
11. Rio (1982)
12. Rodda *et al.* (in press)

Late Pliocene

Datum	Chron	Magnetic polarity Chron and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. FAD *Gephyrocapsa oceanica*	M	just above Olduvai Subchron (anom. 2)	1.68	1,2,8	
2. FAD *Gephyrocapsa caribbeanica*	A T	just below top olduvai Subchron (anom. 2)	1.74	1,2	
3. LAD *Discoaster brouweri* and *D. triradiatus*	U Y A	just below Olduvai Subchron	1.90	3,6,7,8	Near base or within Olduvai Subchron (ref. 1, 2).
4. FAD *Gephyrocapsa aperta*	M A	below Olduvai Subchron	2.2	1,2	
5. LAD *Discoaster asymmetricus*	(2)	below Olduvai Subchron	2.2	7	
6. LAD *Discoaster pentaradiatus*		reversed interval between Reunion Subchron and top Gauss	2.4	1–8	
7. LAD *Discoaster surculus*	G	reversed interval between Reunion Subchron and top Gauss	2.4	1–8	
8. LAD *Discoaster tamalis*	A U	latest Gauss	2.6	6,7	Equatorial region only
9. LAD *Discoaster decorus*	S S (3)	just above Kaena Subchron	2.90	6	First common occurrence recorded near Gauss/Gilbert boundary in Site 503, but sporadically below this level (ref. 8).
10. LAD *Discoaster variabilis*	GAUSS (3)	just above Kaena Subchron	2.90	7	

Refs.: 1. Haq *et al.* (1977)
2. Berggren *et al.* (1980)
3. Backman *et al.* (1983)
4. Gartner (1973)
5. Berggren (1973)
6. Monechi *et al.* (in press)
7. Backman and Shackleton (1983)
8. Rio (1982)

Pleistocene

Datum	Chron	Magnetic polarity Event and/or Anomaly Correlative	Age (Ma)	Ref.	Remarks
1. *Gephyrocapsa caribbeanica*/ *Emiliania huxleyi* dominance-reversal	B R U	1. δ^{18}O stage 4 2. δ^{18} substages 5a–5b	0.073 0.085	1,2	Datum event occurs earlier in tropical (low latitude) waters than in transitional (mid-latitude) waters.
2. FAD *Emiliania huxleyi*	N H	δ^{18}O stage 8	0.275	1,2	
3. LAD *Pseudoemiliania lacunosa*	E S (1)	δ^{18}O stage 12	0.474	1,2	
4. LAD *Helicosphaera sellii*			1.37	5,6	Equatorial region only. Apparently diachronous elsewhere (ref. 5).
5. LAD *Calcidiscus macintyrei*		between Olduvai and Jaramillo	1.45	3	recorded at top Olduvai Subchron at Site 577 (ref. 4).

Refs.: 1. Thierstein, J.R. *et al.* (1977)
2. Berggren, W.A., *et al.* (1980)
3. Backman *et al.* (1983)
4. Monechi *et al.* (in press)
5. Backman and Shackleton (1983)
6. Rio (1982)

Appendix III

'Tie points' in the comparison of continental biochronology to the Neogene and geomagnetic-polarity time-scale

Equus Datum
Earliest *Equus*, mid MN-16b [2.6] = mid upper Gauss [2.6]
Siwaliks; Ponto-Caspian region; France: immigration of *Equus* corresponds to levels with K/Ar ages and MPTS 'dates' in synchrony (4; 6; 11; 20)

Pinjor
Base 'Elephas' zone, early MN-16b [2.8] = upper Gauss [2.8]
Siwaliks, N. Pakistan (Potwar Plateau): mammal faunas with last hipparion, first *Elephas* and cervids, in base of upper normal interval of Gauss (Chron 3) (4; 6; 15)

Arcille (Baccinello V4)
Late Ruscinian, MN-15 [3.7] = late Zanclean, PL2 [3.6]
Grosseto, N. Italy: small mammals, with *Mimomys* and *Blairinoides*, in lignites lying directly over marine transgressive tongue with *G. margaritae* foram fauna (14).

Anza-Borrego
Early Blancan [4.0] = Gilbert C1 [3.9]
Imperial Valley, California: mammals in lower 'Layer Cake' zone of Anza-Borrego sequence are associated with normally magnetized sediments, identified according to a palaeomagnetic profile that extends into the Pleistocene (22).

Yepomera
Latest Hemphillian [4.8] = mid Gilbert C2 [4.6]
Chihuahua, Mexico: mammals in normally magnetized sediments overlain by reversely magnetized beds, not yet Blancan but younger than late Hemphillian Mount Eden l.f. (19).

Mount Eden
Latest Hemphillian [5.0] = earliest Gilbert [5.0]
San Bernardino, California (San Timoteo badlands): mammals in reversely magnetized beds; palaeomagnetic age constrained by older (Warren) and younger (Yepomera) latest Hemphillian local faunas of south-west N. America found in positively magnetized beds (19).

Tatrot
Base *Hexaprotodon* zone, late MN-15 [5.4] = upper Chron 5 [5.4]
Siwaliks, N. Pakistan (Potwar Plateau): mammal faunas with first *Hexaprotodon* in upper normal interval of Chron 5 (6; 15)

Upper Bone Valley
Late Hemphillian [5.5] = Zone N.17–18 [5.2]
Southern Florida: mammals in channel-filling deposits associated with end-Miocene rise in sea-level (18; 32).

La Alberca
Latest Turolian, late MN-13 [6.0] = mid N.17 [6.0]
Murcia Province, Spain (Fortuna Basin): small mammals, more advanced than late Turolian Crevillente 6, but not yet Ruscinian, in beds intercalated with 'Miocene Terminal' containing *G. conomiozea — G. mediterranea — G. multiloba* foram fauna (9)

Crevillente 6
Late Turolian, mid MN-13 [6.5] = mid N.17 [6.0]
Alicante Province, Spain (Fortuna Basin): small mammals (with earliest *Apodemus, Paraethomys*) similar to Librilla (K/Ar 6.7 Ma), in beds correlated to 'Miocene Terminal' with *G. conomiozea, G. mediterranea* foram fauna (9).

Chamita
Late Hemphillian [6.5] = mid Chron 6 [6.2]
Rio Grande Valley, New Mexico (Espanola area): mammals directly equivalent to type Hemphillian (Coffee Ranch, 6.6 Ma) in lengthy reverse-polarity zone; local fission-track ages (*c.*5.0 Ma) suggesting early Gilbert correlation are discounted (17; 29).

Late Maeotian
Late Turolian, early MN-13 [6.8] = Mollusc zone NSM-6 [7.0]
Ukraine region, USSR: mammal faunas (Tudorovo, Avgustovka) with earliest *Protoryx* correlated to mollusc faunas associated with mid Chron 7 (11; 27), in present notation.

Dhok Pathan
Base 'Selenoportax' zone, mid MN-12 [7.5] = lower Chron 7 [7.4]
Siwaliks, N. Pakistan (Potwar Plateau): mammal faunas with last *Deinotherium* and first large giraffids, just above Chron 8 (6; 15), in present notation.

Sidi Salem
Early Turolian, MN-11/12 [8.5] = upper N.16 [8.0]
N. Algeria, southern Oran district (Beni-Chougrane massif): small mammals at top of Bou Hanifia Formation, unconformably overlain by lower N.17 marine strata (23)

Crevillente 1–3
Earliest Turolian, MN-11 [9.0] = upper N.16 [8.0]
Alicante Province, Spain (Fortuna Basin): small mammals, with earliest *Parapodemus*, in beds overlying marine sediments with *G. acostaensis — G. psuedomiocenica* foram fauna (9)

North Tejon Hills
Late Clarendonian [9.5] = late early Mohnian, CN7/8 [8.5]
S. San Joaquin Valley, California: mammals in shallow marine Chanac Formation, correlative to basal Etchegoin Formation (3; 24; 26; 30).

Nagri
Base 'Hipparion' zone [9.5] = lower Chron C5N1 [10.0]
Siwaliks, N. Pakistan (Potwar Plateau): mammal faunas with first Indo-Pakistan area hipparion (F/T 9.5 Ma) located 1/3 up from bottom of 'long normal' (Anomaly 5 correlative) polarity zone (6; 15).

Kastellios
Late Vallesian, early MN-10 [10.5] = earliest N.16 [10.0]
Island of Crete: mammals in mid to late Vallesian sites (with primitive hipparions, *Progonomys* spp.) directly overlain by marine strata with *G. acostaensis — Neogl. falconarae — G. apertura — G. ventriosa* foram fauna (8; 10).

South Tejon Hills
Early-mid Clarendonian [10.5] = early Mohnian, CN6/7 [9.8]
S. San Joaquin Valley, California: mammals in non-marine upper Santa Margarita Formation, younger than 'Delmontian'

Belridge Diatomite and older than mid-Mohnian Etchegoin Formation (1; 3; 26; 30).

Sycamore Creek
Early Clarendonian [11.5] = early Mohnian, CN6/7 [9.8]
San Francisco Bay, California (Berkeley Hills area): mammals in Cierbo Sandstone Member of San Pablo Formation, correlative to 'Delmontian' upper Monterey Formation in type area (21; 24; 26;).

Khersonian
Late Vallesian, MN-10 [10.5] = Mollusc zone NSM-4 [11.0]
SE Ukraine — Caucasus region, USSR: mammal faunas (Eldar, Berislava) with *Ictitherium*, *Samotherium*, *Miotragocerus*, and '*primigenium*'-type hipparion (older than early Maeotian K/Ar 10.0–10.6 Ma), correlated to molluscs associated with lower Chron C5R, in present notation (2; 11;)

Bou Hanifia
Mid-Vallesian, MN-9/10 [11.0] = basal N.16 [10.5]
N. Algeria, Beni-Chougrane massif: mammals (with *Hipparion primigenium*, *Samotherium*, *Progonomys cathalai*, *Zramys*) in beds intertonguing with marine strata containing *G. continuosa* fauna, with tuff (K/Ar 39/40 age of 12.1 Ma) at base; overlies N.14 Anaseur Formation (23)

Comanche point
Early Clarendonian [11.5] = early Mohnian, CN5/6 [10.8]
S. San Joaquin Valley, California (Tejon Hills area): mammals in shallow marine lower Santa Margarita Formation, correlative to 'Delmontian' Belridge Diatomite (1; 21; 26; 30).

Bessarabian
Early Vallesian, MN-9 [12.0] = Mollusc zone NSM-2b [12.5]
Black Sea basin: Mammal faunas (Jeltokamenka) with primitive hipparion and *Anchitherium*, correlated to mollusc faunas associated with upper half of Chron C5A (2; 11), in present notation.

Beglia
Base Vallesian, MN-8/9 [12.5] = Zone N.12 [12.0]
Metlaoui district, central Tunisia (Bled Douarah basin): *Hipparion* datum, in beds that grade northward to late Serravallian lower Saouaf Formation (7; 33)

Volhynian
Latest Astaracian, MN-8 [13.0] = Mollusc zone NSM-1 [13.2]
Azov basin, SE Ukraine: mammal faunas (Kryvoi-Rog) with *Anchitherium*, *Euprox*, and *Aceratherium incisivum* correlated to mollusc faunas associated with Chron C5AB (2; 11), in present notation.

Pojoaque
Late Barstovian [13.5] = Chron C5AA [13.0]
Rio Grande Valley, New Mexico (Espanola area): mammals in normal-polarity zone in upper Santa Cruz sites, upper Tesuque Formation, correlated to GPTS in dated palaeomagnetic profile (5; 29).

Sharktooth Hill
Mid Barstovian [14.0] = late Luisian, lower CN5a [14.0]
SE San Joaquin Valley, California (Kern River area): mammals, mostly marine, in upper Round Mountain Silt; other mid-Barstovian localities in SW San Joaquin Valley, where the upper Caliente Formation intertongues with Luisian White Rock Bluff Member of the Branch Canyon Formation (3; 24; 26; 30)

Skull Ridge
Early Barstovian [15.0] = Chron C5AD [14.5]
Rio Grande Valley, New Mexico (Espanola area): mammals in normal-polarity zone in middle Tesuque Formation, below Santa Cruz sites, correlated to GPTS in dated palaeomagnetic profile (5; 29).

Burkeville
Early Barstovian [15.0] = Zone N8/N9 [15.3]
Texas Gulf Coast: mammals sandwiched between marine transgressive tongues in upper Fleming Formation, '*Potamides matsoni* Zone' (13; 30).

Neudorf-Sandberg
Early Astaracian, MN-6 [15.0] = mid CN5a [14.0]
CN5a [14.0]
Devinska Nova Ves, CSSR (lower Morava valley): mammals (including dryopthecines, *Conohyus*, and last *Listriodon*) in marine sands of the lower *Bulimina-Bolivina* zone (K/Ar 15.0 Ma) (16; 20; 25; 28; 31;)

Kleinhadersdorf
Early Astaracian, MN-6 [15.5] = basal CN5a [14.5]
CN5a [15.0]
Poysdorf district, N. Austria: mammals (including first dryopithecines, *Conohyus*, *Gazella*) in marine sands of the upper Lagenid zone or basal *Spiroplectammina* zone (16; 25; 28; 31)

Neudorf-spallte
Late Orleanian, MN-5 [16.0] = early Badenian, mid CN4 [15.5]
Devinska Nova Ves, CSSR (lower Morava valley): mammals in fissure deposits (with pliopithecines, *Dicroceros*, *Palaeomeryx*, *Taucanomo*) overlain by transgressive marine beds of lower Lagenid zone (K/Ar 16.5 Ma) (12; 16; 20; 25; 31)

Barker's Ranch
Latest Hemingfordian [17.0] = early Relizian, CN3 [17.5]
S. San Joaquin Valley, California (Kern River area): marine and terrestrial mammals in Olcese Sand, intertonguing with base of Round Mountain Silt. In SW San Joaquin Valley, other late Hemingfordian localities occur in the upper Caliente Formation, which intertongues with the lower Relizian Saltos Shale Member of the Branch Canyon Formation (3; 21; 24; 26; 30).

Midway
Late Hemingfordian [17.5] = Zone N6–N7 [17.6]
Florida panhandle: mammals in upper Torreya Formation, which intertongues with basal Chipola Formation (13; 18; 30).

Seaboard
Mid Hemingfordian [18.5] = mid N.6 [18.5]
Florida panhandle: mammals in middle Torreya Formation: also in central Florida, where mid-Hemingfordian Thomas Farm mammals occur at the same level in lower Hawthorne Formation (13; 18; 30).

Cuyama Pato Red
Latest Arikareean [21.0] = Mid Saucesian, CN1c [20.5]
Transverse Ranges, California (Cuyama Valley area): mammal localities in Pato Red Sandstone, which intertongues

with 'Upper Vaqueros' Painted Rock Sandstone. (3; 26).

Pyramid Hill

Late Arikareean [22.0] = early Saucesian, CN1b/1c [22.5]

S. San Joaquin Valley, California (Kern River area): mammals (with *Anchitherium*) at base of Jewett Sand; just above Tecuja Dacite (K/Ar 22.3 Ma). In SW San Joaquin Valley, other late Arikareean mammal localities occur in the Lower Caliente Formation, which intertongues with earliest Saucesian Agua Sandstone of Santos Shale Formation (3; 21; 26).

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An interim time-scale

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SUMMARY: The time-scale presented here endeavours to take cognisance of the views expressed in the Symposium, those expressed in other recent publications, notably the reviews and collations of Odin (1982) and Harland *et al.* (1982; based on Armstrong 1978), and the suggestions of the Subcommission on Precambrian Stratigraphy (Sims 1980). The conclusions of the contributors to these various works are not always compatible and this scale involves elements of both compromise and personal judgement. No plus or minus figures have been assigned to the boundaries although estimates of the uncertainty to be attached to the boundaries are commonly to be found in the more recent literature. It should be emphasized that these uncertainties are generally expressions of the limits within which a particular boundary probably lies *according to the author concerned*, they are thus apt to be somewhat subjective.

The Precambrian

Cowie & Johnson (this volume) discuss the late Proterozoic time-scale and compare the chronometric scale proposed by the Subcommission on Precambrian Stratigraphy in 1979 (Simms, op. cit.) with the chronostratigraphic scale proposed by Harland *et al.* (op. cit.) 'designed to satisfy the aspirations of the Australian, Chinese, Russian and Scandanavian communities ...'. The former is virtually the same as the proposed for the USA and Mexico (Harrison & Peterman 1980), it is essentially lithostratigraphic in that the boundaries were selected so as to 'split as few major rock sequences as possible'. Other lithostratigraphic scales can be drawn up for the other continents with significantly different boundaries but which are equally relevant and functional for the regions concerned — compare for example the scales for the Canadian Shield, Africa, and the USA and Mexico (Snelling, this volume); thus, despite the advocacy of the Subcommission on Precambrian Stratigraphy there would seem to be no reasons for adopting their proposed chronometric scale in preference to any of the others.

The latter chronostratigraphic scale is a clear reflection of the world-wide effort that has been put into the application of the principles used to define Phanerozoic chronostratigraphic divisions to the supracrustal rocks of the Precambrian; including the vigourous persual of stromatolite and microphytolite biostratigraphy, and the correlation potential of glaciogenic deposits, palaeomagnetism, and the conventional methods of geochronology (Trompette 1982). The proposal to use the name Sinian for the youngest era or sub-era of the Precambrian, however, could introduce a considerable degree of confusion. As used by Chen Jinbiao *et al.* (1981) the 'Sinian Suberatherm' extends from c.1950 Ma to the lower boundary of the Palaeozoic, while the 'Sinian System' is the youngest chronostratigraphic unit in the Sinian Suberatherm. Even if this problem of nomenclature can be resolved a restructuring of the late Precambrian chronostratigraphy would be called for, specifically 'a radical restriction of R_3' — the Karatavian (Harland *et al.* 1982). Despite some confusion in the classification of the late Riphean, the 'Riphean' is clearly regarded by Soviet geologists as an entity as emphasized by their two-fold division of the Proterozoic into Lower (unnamed) and Upper (Riphean) units only. The proposal of the Comité français de Stratigraphie (1980) to assign the Upper Riphean (R_3 and R_4) and the Vendian to the Upper Proterozoic, with the Middle and Lower Riphean (R_1 and R_2) constituting the Middle Proterozoic, would allow such a restructuring of the late Proterozoic chronostratigraphy, although it might not initially find favour with Soviet specialists. These are taxing but not insoluble problems, but their resolution can only be achieved through international co-operation and co-ordination.

The Lower Proterozoic represents a long period of time, tbe subdivision of which is only briefly touched upon by Harland *et al.* (1982). An important division can clearly be made at c.2000 Ma which is marked by the appearance of the oldest fossils known to display a clear differentiation into two or more types of cells (the Gunflint microbiota; Cloud 1983), the near limitation of the banded iron formations to rocks older than about 2000 Ma (98%; Goodwin 1982) and the appearance of the oldest conspicuous 'red beds' at about this time. Irrespective of arguements concerning the relation between these phenomena and the actual oxygen level in the atmosphere, these observations seem sufficient to justify a division at about 2000 Ma. The period of time from c.2000 Ma to c.1650 Ma could then constitute an era or sub-era for which I would suggest the name 'Guianian' reflecting the spectacular development of the Roraima 'red beds' Supergroup in the Guiana Highlands of South America, deposited in this time interval.

The foregoing discussion leads again to consideration of the rejection of the etymologically unsatisfactory Proterozoic and the substitution of Proterophytic and Palaeophytic, both of eon status, the former extending from the 'conventional' end of the Archaean at 2500 Ma to c.2000 Ma, and the latter from 2000 Ma to the Precambrian: Cambrian boundary. It would seem not impossible to define the Proterophytic: Palaeophytic boundary following chronostratigraphic principles. The Palaeophytic Eon would thus consist of three Eras, viz: the Guianian from c.2000 Ma to c.1650 Ma; the Riphean from 1650 Ma to ?c.680 Ma; and the Vendian (± the Kudashian, R_4) to the Precambrian-Cambrian boundary. Restructuring of the late Palaeophytic could be expected with advances in stromatolite biostratigraphy (Bertrand-Sarfati & Walter 1981) and further research on Ediacaran faunas.

Consideration of the recent literature suggests that there is still considerable scope for the International Commission on Stratigraphy to co-ordinate and rationalize work on the Precambrian time-scale and the prospect that a functional chronostratigraphic scale will be developed seems to be very real. Precambrian chronstratigraphy is developing rapidly particularly under the auspices of the IGCP Projects 99 and 118 (see Precambrian Research, **15**, 1981 and **18**, 1982; Cowie & Johnson, this volume) suggesting that the international adoption of the chronometric scale proposed by the Subcommission of Precambrian Stratigraphy in 1979 might be premature.

The Precambrian: Cambrian boundary and the Palaeozoic and Mesozoic time-scale

The continuing resolution of biostratigraphic problems within the Cambrian System, including a significant measure of agreement on the definition of the base of the Cambrian System at the Precambrian–Cambrian boundary has not been accompanied by comparable improvements in our understanding of the chronology of the Cambrian Period. Indeed Gale (1982) considered it necessary to reject much of the earlier geochromometric data and from consideration of a few (in his view reliable) age determinations suggested an age for the base of the Cambrian (the Precambrian–Cambrian boundary) of 530 ± 10 Ma, an interpretation which was further amplified by Odin *et al.* 1983. This view has been slightly modified following the demonstration that the critical Caer Caradoc granophyre intrudes 'lower' Cambrian sediments, although its relationship to the lowest fossiliferous Cambrian sediments is not so certain. In their review of the current situation Odin *et al.* (this volume) suggests an age for the Precambrian–Cambrian boundary not much older than 540 Ma. Cowie & Johnson (this volume) would regard much of the data considered by Odin *et al.* (this volume) as ambiguous or unreliable and tend to stress Chinese data, particularly a uranium-lead isochron determination on Atdabanian black shales which indicates an age of *c.*570 Ma. So far, however, the Chinese data have received inadequate documentation and remain difficult to assess; Cowie & Johnson are unable to come to any firm conclusion regarding the age of the Precambrian–Cambrian boundary but leave open the possibility that it may be appreciable older (+ 570 Ma) than the younger age favoured by Odin *et al.*

The writer's view of the Cambrian time-scale is conditioned by a willingness to accept: (1) that the Vire-Carolles granite appears to be pre Atdabanian (relation to Tommotian is less certain) and that the most reliable age that can be extracted from the plethora of discordant data from this intrusion is 540 ± 10 Ma (Pasteels & Doré 1982); (2) that the Caer Caradoc granophyre — 533 Ma — possibly intrudes trilobite bearing Atdabanian sediments (Cowie & Johnson, this volume). None of the other data discussed by Cowie & Johnson, and Odin *et al.* seem to be sufficiently unambiguous to throw any more helpful light on the problem. This estimate for the age range of the Atdabanian is of course at variance with the age of *c.*570 Ma determined on uraniferous sediments from China of Atdabanian age (this seems to be the least ambiguous of the considerable body of Chinese data). As far as the Precambrian–Cambrian boundary itself is concerned a reviewer can do little more than follow Lambert (1971) and accept the late Precambrian Holyrood granite as being 'significant'. Gale (1982) has recalculated the age of this intrusion to 585 ± 15 Ma. McCartney *et al.* (1966) suggested that an arbitrary 15 Ma may have elapsed between the emplacement of this granite and the deposition of overlying Early Cambrian (possible pre-Atdabanian) sediments. Tommotian-type (s.l.) fossil assemblages are present in Newfoundland, and one would not seem to be doing too much violence to the generally unsatisfactory evidence to retain as an interim figure an age of 570 Ma for the

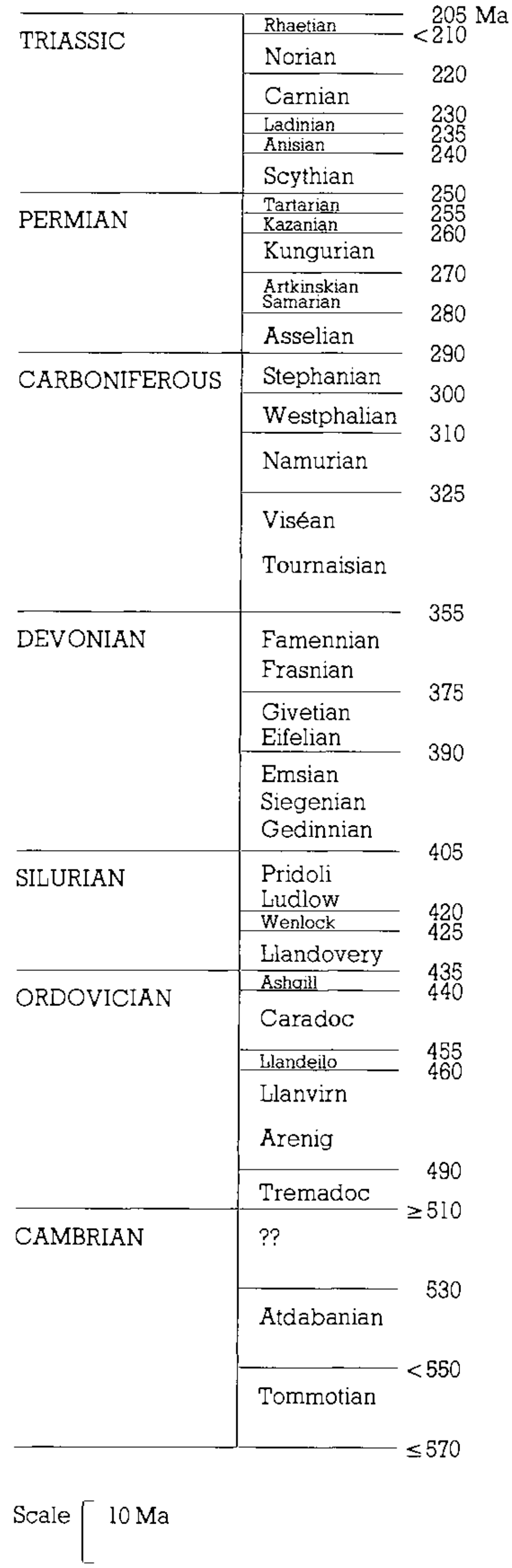

FIG. 1. The Phanerozoic time-scale: Cambrian to Triassic.

			65 Ma
CRETACEOUS	Late	Maastrichtian	
			72
		Campanian	
			83
		Santonian	
			86
		Coniacian	
			88
		Turonian	
			91
		Cenomanian	
			95
	Early	Albian	
			107
		Aptian	
			114
		Barremian	
			116
		Hauterivian	
			120
		Valanginian	
			128
		Berriasian	
			135
JURASSIC	Late	Tithonian	
			139
		Kimmeridgian	
			144
		Oxfordian	
			152
	Middle	Callovian	
			159
		Bathonian	
			170
		Bajocian	
			176
		Aalenian	
			180
	Early	Toarcian	
			188
		Pliensbachian	
			195
		Sinemurian	
			201
		Hettangian	
			205

Scale 10 Ma

FIG. 2. The Phanerozoic time-scale: Jurassic and Cretaceous.

Precambrian–Cambrian boundary.

There are no data relevant to the younger limit to the Cambrian. A figure 510 Ma is adopted here based on the extrapolations to the base of the Tremadoc by McKerrow *et al.* and Gale (this volume). The remainder of the Palaeozoic time-scale (Fig. 1) generally follows the views expressed by the various contributors to this volume; the ages having been rounded off to the nearest five million years. This rounding-off seems justifiable in view of the debatable methods of extrapolation used by McKerrow *et al.* and Gale. It is only in the late Silurian and early Devonian that there are significant differences among the various contributors which are largely due to inclusion by Gale of the Stockdale Rhyolite data. The data presented in this volume on the Esquibel Island hornblende and Hopedale biotite (see Kunk *et al.*, this volume), and the results of Wyborn *et al.* (1982) on the Laidlaw Volcanics all suggest that the Stockdale Rhyolite age is anomalous; it is here excluded from consideration. An age of 405 Ma has been adopted for the Silurian: Devonian boundary as a compromise between the ages determined by McKerrow *et al.* (this volume, 412 ± 5 Ma) by Gale, 1982 (400 ± $^{10}_{5}$ Ma), and by Harland *et al.* 1982 (408 Ma).

The Upper Palaeozoic and Triassic time-scales of Foster & Warrington, and Odin (this volume), shown an encouraging convergence and the scale by Forster & Warrington, slightly modified in accordance with some of Odin's comments is adopted here (Fig. 1). The Jurassic and Cretaceous time-scale of Hallam *et al.* (this volume) is largely based on that by Kennedy & Odin (1982). The most significant divergence in scales is at the geochronologically poorly defined Jurassic–Cretaceous boundary for which Hallam *et al.* prefer an age of 135 Ma in contrast to 130 Ma adopted by Kennedy & Odin (op. cit.) and 144 Ma adopted by Harland *et al.* (1982). One new datum on sanidine from an Upper Aptain fuller's earth (Jeans *et al.* 1982) has prompted Hallam *et al.* to adopt an age of 114 Ma for the base of this stage. Kennedy & Odin (op. cit.) had suggested that the Barremian-Aptian boundary lay in the interval 110–114 Ma with the higher figure most likely. The prefered scale in Fig. 2 attempts a compromise between the aforementioned alternatives. However, greater weight has been given in most cases to the Kennedy & Odin scale for which the assessment of the time-scale utility of the analysed samples is exhaustively documented.

The Cenozoic

Although there would now seem to be a reasonable convergence of opinion concerning most of the Palaeozoic and Mesozoic time-scale there is a strong divergence of opinion for the Paleogene where no compromise would seem at present to be acceptable. The problems of Paleogene correlations have been effectively summarized by Hardenbol & Berggren (1978) and Harland *et al.* (1982). The various proposed stratotypes are scattered among many small outcrops in basins all over Western Europe where the intercalation of marine, marginal marine and non-marine sediments together with lateral facies changes 'all combine to render inter-basin correlation difficult, and the possibility of world-wide correlation even more remote'. (Harland *et al.* 1982, p. 34). However, the development of biostratigraphic zonations based on planktonic foraminifera, calcareous nannoplankton, and radiolaria is generating a biostratigraphic framework into which the European Paleogene stratotypes may be placed.

The time-scale presented by Berggren *et al.* (this volume) is a comprehensive integration of the aforementioned biostratigraphic zonations with ocean floor spreading (paced by ocean crust magnetic lineations), the rate of which is calibrated by age determinations on terrestial volcanics. The rationale here is that terrestial volcanics will give more reliable ages compared with glauconies which unless carefully selected can yield K:Ar ages which may be too low or too high. In generating the time-scale, however, assumptions must be made about the rate of sea-floor spreading — it is assumed to be more or less constant, but does it vary significantly with time? (see Curry, this volume); foraminifera, nannoplankton and radiolaria zones must be correlated with terrestrial faunas associated with the dated volcanics; and the polarity signature of the dated volcanics must be related to the ocean floor magnetic lineaments.

The alternative approach is to undertake age determinations on glauconies which are common in the Paleogene basin sediments of western Europe, to supplement this data where possible by dating intercalated volcanics, and to integrate such data with the stratigraphy in the conventional way. The principle difficulties of this approach lie in the problems of inter-basin and worldwide correlation, and the heavy, thought not exclusive, reliance on glauconies as the dated mineral (the suitability of this mineral for dating must be assessed with great care). Such a time-scale for the Paleogene has been established by Curry & Odin (1982).

The Paleogene time-scale derived by Berggren *et al.* (this volume) is summarized in Fig. 3 and the Paleogene scale determined by Curry & Odin (1982; the scale is fully documented in this reference) is given for comparison. That such different scales have emerged is a clear indication of the difficulty of the task and of the need for further research. The discrepancy may in part reflect significant variations in the rate of sea-floor spreading as emphasized by Curry (this volume), and both approaches may be subject to uncertainies in stratigraphic correlations, and distortions due to the vagaries of the dated rocks and minerals. An interim scale cannot yet be drawn up for the Paleogene, and it must be left to interested individuals to make their own assessments of the voluminous documentation which back up both scales. For the Neogene the scale proposed by Berggren and his co-workers in Jenkins (this volume) is here adopted, viz: base of Miocene 23.7 Ma, base of Pliocene 5.3 Ma, and the Pliocene–Pleistocene boundary at 1.6 Ma. It should be noted, however, that Curry & Odin (1982) would date the base of the Miocene at 23 Ma and that the definition and hence the age of the base of the Pleistocene remains the subject of debate (Bowen, *In*: Jenkins, this volume).

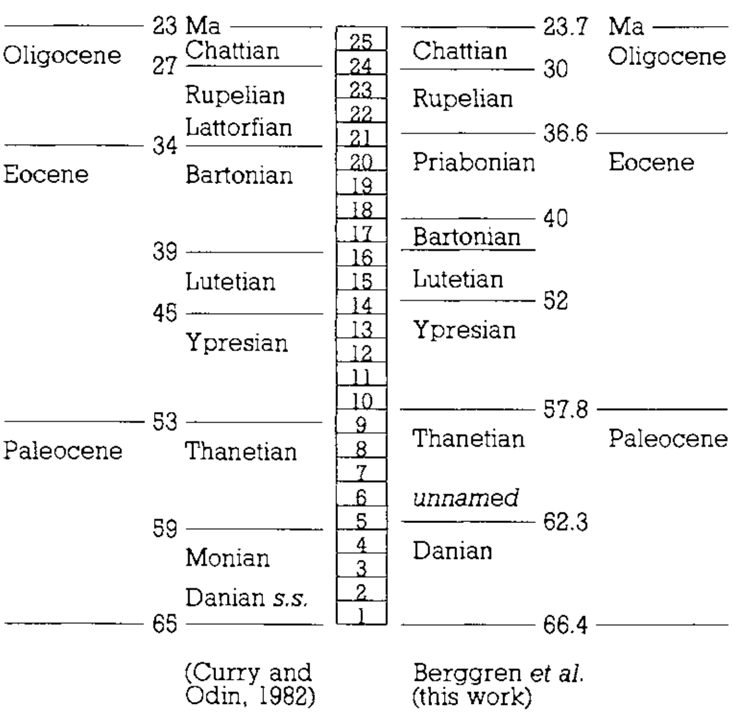

FIG. 3. Paleogene time-scales related to calcareous nannoplankton.

Concluding comments

The Phanerozoic time-scale suggested here differs remarkably little from that proposed following the Phanerozic Time-Scale Symposium of 1964. However, the 1964 scale gave time boundaries which were often dependent on a general agreement of individually unreliable sets of figures. Many of the 1964 data have now been discarded and much of the time-scale suggested here is based on relatively new and more reliable analytical data, though often the stratigraphic constraints are still poor. Except for the Cambrian where there is a dearth of universally acceptable data, there is an encouraging measure of agreement among the various contributors about the Palaeozoic time-scale, an agreement which also extends upwards through the Triassic. Indeed the time-scale for the Permian and Triassic presented here is a major improvement on any previous scale. The Mesozoic time-scale is similarly much improved mainly because of our better understanding of glaucony as a geochronometer, and the thorough assessment of analysed samples from both the geochemical and stratigraphic points of view by Odin and his collaborators (Odin 1982). However, the resolution of stages within the Jurassic and Cretaceous remains very poor and the lingering element of subjectivity in the assessment of glaucony ages inevitably weakens confidence in the accuracy of the Mesozoic scale. Improvements in resolution can still be hoped for, at least down to the limits imposed by the typical experimental errors of *c.* ± 3% of the age. The more pressing need, however, is for assurances concerning the accuracy of the scale (see for example the discussion by Hallam 1984, concerning the problems of mass extinctions); this can best be bought about — given the slight subjectivity of glaucony ages and the 'element of vagueness associated' with dated plutons (Lambert 1981, p.17) — by the search for and successful dating of stratigraphically constrained volcanic horizons.

Although an impressive potential for resolution has been demonstrated for the Cenozoic the question of accuracy is here particularly acute. That the scales of Berggren *et al.* (this volume) and Kennedy & Odin (1982) can be more or less concordant at the beginning and the end of the Paleogene but differ by as much as 10% in the Eocene, speaks of systemmatic error(s) in one or both scales which have yet to be recognized.

The definitive time-scale still remains elusive, however, the resolution of the problem of the half-life of rubidium-87 has resulted in the wider application of the Rb-Sr method on whole rock systems, as appealed for by Lambert (1971), with significant improvements in the Palaeozoic and early

Mesozoic parts of the time-scale. Our better understanding of the glaucony geochronometer has equally led to improvements in the Mesozoic and Cenozoic time-scales. While there still remains room for improvements in analytical methods, particularly perhaps in the dating of accessory minerals such as zircons from younger intrusives and volcanics by the U:Pb method, and the wider applications of the Sm:Nd method, it remains appropriate to echo Lambert's view that 'the present situation regarding the development of the scale is clear enough: there is a shortage of raw material suitable for its further refinement ____' It is encouraging that exploration, development and reassessment in the earth sciences continues to produce such raw materials, and that the hope 'that this co-operative enterprise will continue' as expressed by Harland in the Preface to the *Geological Society 1964 Phanerozoic Time-scale*, continues to be fulfilled.

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Oceanic magnetic lineaments and the calibration of the late Mesozoic – Cenozoic time-scale.

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SUMMARY: The pattern of magnetic reversals from mid-Cretaceous times onwards is now well-known from the study of ocean-floor spreading systems. Integration of the pattern of polarity reversals with the standard biostratigraphical time-scale is now well-advanced and, as a result, a series of recently proposed time-scales use the detailed pattern of magnetic lineaments, calibrated radiometrically at a few points only, as their major control. The anomaly sequence generally accepted as standard is based on a single track across the South Atlantic and its representativeness of at least the greater part of the Cenozoic has never been seriously challenged. Assumption of a constant spreading rate with time has enabled detailed and apparently very accurate time-scales to be generated. However, spacings within anomaly sequences in the same oceanic area vary notably from track to track and oceanic areas themselves may exhibit characteristic patterns. Clearly, in detail, the rate of ocean-floor spreading with time is not constant, and the role of patterns of magnetic reversals in the construction of time-scales needs reevaluation in the light of this fact. Keywords: time-scales, oceanic magnetic lineaments, ocean-floor spreading.

Many writers have used the spacing of ocean magnetic stripes for the subdivision of time, but the first application to a long time-span was that of Heirtzler *et al.* (1968). These authors chose a single anomaly sequence from the South Atlantic (Vema-20) on an E-W line approximately at 28°S latitude. From this they constructed a time-scale (HDHPL68) by extrapolation from a radiometric date of less than 4 Ma to approximately 80 Ma. In spite of the extreme degree of extrapolation involved, this scale has proved to be astonishingly accurate in its predictions. The majority of subsequent scales for that period of time have been based on the Vema anomaly sequence, modified to take account of new biostratigraphical correlations and preferred spot radiometric ages. The fundamental assumption of HDHPL68 was, of course, that sea-floor spreading proceeds at a steady rate. However, the fact that substantially all subsequent magnetically-calibrated time-scales are based on HDHPL68 means that if any changes in velocity are represented within it they have become an integral and perhaps unrecognized part of the structure of scales in general. One result of this situation has been that where discordant radiometric data have been encountered, those more in accord with HDHPL68 have been selected preferentially, an action tending to disguise deviations from the original constant-spreading hypothesis.

There is, of course, nothing unique about the spreading regime in the South Atlantic. It just happened that, at the time they wrote, Heirtzler *et al.* thought *it* to be the most reliable one available. They were quite aware from comparisons between spreading systems in the Indian, Pacific and South Atlantic oceans that spreading rates varied both between and within systems. Their observations were relative only, however, and they had no means of calculating absolute values of speading in any particular case. One may therefore conjecture what time-scale might have emerged if a different spreading system had been chosen by Heirtzler *et al.*, and what alternatively would result if the patterns of all known systems were given equal weight, to explore a hypothesis of mean constant-rate spreading world-wide. This situation is examined in the paragraphs which follow.

The analysis is a preliminary one, based on maps in Owen (1983), which provide six azimuthal equidistant projections of the earth's surface at the present day, marked with the ridge-systems and associated numbered anomalies. Areas have been identified where long, continuous successions of anomalies are present and the distances between the originating ridge and selected anomalies have been measured. Those chosen, because most commonly plotted, are 5, 13, 21, 25, 31 and 34 (young). In some cases, in order to complete a sequence, the position of an anomaly has been estimated by interpolation or a small extrapolation. Ideally the distances should be measured on the surface of a globe, but measurements from the maps in Owen are believed to be sufficiently accurate for the present purpose. Examination of the maps reveals that there are comparatively few regions along the world's ridge systems where long series of lineaments extending to the present day are developed, quite the most important of these being along the Atlantic ridge. Six sequences (at about 15, 20, 28, 35, 40 and 45 degrees south) were measured in the South Atlantic, and seven (at about 15, 20(2), 25(2) and 45(2) degrees north) in the North Atlantic.

In the Pacific, four sequences were measured SE of New Zealand and two west of North America. Measurements made to the south of Australia and along the Carlsberg and Reykjanes ridges were not included in the analysis which follows because their anomaly sequence was insufficiently long, and a long sequence to the SE of Madagascar was discarded because of evidence of very rapid spreading in the last 40 Ma. Measured distances between the chosen anomalies were compared and relative ages were calculated, using a constant-spreading hypothesis for each set. To express the results in years it is necessary to assume one numerical date; this was chosen to be 84 Ma at anomaly 34 (young), to agree with that accepted in the time-scales of Lowrie & Alvarez (1981) and Berggren *et al.* this volume, here referred to as LA81 and BKF85 respectively. Results are summarized in Table 1. Replication established that measurements could be made to an accuracy of the order of ± 0.1 mm, corresponding to a possible error in time of less than 0.3 Ma in the Pacific and South Atlantic, but as much as 0.6 Ma in the northern part of the North Atlantic, where spreading rates are much lower. Comparison with the standard errors actually observed (see Table 1) indicates that the elimination of errors due to measurement would not significantly change the arguments advanced herein.

Very few regions in the world exhibit anomaly sets extending as far back as anomaly 34, so to widen the scope of

TABLE 1. Dating of oceanic magnetic anomalies.

	Analyses based on sea-floor spreading (s-f-s)						radio metric (r-m)	s-f-s + r-m
anomaly	HDHPL68 reset to 84Ma	South Atlantic mean of 6	North Atlantic mean of 7	NE Pacific mean of 2	South Pacific mean of 4	weighted mean cols 3–6	CO82	NLC80
5 (old)	9.4	9.9 ± 0.6	9.2 ± 0.7	9.4 ± 1.0	13.2 ± 0.7	10.3	10.0	10.3
13	36.3	33.9 ± 1.4	29.1 ± 2.1	32.5 ± 0.5	33.2 ± 2.4	31.8	33.5	34.8
21	50.9	46.6 ± 3.5	41.2 ± 3.0	45.4 ± 0.4	42.4 ± 2.0	43.6	43.5	48.5
25	60.3	°55.5	53.5 ± 2.6	54.2 ± 0.3	50.6 ± 2.6	53.6	55.2	58.1
31	68.6	62.1 ± 1.7	64.0 ± 1.4	*63.0	*63.0	63.1 (cols 3–4)	66.5	69.5
34 (young)	°*84.0	*84.0	*84.0				84.0	85.9

°Position of anomaly estimated. *Assumed date.
Dates refer to mid-point of anomaly unless stated otherwise.

the present analysis an additional datum point was established at anomaly 31, as follows. As Table 1 demonstrates, the dates derived for anomaly 31 for the North and South Atlantic agree rather well, so their mean was taken at 63 Ma to provide a basis for the calculation of the dates in the Pacific. From the total of 19 tracks studied, weighted mean ages were calculated for the selected anomalies and these are compared in Table 1 with those for the individual spreading systems, with that yielded by Vema-20 (from which HDHPL68 was compiled) and with two recently proposed but discrepant time-scales, NLC80 (Ness *et al.* 1980) and C082 (Channell 1982). NLC80 is one of a series (LKC77 (LaBrecque *et al.* 1977), NLC80, LA81 and BKF85) based in detail on HDHPL68, but each adjusted differently to fit a small number of tie-points thought by the compilers to have good radiometric control. CO82, on the other hand, is based entirely on radiometric dates, with no input in its construction from palaeomagnetism. An alternative presentation of the data is made in Fig. 1, which enables changes in relative spreading rate to be visualized more easily.

Discussion and interpretation

Table 1 demonstrates considerable discrepancies between the calculated dates for the four spreading areas; in particular in the time-span 50–30 Ma. There is clear evidence for fast spreading in the last 10 Ma or so in the four South Pacific tracks documented and this was noted in other tracks from that general area. During that period North and South Atlantic and NE Pacific tracks are approximately concordant, as are those before about 60 Ma in the North and South Atlantic, as already mentioned. In the intervening period, however, there are major differences between North and South Atlantic. It was suspected that the grouping for calculation purposes into north and south might conceal a single trend spread over the whole distance from 45°N to 45°S so the figures from individual tracks were examined in a latitude context. Both systems are internally discrepant as the error figures quoted indicate, but no evidence for a cline from north to south was observed. It seems, therefore, that the North and South Atlantic systems moved independently. Translated into spreading velocities the contrast between north and south implies relative deceleration in the South Atlantic at around 60 Ma, followed by relative acceleration at about 40 Ma. The figures for the NE Pacific resemble those for the South Atlantic, whilst those for the South Pacific differ markedly from those of the NE Pacific but have some features in common with those for the North Atlantic.

The dates for both HDHPL68 and NLC80 almost throughout the scale are notably higher than those provided by the 19 tracks under study. Figure 1, however, demonstrates that the shape of HDHPL68 agrees closely with that of the mean for the South Atlantic; in other words, that a change in the slope of HDHPL68 would bring columns 2 and 3 of Table 1 into close agreement. The slope of the plots is of course controlled by the calibrating date of 84 Ma and by the observed or estimated position of anomaly 34. Track Vema-20 terminates somewhere in anomaly 32 and it seems possible therefore that the anomalously high dates calculated for HDHPL68 by later authors relying on it result from an incorrect estimate of the position of anomaly 34 in relation to the charted sequence.

NLC80, as might be expected, agrees rather closely with HDHPL68, as does LA81. The radiometric dates by which these scales are calibrated are essentially those preferred by Berggren *et al.* (1978) and are derived almost exclusively from high-temperature rocks. LKC77 also agrees with HDHPL68, in spite of the fact that it is calibrated to 79.5 Ma at anomaly 34 (young).

Time-scale CO82 is quite differently constructed. It was compiled from radiometric data which were linked with the associated biostratigraphy. For the most part it was based on glauconies, supplemented by some data from high-temperature rocks. The glauconies typically have good biostratigraphic control; nevertheless the dates which this particular chronometer has yielded have in general been rejected by Berggren *et al.* (1978). As stated earlier, magnetostratigraphy was not used in the construction of CO82. It is interesting therefore to observe that CO82 agrees rather well with the weighted means derived in Table 1, which are derived *exclusively* (apart from the common starting date) from ocean palaeomagnetics. Thus we have the curious situation that time-scales supposedly based firmly on oceanic palaeomagnetics (LKC77, NLC80 and LA81) do *not* agree with the time-

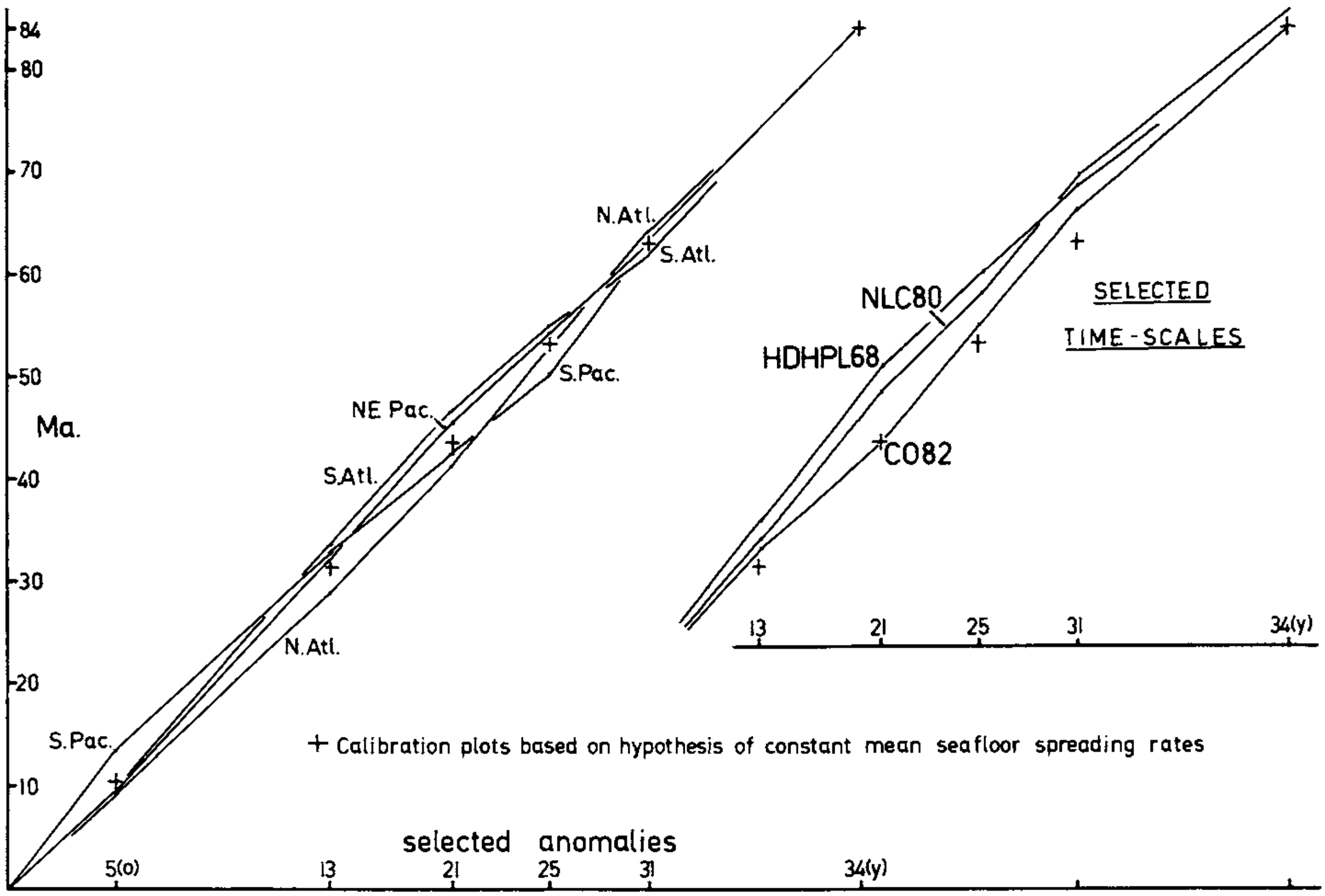

FIG. 1. Time-scales and dates derived from magnetic anomaly patterns.

scale derived from average spreading rates, whilst the time-scale (CO82) *not* based on palaeomagnetics does agree with that same spreading scale.

On the basis of the evidence so far presented the theory of a mean average sea-floor spreading rate, and consequently of the reliability of the CO82 time-scale, approximately concordant with it, might provisionally be accepted. However, problems arise when these are related to some widely-accepted dates:

1. Anomaly 34 (young), 84 Ma. This well-defined magnetic event (the end of the Cretaceous quiet period) is used as a basis for calculations in this paper and is accepted by LA81 and CO82 (NLC80 arrived at 85.9 Ma, obtained by extrapolation). The date of 84 Ma is derived from a single determination (see NDS 107; 84.4 Ma in Odin 1982) on late Santonian beds, with some supporting dates. The event has been identified in Italy a quarter of the way up the local Campanian succession. The Campanian has an estimated duration of 10 Ma so anomaly 34 (young) might therefore be as young as 82 or 81 Ma. Any reduction in the age assumed for this event would of course involve a proportional reduction in all the calculated ages in Table 1.
2. Anomaly 31. Using anomaly 34 (young) as a datum, its mid-point is dated by sea-floor spreading at 63 Ma, which implies a figure of about 61.5 Ma for the base of the Cenozoic (between anomalies 29 and 30). This latter figure is well below established radiometric dates for the boundary, which lie in the range 66.7–63.5 with a preferred mean of 65 Ma (Odin 1982).

The discrepancy between radiometric and sea-floor-spreading dates is so large that it must, it seems, be the result of a change in spreading rate affecting both halves of the Atlantic. Figure 1 indicates where slower and faster episodes appear in the comparative spreading records. Note that its indications are for the most part relative ones — there is no means in the time-spans without accepted radiometric tie-points (in particular between anomalies 13 and 31) of calculating absolute rates. The simplest explanation for the relative velocities observed is that there was a general reduction of about 30% in spreading velocities, but at different times in different systems; at about 60 Ma in the South Atlantic, about 50 Ma in the South Pacific and 40 Ma in the North Atlantic. A reversion to about the original speed occurred in the South Pacific at about 20 Ma. Nothing is deduced about a possible velocity change in the NE Pacific because of insufficient control before anomaly 31 times. The large-scale variations in relative spreading rate mentioned above are almost certainly accompanied by smaller-scale ones, as is suggested by the large error-bars derived in Table 1. Thus, not only have there been substantial changes in spreading rates in different oceans since 84 Ma but also the time of onset of such changes has been different in the different areas. In addition, a preliminary examination of earlier spreading rates in the Atlantic has suggested that relatively low rates prevailed in the time prior to the Cretaceous quiet period before a speed-up at around 100 Ma, and the subsequent slowdown at 60–40 Ma referred to earlier.

Conclusion

The search for a unifying theory relating to rates of ocean spreading has to date proved unsuccessful. There can be no doubt of the great value of ocean magnetic lineaments in providing a framework with which the necessarily incomplete land-based magnetic sequences can be correlated and of their further value in providing approximate relative and absolute ages in time-spans with little radiometric control. However, the variability in spreading rates between and within regions which has been demonstrated herein makes any attempt to use such lineaments as a main tool for the construction of a time-scale very hazardous, even undesirable. The only reliable basis for such scales is isotopic dating. It is unfortunate that that method has to date so often produced discrepant results. The answer is *not* to build time-scales based on sea-floor spreading and calibrated by the author's (perhaps idiosyncratic) choice of a few radiometric tie-points. Rather, isotope geochronologists should be encouraged to redouble their efforts to produce new dates using additional chronometers and better techniques and, in particular, to find means of resolving discrepancies between published dates. Many, many more reliable dates are needed, to ensure that the main framework of the scale is soundly based. At that time, detailed palaeomagnetic study should be able to refine the scale yet further, because its potential for fine correlation is much higher than isotope work is ever likely to achieve, or than biostratigraphy promises.

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The ammonoid time-scale and ammonoid evolution

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SUMMARY: For a period of some 330 × 10^6 years, from the early Devonian to the late Cretaceous, ammonoid zones and subzones in marine rocks provide a time *discrimination* of the order of 0.25−1.0 × 10^6 years or a *resolving power* of 1:100 to under 1:500. Several time intervals show much better results. The progressive improvement of these techniques is illustrated.

The evolutionary tree of the Ammonoidea is as well documented as for any group. The record, with its periods of extinction and diversification at several taxonomic levels can be viewed as a bio-seismograph of the factors controlling evolution. The rôle of early stage modifications in evolutionary strategy is emphasized. A relationship between environmental change and critical events in ammonoid history is examined. Examples illustrate punctuated as well as gradualist mode confirming Darwin's view on the broad range of evolutionary mechanism. The ammonoids show no evidence supporting an hypothesis of regular diversity decay. The case is made for a more collaborative and systematic study of evolutionary fluctuations as a key to documentation of primary causations which may be global or cosmic.

The evolution of the Ammonoidea has been the object of study for over a century. It provides a powerful tool for the analysis of geological events through time. This paper attempts a brief review of the patterns of ammonoid evolution and their contribution to evolutionary theory. This will be preceded by comments on the ammonoid time-scale and precision in dating past events. Emphasis is placed on how the evolutionary pattern itself provides a delicate and sensitive record of changes in the environment at all scales and how this gives a guide to the factors influencing evolution itself. These factors are likely to be the end member of a series of causations the primary cause being global or cosmic physical changes and hence they serve to document these with precision.

A current view on the evolution within the Ammonoidea is illustrated in Fig. 1. This is compiled from a range of data assembled by House, Kullmann, Glenister, Tozer, Donovan, Howarth, Callomon and Wright in a recent synthesis (House & Senior (eds) 1981). A comparison with an earlier synthesis (House 1963) based on data in the *Treatise* (Arkell *et al.* 1957) shows considerable refinement in recent years.

The ammonoid time-scale

Radiometric dating giving ages, in years, based on calculations regarding isotope decay and interpretations of geological relationships has great value in giving a time-scale for many past events. But for the sedimentary rocks of the Phanerozoic these methods are too uncertain in detail, and too limited in application, to be relevant to the fine scale discrimination of most past events. Here the Radiometric Time-Scale gives way to the Evolutionary Time-Scale which, being a relative scale, enables greater precision. The time order of distinct faunal assemblages, grouped as zones or subzones, may usefully be integrated with radiometric data to specify something of the resolution and precision available in the Evolutionary Time-Scale. It is convenient to define terms in the following way for purposes of crude analysis.

Precision, or P, is used here to refer to the discrimination in years of average zonal or subzonal units. If t is the span of radiometric time for a group of zones, and z the number of zones or subzones within that span, then $P = t/z$. Resolution, or resolving power, R, may be used to express the resolving power precision represents in terms of the radiometric time span for the zones considered. If T is the mean radiometric age for the time span t of z zones, then $R=T/P$.

As an illustration of the improvements in the Evolutionary Time-Scale the situation in the Jurassic may be considered. If T is taken at 170 Ma and the radiometric limits of the system as 135−205 Ma comparisons can be made. D'Orbigny in 1842 distinguished 25 faunal divisions within the Jurassic. Currently some 117 zones and subzones are recognized. The gradual improvement in precision and resolving power since d'Orbigny may be elaborated in the following table.

	z	P	1:R
1842−1849 d'Orbigny's divisions	25	2.8	1:60
1859 Oppel's zones	33	2.1	1:81
1922 Buckman's ages	42	1.7	1:100
1956 Arkell's zones	64	1.1	1:154
1978 Total zones and subzones	117	0.6	1:284

Finer subdivisions are recognized locally today but the table serves to indicate the significant progress in improved discrimination which has been made. Of course, the stratal sequence within zones allows far greater resolution.

The example given relates only to one group. The development of integrated zonations based on several fossil groups as well as improvements in the ammonoid scheme itself will open the way to greater precision. The actual figures change according to the time-scale used. Those above refer to figures given to this symposium. Using the scale (ii) of Lambert (Harland *et al.* 1973) the calculation for 1978 gives P = 0.5 and $1{:}R$ = 1:321, and this is probably an indication of the high error in the radiometric scheme.

The present state of zonal precision and its relation to resolution for several fossil groups using the Lambert (ii) scale is shown in Fig. 2. It may be pointed out that finer schemes of zonal subdivision have been proposed in the Phanerozoic. For example, Ramsbottom (1979) has suggested a span of only 25 000 years for fine divisions using goniatites

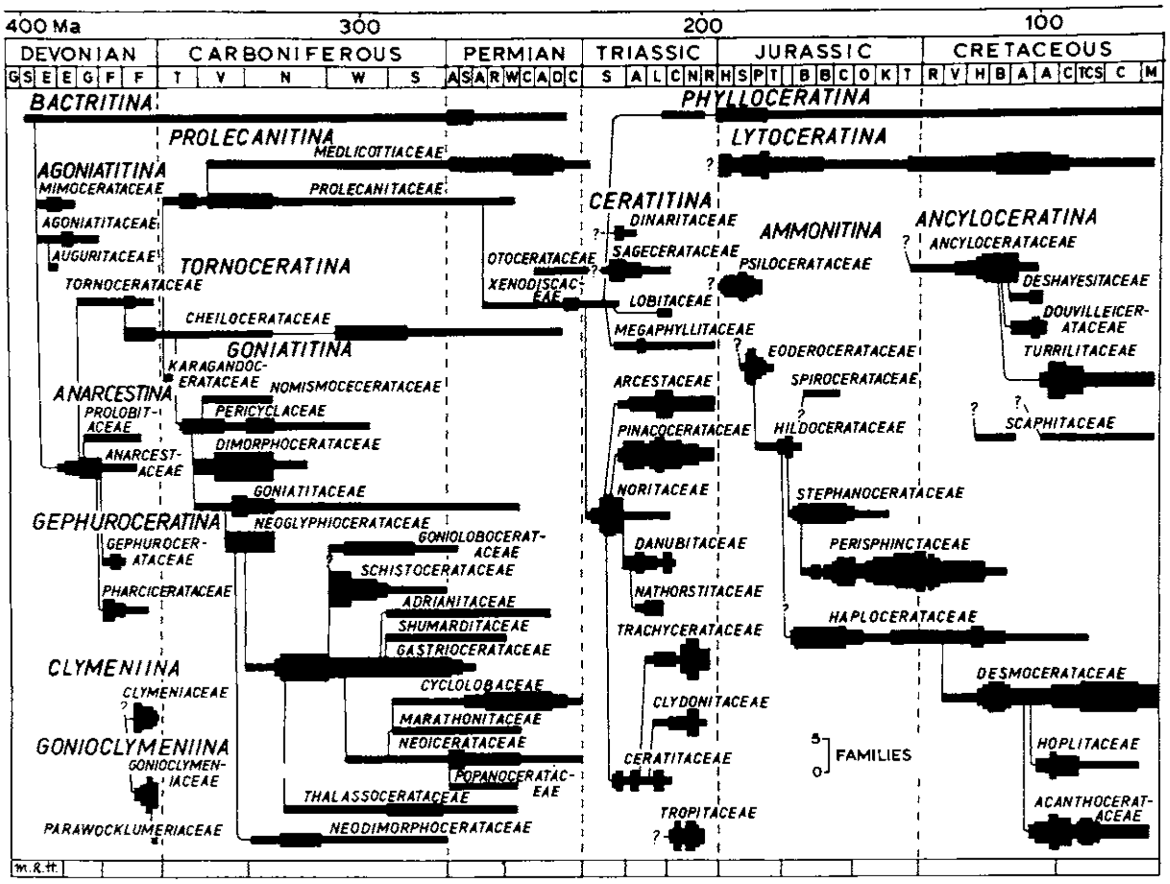

FIG. 1. Evolution of the Ammonoidea based on views of many specialists (from data in House & Senior (eds) 1981).

in the Namurian. Kauffman (1970) has indicated the refinement possible from overlapping ranges using several groups but facies changes would limit the widespread application of such methods. Figure 2 is intended to give a conservative view based on groups which have more than regional use.

Callomon (*In*: House & Senior (eds) 1981), who has contributed much seminal thought in this field, has suggested that Jurassic ammonite zones are likely to be more nearly of time equality than possible by other methods of approximation so that zones should be the basis for the analysis of processes and rates. Little has been done along these lines.

Young ammonoid stages and evolutionary strategy

The principal, primary distinguishing character separating the Ammonoidea from their ancestral Nautiloidea, if the Bactritina are included in the former, is the possession of a spherical to spindle-shaped protoconch or initial chamber. Other distinguishing characteristics, such as the marginal siphuncle, the thin shell, elaborate sutures and extremes of ornament are found in certain groups of Nautiloidea but it is the form of the protoconch which appears to be the fundamental novelty initiated with the Ammonoidea. Thus it would appear that, whatever secondary modifications were adopted in the Ammonoidea, and they were many, the initial radiation was related to the advantage to the larvae of the egg-shaped protoconch and its associated protoconch apparatus.

Protoconch

There is a substantial size range known for protoconchs. Erben (1964) gives the largest diameters known in primitive ammonoids as 0.6–0.75 mm, but the author has protoconchs of *Agoniatites* from Cornwall (Terebratum Zone) which reach 1.7 mm diameter. Among the smallest known are *Distichoceras* and *Taramelliceras* of 0.24–0.3 mm diameter recorded by Palframan (1966, 1967) from the Oxfordian. Recently there has been an improvement in knowledge of protoconch dimensional data. Here, seventy records are plotted against a time-scale based on information published by Miller & Unklesbay (1943), Spath (1934, 1951), Drushchits and Doguzhayeva (1974, 1981) and Zakorov (1974) and including Devonian data of the authors own. Some records only have imprecise stratigraphical information. The overall result of the plot (Fig. 3) suggests a very limited trend towards size reduction in later ammonoids. The range in volume which this represents is not so marked since the protoconch of early ammonoids is subspherical, that of later ones spindle-shaped. Volumes range from a maximum estimated as 1.5 mm^3 in *Agoniatites* to a minimum estimated at .01 mm^3 in *Distichoceras*

Ammonitella

Following the formation of the protoconch, the ammonoid grew a first whorl often in excess of 360° before the formation of a distinctive structure termed the nepionic constriction, the next distinctive growth stage. In recent years the useful term *ammonitella* has been used for ammonoid young up to this stage. Ammonitella dimensions which are available range from 0.48 mm diameter in Oxfordian *Distichoceras* (Palframan 1967) to the author's own estimate of probably 4.4 mm diameter in *Agoniatites*, a substantially greater range than that of 0.6–1.5 mm given by Zakarov (1974). Length of the

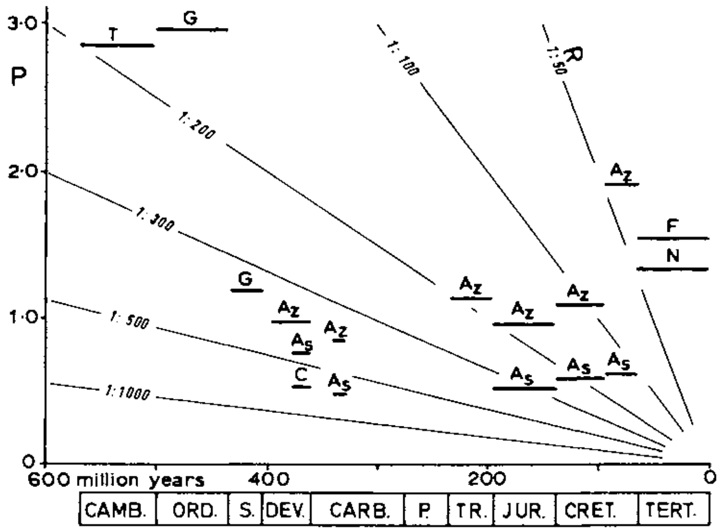

FIG. 2. Resolution and precision in some fossil groups of biostratigraphical value: A, Ammonoidea; C, conodonts; F, Foraminifera; G, graptolites; N, nannoplankton; T, trilobites: subscript z, zones; subscript s, subzones. For explanation see text.

body chamber needs taking into account but curiously, that of *Agoniatites* is short.

It has been argued that the marked change in shell ornament at the close of the ammonitella stage, with the subsequent development of elaborate apertural margins then forming, corresponds to the development of a functional hyponome and the achievement of an active life free from the egg capsule (House 1965). But ornament changes or, indeed, ornament details at all, are available for very few ammonoid genera (see review by Birkelund *In*: House & Senior (eds) 1981.

Evolutionary strategy

The importance of egg size in evolutionary strategy has long been recognized. Recently this has been expressed in the form of r and K selection modes. The trend to small protoconchs would suggest an overal trend towards the r-endpoint in which 'the optimal strategy is to put all possible matter and energy into reproduction, with the smallest practicable amount into each individual offspring, and to produce as many total progeny as possible' (Pianka 1970). That is, smaller protoconchs suggest a situation in which intra- and inter-specific effects are minimal and competition lax. However, it will be observed from the data that there is a spread of size at most periods. Details of the groups involved, their habit, and their adult size need to be analysed before any meaningful deductions can be made. Most of the data for the Cretaceous comes from the Phylloceratina and Lytoceratina, and this may explain the rather larger size which seems shown in the Cretaceous rather than the Jurassic and Triassic. Those two groups are thought to have lived in deeper water conditions and perhaps followed the K-selection mode. Data on Cretaceous Ammonitina is minimal. There is much interesting work to be done in this field, particularly in relation to protoconch size in periods of evolutionary radiation, but volumetric rather than linear dimensions need to be considered.

Data for the comparison of ammonoids with nautiloid groups is almost wholly lacking. *Nautilus* hatch at about 15.5 mm diameter (Zakarov 1974), and this is well over an order greater than the range for most ammonitella. This would suggest that ammonoids had a wholly different larval strategy than *Nautilus*.

The nature of the protoconch apparatus in the ammonitella also suggests novelty in ammonoid development. It seems highly likely that the caecum within the protoconch plays a rôle in liquid absorption and hydrostatic adjustment. The buoyancy which could result from liquid extraction and gas infill of the protoconch would lift the ammonitella from the sea floor and remove them from attack from benthonic organisms. This advantage would accrue the more if it were to be achieved at a smaller size. Thus r-selection need not be the only reason for small protoconchs.

Finally there is the important matter of the characters which distinguish the major ammonoid groups. The Devonian goniatites and clymenids represent some half of the 14 suborders currently recognized in the Ammonoidea, and these are distinguished by differing patterns of early sutural ontogeny (House *In*: House & Senior (eds) 1981). It seems inescapable that slight advantages in the novel modifications to the early stages had immense evolutionary importance. This is not so well documented for post-Devonian groups, but other factors, such as the progressive evolution of prosepta, may be significant.

The pattern of ammonoid evolution

There has been a refreshing and entertaining, if not always enlightening, spate of papers on matters relating to the evolutionary process in recent years. One debate has led to a supposed polarization into two opposing schools one invoking gradualism alone, the other punctuated equilibria, for the detailed nature of the process of change in evolution. The latter stresses periodic change separated by *stasis*, the former supposing a more continuously operating process indicated by slow sequential change. The distinction is largely made by proponents of the punctuated equilibria school who often apply the term Darwinian to gradualism quite unfairly implying that Darwin, who was brought up in the changeable English climate, and who bobbed around the stormy seas of the world under sail in a 235 tonner for almost five years, would for one moment have imagined that natural selection would operate only in an even way. As Berry (1982) has pointed out one of the best definitions of the punctuated

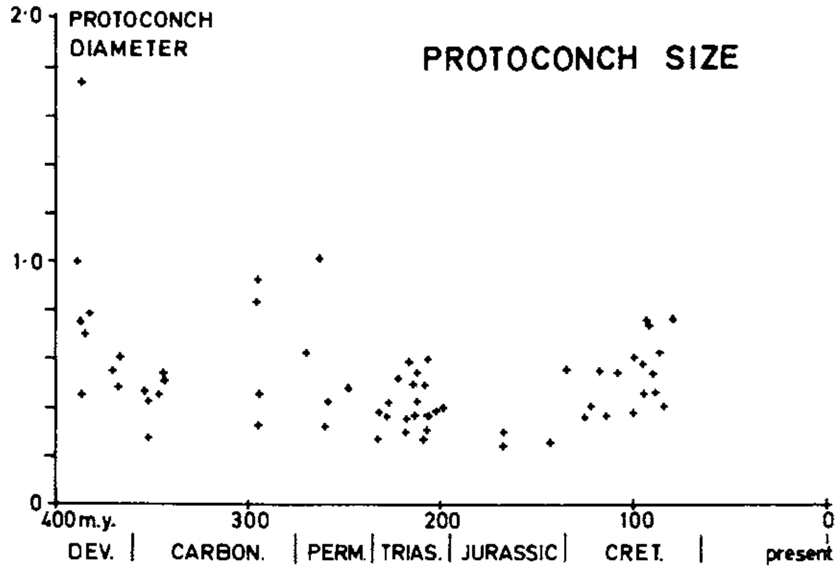

FIG. 3. Protoconch diameter plotted against age for ammonoid records available. For sources see text.

equilibria position is given by Darwin.

This debate has generally lacked contribution from those that have given serious and prolonged study to the fossil record. But the information from various groups is different, and even for one group assessment can be done at many levels, from different subgroups, and in the context of differing environmental parameters. Failure to disentangle different aspects is part of the muddled thinking which makes the current theoretical debate so unhelpful. In considering ammonoid evolution here it seems best to separate discussion into rough taxon levels, but higher grade evolution is but a summation of surviving lower grade changes so that macroevolution is a matter of scale and focus.

It is argued here that the fluctuations in ammonoid evolution are but a reflection of changing environmental and evolution-controlling factors. As such the record is a bioseismograph of such events. Analyses may therefore tell as much about the events as the process. A crude tot of basic data for genera and families is given in Fig. 4, and this gives a crude view of fluctuation in diversity in the group through time. The generic plot is based on the *Treatise*, the crudeness of the stratigraphical framework of which renders it of little serious value. The family data is much more up-to-date (most data is from House & Senior (eds) 1981) and the stratigraphical discrimination is at zonal or stage level rather than series level.

The species level

Some comments are needed about the nature of occurence of ammonoids in the fossil record. Generally they occur at specific levels separated by a rock sequence without ammonoids. In the Carboniferous the term goniatite band is used to illustrate this. A similar pattern is seen in the Devonian (House 1978), even the so-called continuous sequences of the cephalopodenkalk facies are now known to be very condensed and to exhibit non-sequences and hardground horizons. A similar pattern is well shown in the European Mesozoic. This is so well known to specialists that the disjunct nature of the ammonoid record is rarely pointed out. Thus evolutionary studies are likely to be based on interrupted sequences.

Secondly, the bands themselves often represent levels favourable to preservation for syngenetic or diagenetic reasons and hence cover a wide time span in themselves. Ramsbottom (1979) has estimated that for Namurian goniatite bands each horizon may be of the order of 10000–12500 years in span. Each will contain shells of different age which result from growth under different environmental controls.

Thirdly, when rich faunas are known at particular levels an incredible range of variation is recognized which often involves a continuous morphological series with all intermediates recognizable. Good examples have been described by Reeside & Cobban (1960) in neogastroplitids, by Howarth (1973) in dactylioceratids and by Arkell (1935–1948) in cardioceratids (although he may not have admitted the continuity). Kennedy & Cobban (1976) have reviewed some other examples. To this must be added the problem of sexual dimorphism, now claimed in virtually all major groups, and in many of them producing highly different morphologies.

Fourthly, much of ammonoid evolution seems to result from changes in the ontogeny of the organism. Paedomorphogenesis and other allometric styles have been invoked frequently. Yet very few statistical studies have been attempted on successional ontogenies. Studies of adult only, or incomplete specimens, is likely to be misleading, yet ammonoids are one of the groups to preserve the morphology of all growth stages in the adult stage.

Fifthly, there is the semantic problem relating to specific names, or even genera. These are usually used arbitrarily, are subjective in their limits, bear no relation to biological species, and are used variously by authors in space and time. Names have great use in communication of morphological information, but their coherence has to be tested mathematically before they can be used as a basis for evolutionary studies. This can hardly be said to be usual practice.

Finally poor fossilization, preservation, or collection record add different types of uncertainty to the basic premises of most studies.

Quite a different factor is the lack of scientists prepared to spend enormous time in the collection, preparation, and analysis of faunas locally and internationally. Fame can be more easily made elsewhere and fundamental studies of the

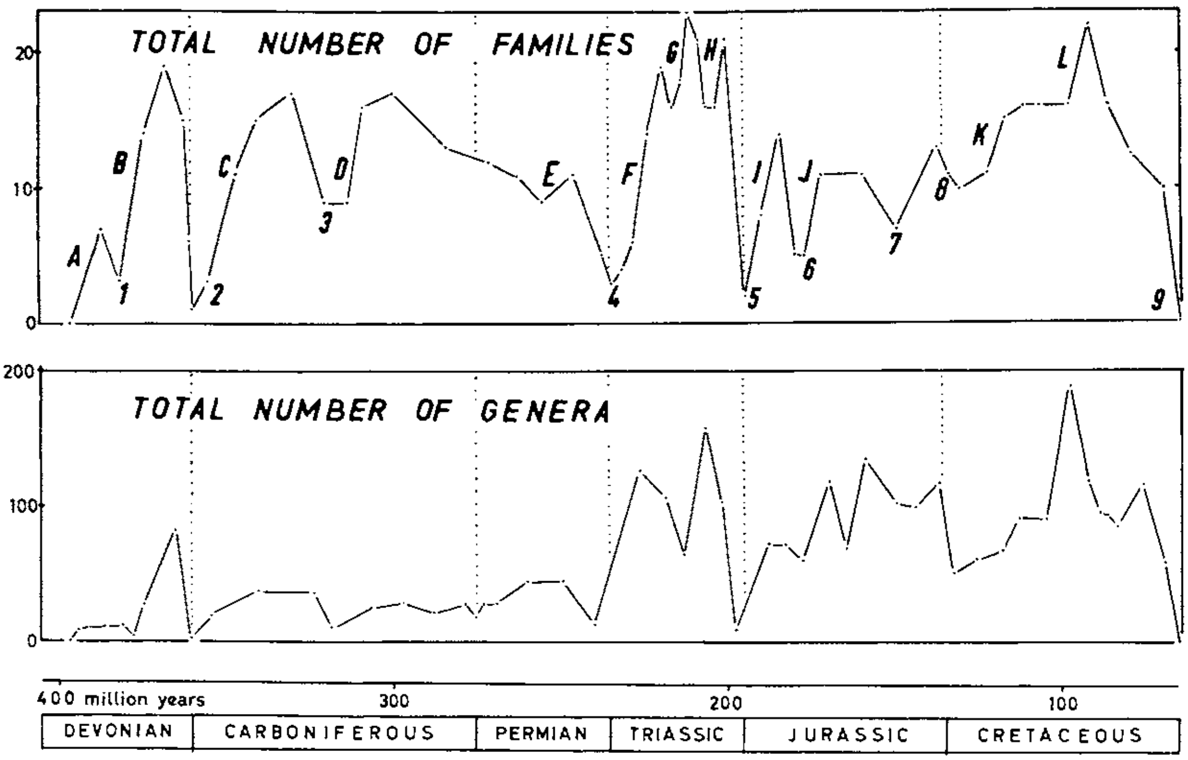

FIG. 4. Numbers of ammonoid families through time (based on data in House & Senior (eds) 1981), and numbers of ammonoid genera through time (data from Arkell *et al.* 1957 and House 1963).

sort required are not fashionable.

Nevertheless, the ammonoid literature is full of records of morphological discriminations which contribute to the problem and a number of small-scale detailed studies have been published. In a study of successional allometries in evolution in the genus *Tornoceras* (House 1965) attention was drawn to the different patterns shown by different morphological characters, a point made in Rowe's classic study of *Micraster*: also the masking affect of phenotypic changes in any attempt to elucidate trends in shell form was noted whilst other characters showed more convincing regularity in trend. A recent study of old data on *Kosmoceras* (Raup & Crick 1981, 1982) illustrates how important it is first to clarify the premises: in this case, tests for dimorphism, tests for validation of Brinkmann's specific and generic groups, in addition to comments made by Schopf (1982) limit the value of deductions drawn. In most studies failure to provide modern analysis of allometric effects and to use evidence applying to similarly aged groups (where this can be recognized) are particular weaknesses.

Generic to family level

At a higher level, broad trends of morphological evolution are too well attested for their clear evidence of the importance of progressive and gradual evolution to be ignored. Examples include the progressive elaboration of the suture in *Sporadoceras* (Petersen 1975); increase in lateral lobes in the Prolecanitidae; the broad trend from unornamented psiloceratids to ribbed and eventually bisulcate and tricarinate venters of the arietitids; the forward sweep of ribs leading from acute venters to corded kells as independently in the amaltheids and the cardioceratids; the progressive development of disjunct ribbing and smooth bands on the venter in many groups, including the parkinsoniids, kosmoceratids and aulacostephanoceratids, and so on.

One of the most noticeable characteristics of the sequence of ammonoid faunas is the successive dominance of particular generic or familial groups and their replacement by others. In the Devonian there are the ammonoid stufen, in the Carboniferous the genus-zones such as E, H and R, and the 'ages' into which the Triassic and Jurassic have been divided. The successive faunas have a good deal of coherence and suggest that one group has merely replaced the other without much change in the gross morphological range and abundance in the ammonoid faunas as a whole. Even at low taxon levels relationships with transgressive pulses have been suggested for the Devonian (House 1975a), Carboniferous (Ramsbottom 1979 and other papers), and there has been a traditional argument in the Jurassic for successive new stocks arising in Tethyan areas and migrating into boreal areas. The last is proving to be an oversimplification.

Arambourg's term evolutionary relay is a neat way of expressing this pattern — a rich, basic ammonoid fauna continuing, merely the taxonomic group which dominates that fauna changes. If the full link with transgressions is demonstrated it is likely that the novel surviving groups are those that survive tbe high selection pressures associated with the intervening regressional events. The term 'take-over' has also been used for the relay pattern, but seems more usefully confined to cases where a quite different group takes over the niche as, for example, in the successive replacement of graptolites, dacryoconarids, and planktonic ostracods in Devonian pelagic environments.

A Devonian example of the relay pattern is illustrated here (Fig. 5) which shows the diversity of three replacing groups in the late Middle and early Upper Devonian. The first group comprises the sample-sutured Anarcestina and Agoniatitina, and only a restricted remnant survives. The replacing group, the Gephuroceratina, is characterized by a novel ontogeny with umbilical lobes developing in the suture. In turn, this group is replaced in dominance by the Cheilocerataceae

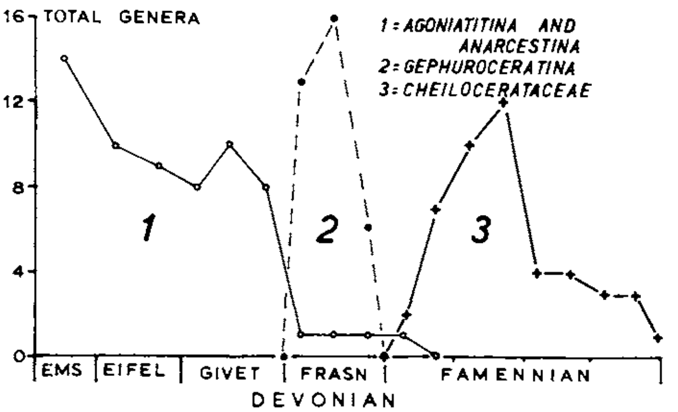

FIG. 5. An example of relay type clade replacement in Devonian ammonoids. The earlier group is virtually extinct before the rise of the replacement taxon.

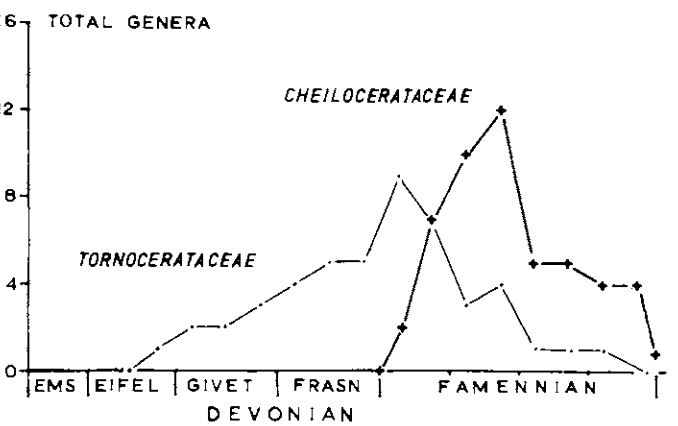

FIG. 6. An example of antipathetic clade replacement in Devonian ammonoids. The decline of the earlier group occurs *pari passu* with the rise of the replacing group.

which reverts to a simpler suture but with different lobe genesis and a different apertural margin. In turn these are replaced by the clymenids, and so on.

The data indicates that there is a decline *before* the rise of the replacing group, as if the stock was confined and the novel surviving groups then more freely radiate. Although illustrated at a rather high taxon level, this pattern is seen also at generic level and species-group level. Ramsbottom has argued a close relationship between cyclothems and this type of change in the Namurian.

Quite a different pattern is shown when there is an antipathetic decline of one group and an increase in another, as illustrated (Fig. 6) by the replacement of the Tornocerataceae by the Cheilocerataceae in the Devonian, and may be seen for other groups on Figs 1 and 7. In tbe Devonian example the chief difference between the two groups is the pattern of the aperture.

These two contrasting styles of replacement suggest that the first is environmentally controlled with near extinction in one group being followed by *de novo* development of another giving a disjunct relay style. In the second, the gradual encroachment of one group on another may be relating to intra-group competition related to some inherent biological advantage in the biological characters of the replacing group, one might term this the antipathetic style. In the first case, the surviving morpbology will be related to characteristics required to survive the high selection period stress and not necessarily to adaptation to the environment the succeeding transgressive pulse (or other change) presents. It is the subsequent radiation which exploits that.

When a range of such examples are examined (Fig. 7) most show an overlap of groups except where events effect many groups and which will be referred to later. Clade shapes may be examined by the methods of Gould *et al.* (1977) but each will show a finer scale pattern at lower taxon grade which it would be unwise to ignore even if such documentation is patchy in the record of the Ammonoidea as a whole.

Highest taxon levels: extinctions

As will be clear from Fig. 1 there are periods in ammonoid history when several unrelated stocks became extinct at about the same time. Such events may be satisfactorily explained by extra-group factors such as changes in the environment and these may most likely be primarily physical changes although their immediate effect may be secondarily biological by, for example, modifications of the food chain, or changes in competition, either direct or indirect. There are nine particularly critical periods of extinction in the history of the Ammonoidea (Fig. 4). The data for family extinctions is based on modern data (House & Senior (eds) 1981) whilst that for genera is based on an earlier review (House 1963) which was based on the older data in the *Treatise* (Arkell *et al.* 1957) with its very crude stratigraphic control (full data are not yet available to revise it).

These nine events are not dissimilar in type to sudden extinctions described in a previous section but many taxa of contemporary ammonoid are affected rather than one group alone. In the particular events listed below, substantial (1–8) or total (9) extinction of the existing families or higher taxa resulted.

It has been argued by Van Valen (1973) that within an ecologically homogeneous high taxon group, evolution occurs at a stochastically constant rate analogous to the radiometric decay constant. Raup (1975) has referred to this as Van Valen's Law, but analogies with a closed atomic system are obviously inappropriate to the complex, open, and fluctuating real biological situation. The extinction rates which may be calculated have no predictive value and hence the term 'Law' cannot be used properly. The theory is a postulate only. Another such postulate might be that in an ecologically established biological system at steady state, high taxon extinctions will only take place as a result of physical changes in the system. If of no other use, at least the latter postulate would encourage the analysis of extinction events that would contribute to the elucidation of the physical events which are the primary cause and control. The increasing evidence from event stratigraphy might suggest that the physical controls may be regularly triggered by mid-ocean ridge activity, sea-level changes, climatic cycles and the like. The only way in which such events will be documented in time is by the detailed study of the biological record and the bioseismograph it represents.

1. Late Givetian The Middle-Upper Devonian boundary was formerly defined by the first major extinction event in the history of the Ammonoidea, but the redefinition of that level in 1982 by the Devonian Subcommission to the base of the Lower *asymmetricus* Zone now places the event within the late Givetian when the Agoniatitina and most Anarcestina became extinct. So far as is known only one genus appears to have survived (*Tornoceras*) but representatives of two other

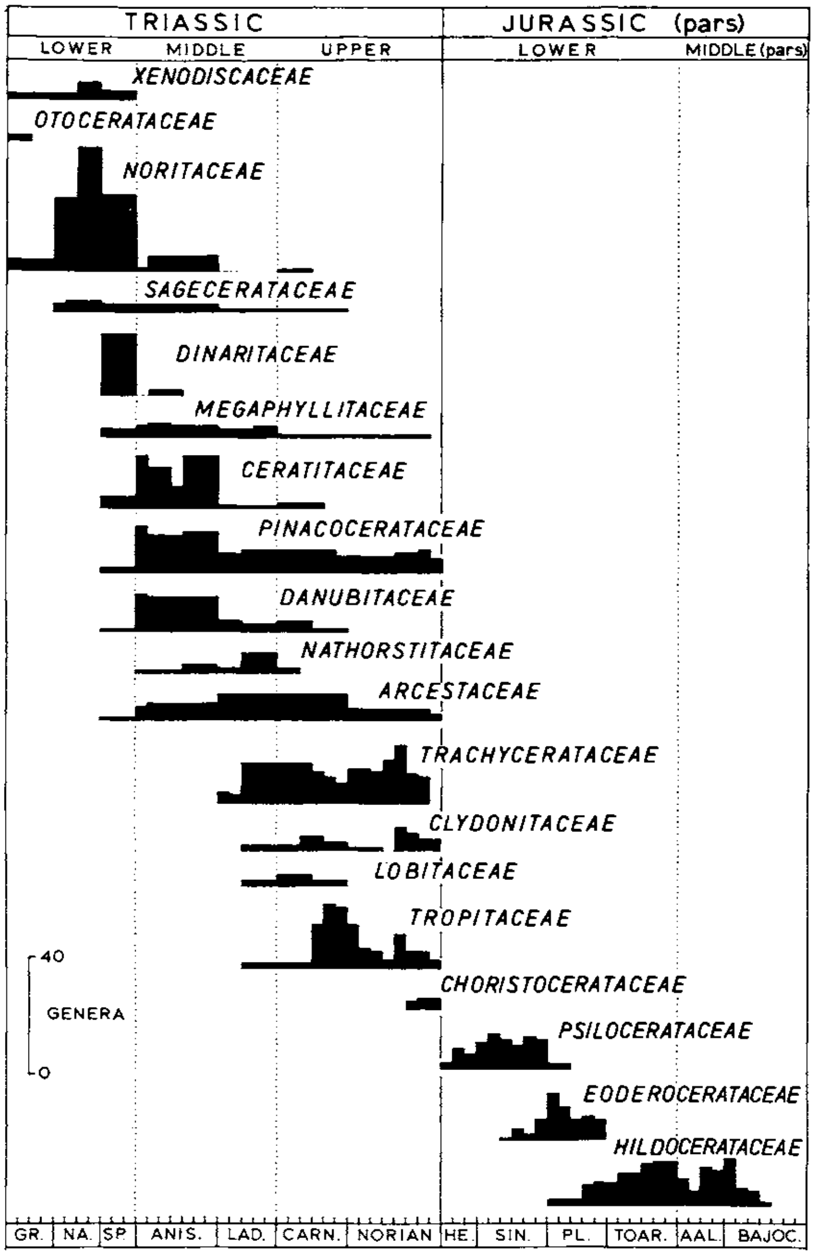

FIG. 7. Successional replacement of superfamilies in Triassic and early Jurassic ammonoids. (Based on data of Tozer *et al. In*: House & Senior (eds) 1981.)

stocks must have also survived (anarcestids and prolobitids) if current views of evolutionary relationships are taken into account. The environmental cause of this event has been argued elsewhere (House 1975a, b) to be the changes associated with the Taghanic onlap of the New World and similar transgressive events in the Old World. Detailed biostratigraphy of the Givetian is not at a high order of precision but it appears that the taxon decline is sudden as illustrated in Fig. 5.

2. *End Devonian* Near the close of the Famennian, the most notable extinction is of all *Clymeniina* (and also the *Tornocerataceae* and *Sporadoceratidae*). Some precise documentation of this is possible (Fig. 9(a)) especially for the terminal part of the Wocklumeria Stufe (Fig. 9(b)) as a result of the study by Schindewolf (1937) of the sequence in the Hönnetal. There the Hangenberg Schiefer introduces a deeper water facies in which clymenids are last seen. Less precise sequences are known in a number of other areas. A late Devonian transgressive pulse is indicated in North America in the late Bedford Shale and Exshaw Shale.

The decline in diversity seems to be a progressive one during the late Wocklumeria Stufe at family, generic and specific levels (Fig. 8(a), 9(a), (b)) giving decline rates, respectively, of 4.3, 5.2 and 10.4 per Ma. Thus the terminal extinction has the effect of a *coup de grâce* on an already impoverishing situation.

3. *Mid Namurian* In the E_2 zone of the Namurian there are some 10 or 11 families present which do not survive into the succeeding H zone (Kullmann 1981) (Fig. 8(b)). This is sufficiently close to the Mississippian–Pennsylvanian boundary to seek a possible cause in the regressional events which mark the acme of the Kaskasia–Absaroka offlap in North America and herald significant clastic input leading to the establishment of coal measure facies in Europe. No detailed analysis of this appears to have been attempted apart from that by Ramsbottom (1979).

4. *End Permian* Progressively from the late Carboniferous there is a decline in diversity at family level in the Permian at a rate of 0.38 families/Ma (Figs 4, 9(c)), a decline so regular in its pattern that the survival of the ammonoids at all is a

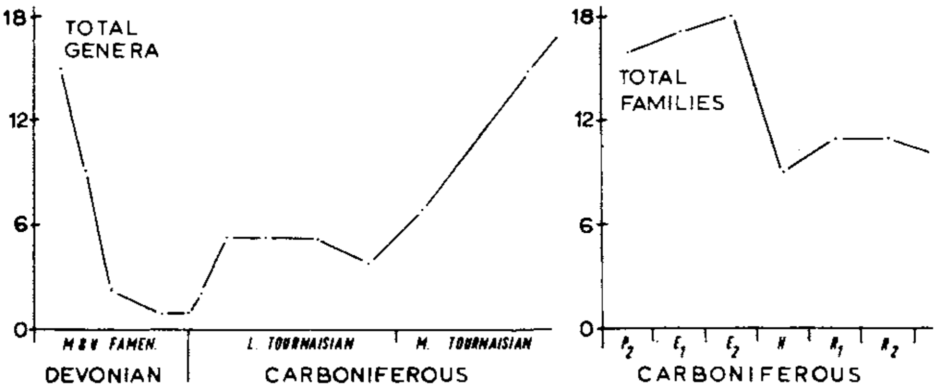

FIG. 8. Recovery after extinction periods illustrated by generic diversity for the Devonian–Carboniferous boundary and family diversity after the Carboniferous E_2 extinctions. (Data from House & Kullmann, *In*: House & Senior (eds) 1981.)

surprise. No members of the Goniatitina survive the Carboniferous–Permian boundary (Glenister *In*: House & Senior (eds) 1981). The long-sustained decline characterizes many other groups (House 1963) and is well documented (contributors to Harland *et al.* (eds) 1967) for brachiopods particularly by Waterhouse & Bonham-Carter (1976). A range of the possible causes were reviewed by Rhodes (*In*: Harland *et al.* (eds) 1967).

5. *End Triassic* This break was so complete (Figs 7, 9(d)) that there is doubt as to how any transition to Jurassic ammonite groups took place (Wiedmann 1973; Tozer 1981a). Perhaps less than 16 genera survived to the topmost Crickmayi Zone of the uppermost Norian (including Rhaetian) but none of these may be in common with Hettangian genera. Here too there is a steady decline in diversity through the Norian (Fig. 9(d)) suggesting rather a longer-term cause than a single extinction event. The Norian decline rate is about 3.4 families/Ma. In a review of other marine invertebrate groups by Hallam (1981) a similar pattern is noted and he follows others in ascribing this to a general regressive setting but adds that Hettangian anoxicity associated with transgression is significant, although this would be after the demise of the Ceratitina.

It has still to be demonstrated how Norian marine extinctions relate in time with the disjunct reptile extinctions documented by Tucker & Benton (1982) and said to be mid Norian, or those of other tetrapod groups (Bakker 1977) which, in general, show loss by rather rapid diversity decline before the late Norian. The significance of general correlation with regressions of epeiric seas, however, seems inescapable as a major cause.

6. *Late Lower Jurassic* The Toarcian deepening (Hallam 1978) and Aalenian–Bajocian regressions (Ager 1981) marks a replacement of the surviving Liassic Eoderocerataceae by the Stephanocerataceae, Perisphinctaceae and Haplocerataceae, the three novel groups arising from the

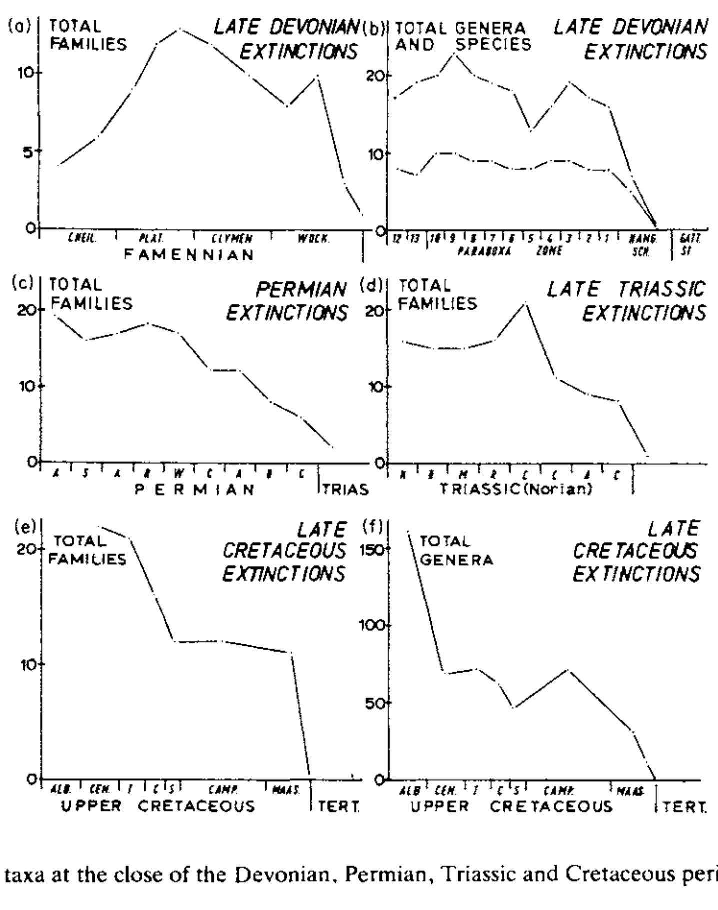

FIG. 9. Data on extinction of taxa at the close of the Devonian, Permian, Triassic and Cretaceous periods. For sources of data see text.

Hildocerataceae in the Aalenian (Donovan & Howarth *In*: House & Senior (eds) 1981), with only the Hildocerataceae surviving in all these stages. The decline of the Liassic groups may be more complex and be related to a late Pliensbachian shallowing which is well shown in Europe and which only the Dactylioceratidae and Hildoceratidae of the Ammonitina survive.

7. *Middle–Upper Jurassic Boundary* The loss of the Reineckiidae and Kosmoceratidae at the close of the Callovian gives a sudden change to Tethyan and Boreal faunas respectively. Other changes are at lower taxon grade. How far Callovian transgressive pulses interplay with province patterns to produce this result is not clear and is complex (Ziegler 1981), but it may follow after maximum Jurassic seaboard onlap.

8. *End Jurassic* Late Jurassic and early Cretaceous ammonoid provincialism has been emphasized by Casey (*In*: Middlemiss *et al.* 1971; Casey & Rawson 1973) and by both Rawson & Calloman more recently (*In*: House & Senior (eds) 1981). An analysis by Gordon (1976) suggests that ammonoid provincialism reached its maximum development at this time. The taxon tots (Fig. 4) show extinctions associated with this. How far this is due to regressional effects alone or changing provincialism or the spurious taxonomic effects which would be associated with these is not clear.

9. *Late Cretaceous* The taxonomic data indicates a long-continued decline in ammonoid diversity through the late Cretaceous (Fig.9(e), (f)) and this has been documented by Hancock (1967), Kennedy (1977) and others. This pattern is accompanied in the Maastrichtian by a progressively restricted geographic distribution of known faunas. A long term decline in diversity is shown by other groups also, including Reptilia. Of the several dozen competing hypotheses to explain Cretaceous–Tertiary houndary extinctions those that take account of this history are the most plausible. Of these, the palaeogeographic regressional setting seems, again, highly important. The relation of major Mesozoic extinctions to continental flooding has been illustrated by Kennedy (1977) and the interpretations follow those proposed by Newell (1963). Whilst regression may be most important in detail there are exceptions and it is to be expected that environmental change (of whatever cause) is bound to present hazard to some groups even if it presents opportunity to others.

Attention needs to be drawn to some of the innaccuracies in using palaeogeographic maps to assess past sea-level movements. Those relating to presumptions on the form of the hypsographic curve are obvious. But the maps themselves usually carry interpretations far beyond the outcrop data. Frequently such maps are quite inaccurate regarding time. National practice regarding boundary definitions, such as those at the Siluro-Devonian, Devonian–Carboniferous, Middle–Upper Jurassic or Lower–Upper Cretaceous make comparisons between continents misleading to the uncritical. Further, the cyclicity within systems means that for any part of a system much crude generalization is required. More potential lies in terms of facies movements and the geometry of cycles themselves.

Higher taxon levels: appearances

At a number of times there are marked periods of apparently significant diversification at family level. These are marked A–L on Fig. 4. On Fig. 1 the same events are indicated by the times when several novel groups appear. As is usual with the appearance of new groups the precise point of origin often is uncertain, but the subsequent diversification is well documented, and generally this is spread over a longer period of time than is the case with extinctions.

(A) Appearance of the Ammonoidea The first ammonoids now appear to have arisen in the Emsian (House *In*: House & Senior (eds) 1981) but the actual sequence cannot be said to be well described but there is new documentation in a recent Czech study (Chlupač & Turek 1983). Within the Emsian the new group becomes world-wide in its distribution. The rise and distribution seems related to an early Emsian transgressive pulse represented by the Zlichov–Daleje, Hünsruck Schiefer deepening of Europe and the Regularissimus Zone event of the USSR.

(B) Mid and late Devonian Several successive evolutionary diversifications are represented leading to the peak of family diversity in the late Devonian (Fig. 4). Successively entering are faunas dominated by the Pharcicerataceae (now late Givetian), Gephurocerataceae (Frasnian) (Fig. 5), Cheiloceratateae (Lower Famennian) (Fig. 6), and the Clymeniina (much of Fig. 9(a)). The first two may be related to the Taghanic onlap and Frasnian transgressions (House 1975a). Although an early Cheiloceras Stufe transgressive pulse seems clear and the initiation of the Cheilocerataceae with this seems plausible. Although the Annulata Schiefer deepening occurs in the Platyclymenia Stufe where clymenids appear that event seems to postdate the actual appearance.

(C) Early Carboniferous The rise in diversity following the end Devonian extinctions is slow (Fig. 8). Again it is characterized by the appearance of groups exploiting new ontogenetic patterns, either of the U-type ontogeny (Prolecanitina) or of the A-type modified by the division of the ventral lobe (Goniatitina). The time involved in the diversification is so great that it can only be correlated in general terms with the early Carboniferous transgressions, the pattern of which seems stepped.

(D) Late Carboniferous Almost all Permian groups of Goniatitina were established following the E_2 extinction event in the following late Carboniferous. Again the rise was slow (Fig. 8). The groups involved are mostly characterized by much more complex sutural styles than typical of the Goniatitina of the Lower Carboniferous. Also the geographical distribution is different and a palaeogeographic cause may be inferred, perhaps the slowly successful Absaroka onlap.

(E) Late Permian A late Permian (Wordian to Dzhulfian) resurgence seems weakly recognizable from both family and generic data (Fig. 4). There is a change from dominance by Goniatitina to early Ceratitina during this period and Glenister (*In*: House & Senior (eds) 1981) has drawn attention to the role of provincialism at this time.

(F–H) Trias The radiation of the Ceratitina is spectacular (Tozer 1981a) but four phases seem recognizable within it, the first being the late Permian production of the Xenodiscaceae and Otocerataceae (Fig. 1). The second (F) corresponds to the (L) Triassic rise of the noritaceans and dinaritaceans (Fig. 7). The third (G of Fig. 4) is the Spathian appearance and Anisian radiation of a range of derivatives from the Noritaceae. The fourth (H) is the Ladinian and Carnian replacement of these by derivatives mostly from the Ceratitaceae (Fig. 1) and includes the appearance of the cryptic Tropitaceae and Choristaceae. It is not clear what the control of these phases might be.

(I) Early Jurassic A new range of diversification followed the late Norian extinctions and generally this can be related

to the widely recognizable early Lias transgression. The Lias succession is of the relay-clade type and, again, the point of origin of several groups is uncertain (Lytoceratina, Psilocerataceae and Eoderocerataceae). Some relationship to the Hettangian and Toarcian deepenings seems probable.

(J) Middle and late Jurassic Four novel superfamilies appear in the Aalenian–Bajocian (Fig. 1) and so far as the Ammonitina are concerned these represent completely replacing clades for Lias groups. Ager (1981) has argued that the Bajocian represents a climatic high, and this may be a contributary factor. The several transgressions within the generally shallow water Bajocian of Europe are not sufficiently correlated internationally to infer causal mechanisms.

On the larger scale, the events of the post-Bajocian Jurassic evolution are contained within stable superfamilies four of which continue into the Cretaceous (Fig. 1). The unlettered rise shown on Fig. 4 in the late Jurassic and earliest Cretaceous corresponds to the diversification of the Perisphinctaceae probably associated with a combination of the provincialism already discussed and the climatic optimum invoked by Ager for the Late Jurassic.

(K) Late lower Cretaceous The significant familial radiations of the Barremian, Aptian and Albian (Wright *In*: House & Senior (eds) 1981) suggest recouping following extinction period 8 (Fig. 4). In general this would seem to be associated with the long-known pattern of increasingly successful transgressions documented by Hancock & Kauffman (1979) and by Vail *et al.* (1977) but in the latter only vaguely correlated with the available precise biostratigraphy. Three phases seem to mark phases of the ammonoid diversification.

(L) Late Cretaceous The peak in family and generic diversification in the Cretaceous corresponds with the Cenomanian transgression although most of the stocks involved appear in the latest Albian. Kennedy (1977) has emphasized the correlation between diversity and continental flooding which reaches a Mesozoic acme at about the same time. A correlation of local abundance of scaphitids and baculitids with a late Turonian transgression is documented by Kauffman (In: Kauffman & Hazel (eds) 1977) and Hancock & Kauffman (1979) but this may not indicate a high taxon event internationally. A subsidiary peak in the Campanian shows in generic plots (Fig. 4, Kennedy 1977) but only as a slight step in the family decline (Fig. 4) and not at all at superfamily level (Fig. 1). This correlates reasonably well with the maximum extent of Cretaceous shallow-water equable seas (Hancock & Kauffman 1979) in which high diversity is to be expected.

The foregoing discussion suggests a general confirmation of the thesis that regression gives extinction and transgression leads to novel groups appearing. But in many cases the correlation leaves much to be desired and the effect is different at different taxon levels. Nevertheless it seems clear that an integrated analysis of evolutionary fluctuations in time in many groups of organisms and collated with more precise palaeogeographic and sedimentological models than usually used will offer the potential for great refinement in the documentation of global periodic events.

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The mass-age distribution of Phanerozoic sediments

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SUMMARY: The mass of the Phanerozoic sediments is about 2.1×10^{18} metric tons, and between a quarter and one-third of it is distributed on the present continental margins and deep sea floor. The survival rate (surviving mass per unit time of deposition) seems to decrease exponentially with advancing age back to the Carboniferous, beyond which the tail of the distribution holds up and is somewhat irregular. The distribution for the past 300 Ma can be expressed by the equation $\log s = 10.01 - 0.24t$, where s is the survival rate in metric tons per year and t is in units of 100 Ma. For a constant sediment mass with constant probability of destruction, this corresponds to a mean sedimentation rate since Devonian time of 10^{10} metric tons per year and a half-life for the post-Devonian mass of about 130 Ma. The overall distribution, however, is the sum of the distributions for three major realms, cratonic, marginal and pelagic, each with its own characteristic pattern. The last two account for most of the exponential-looking trend in the later Phanerozoic, but it is not clear (nor, in the case of the pelagic sediments, likely) that they are themselves exponential in character.

The surviving masses (per unit time) of the Phanerozoic Systems tend to decrease with advancing age. This trend was revealed by a series of global volumetric estimates for the Devonian through Jurassic Systems (Ronov 1959) and extrapolations to the rest of the Phanerozoic based on a rough correlation between the volumes of the Systems surveyed and their maximum known thicknesses (Gregor 1967). Ronov and his colleagues have lately published (Ronov *et al.* 1980) complete volumetric estimates for those parts of the Phanerozoic Systems that underlie the global land surface. These estimates (after subtraction of the volcanic component and conversion from volume to mass) are summarized here in Table 3, Column 4. To them, in order to obtain the complete Phanerozoic mass-age distribution, must be added the sediments of the continental margins and deep sea. For convenience of study, the submarine realm can be divided into:

1. 'Passive' continental margins as defined by the US National Academy of Sciences (1979) and by Sclater *et al.* (1980). Examples are the margins bordering the Atlantic and Arctic Oceans and the Indian Ocean as far east as the Andaman Sea;
2. Marginal basins (Sclater *et al.* 1980) separated from ocean ridges by tectonic barriers (trenches, arcs, continental crust). They include such regions as the Caribbean and Mediterranean Seas and the marginal basins of the western Pacific;
3. The deep sea floor. The areal distribution of these submarine sedimentary environments, as determined by Sclater *et al.* (1980) is, in millions of km^2: passive margins, 52.2; marginal basins, 26.9; deep sea floor, 281.7.

Passive margins

The most readily available data on passive margins are in the form of seismic profiles, many of them divided stratigraphically. To estimate the total sediment mass 84 profiles were studied, of which 37 furnished additional data on the volumetric ratios of the different geologic Systems. Space does not permit acknowledgement of the primary sources; all but one have been published in Maxwell (1970), Nairn & Stehli (1973, 1974 and in press), Burk & Drake (1974), Watkins *et al.* (1979), or US National Academy of Sciences (1979). The single exception is a profile of the Douala Basin (Gulf of Guinea), for which I am indebted to Dr E. J. Webb. Cross-sectional areas were determined by reprinting the profiles (enlarged where necessary) on a medium weight paper of good quality (Nekoosa Ivory 20-lb) and cutting them out and weighing the pieces. (This procedure was found to give better reproducibility than could be obtained by planimetry.) Each area was then divided by the length of the profile to give a mean thickness, and the results were grouped and averaged for representative regions whose areas (32 million km^2 in all) were estimated from the maps of Briden *et al.* (1974) and then prorated to make up the total of 52.2×10^6 km^2 (see Table 1).

There is need for caution in evaluating the average thicknesses obtained by this method, because (1) the profiles are sometimes incomplete and then tend to emphasize the thicker parts of the margin under the shelf edge and slope; (2) the basement generally undulates along the margin, and for economic and other reasons, thickly-filled basins have been more often surveyed than the thinner, intervening swells; and (3) while some basins (e.g. in the north-west Atlantic) are elongate enough for two-dimensional treatment in transverse section, in others (as in the south Atlantic) the mean thickness may be overestimated in this way, as happens in a spherical segment. The combined effect of (1) and (2) can be considerable: the mean sediment thickness obtained from 10 profiles between Newfoundland and Cape Hatteras was 4.2 km, while that derived from an isopach map by Drake *et al.* (1959), based on 297 seismic observations covering the same area, was only 1.8 km. Smaller discrepancies are the rule, however, and a study of other stretches of passive margin indicates that the true mean thickness is about 64% of the mean determined from profiles. The 'shape' effect (3) was tested by comparing profiles in two south Atlantic basins (Colorado and Malvinas) with well-dispersed seismic observations (Ludwig *et al.* 1979a, b). No significant discrepancy was found, and the effect is here neglected.

The volumes shown in Table 1 have been calculated by multiplying the prorated areas by the corresponding mean thicknesses determined from the seismic profiles and including a factor of 0.64 to correct for over-representation of the thicker parts of the margin as explained above. The 37 stratigraphically-divided profiles (weighted according to the areas they represented) were used to estimate the volumetric ratios of the different Systems, and the whole was converted to mass using an assumed density of 2.5 $g.cm^{-3}$. The results are entered in Table 3, Column 5. As can be seen from their

TABLE 1 Mass-age distribution of sediments on passive margins

Province	*prorated area a (10^6 km^2)*	*N*	*mean thickness, t (km)*	σ	*volume = a × .64t*
N. Atlantic, Arctic	22.45	24	4.29	1.76	61.64
Gulf of Mexico	1.57	3	6.10	3.36	6.13
S. Atlantic	8.35	38	3.43	1.82	18.33
Indian-Southern	18.27	13	3.85	1.39	45.02
NE Pacific	1.57	6	3.52	1.78	3.54
	52.2	84			134.7 × 10^6 km^3

Total mass at density 2.5 g.c^{-3} = 337 × 10^{15} metric tons.

Weighted distribution estimated from 37 stratigraphically-divided seismic profiles: Triassic, 16.8 × 10^{15}; Jurassic, 65.3 × 10^{15}; Cretaceous, 114.3 × 10^{15}; Tertiary, 140.4 × 10^{15}.

Areas from Briden *et al.* (1974) and Sclater *et al.* (1980). N, number of profiles measured; σ, standard deviation of mean.

standard deviations, the thickness estimates are subject to considerable error (on the order of 50%). The Triassic may be under-represented for want of data on the Madras-Orissa shelf, the offshore part of the Karroo basin, and other areas known to contain large masses of Triassic sediment.

Marginal basins

The most comprehensive survey of sediment thicknesses in marginal basins presently available is that of Mrozowski & Hayes (1978), which covers the west Pacific basins from the

TABLE 2 Mass-age distribution of pelagic sediments

1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
0–4	4–9	9–20	20–35	35–52	52–65	65–80	80–95	95–110	110–125	125–140	140–160	160–180		basement age, Ma
12.5	18.0	29.7	35.1	33.4	27.0	32.3	27.9	24.8	13.9	15.4	8.3	3.4	281.7	area, 10^6 Km^2
			accummulation in units of 10^{14} metric tons										total	sediment age, Ma
1.5	4.3	7.1	8.4	8.0	6.5	7.8	6.7	6.0	4.5	3.7	2.0	0.8	67.3	0–4
	2.7	8.9	10.5	10.0	8.1	9.7	8.4	7.4	4.2	4.6	2.5	1.0	78.0	4–9
		9.8	23.2	22.0	17.8	21.3	18.4	16.4	9.2	10.2	5.5	2.2	156.0	9–20
			15.8	30.1	24.3	29.1	25.1	22.3	12.5	13.9	7.5	3.1	183.7	20–35
				17.1	27.5	32.9	28.5	25.3	14.2	15.7	8.5	3.5	173.2	35–52
					10.6	25.2	21.8	19.3	10.8	12.0	6.5	2.7	108.9	52–65
												Tertiary	767.1	
						14.6	25.1	22.3	12.5	13.9	7.5	3.1	99.0	65–80
							12.6	22.3	12.5	13.9	7.5	3.1	71.9	80–95
								11.2	12.5	13.9	7.5	3.1	48.2	95–110
									6.3	13.9	7.5	3.1	30.8	110–125
										7.0	7.5	3.1	17.6	125–140
												Cretaceous	267.5	
											5.0	4.1	9.1	140–160
												2.1	2.1	160–180
												Jurassic	11.2	
												grand total	1046	

Based on an assumed mean global accumulation rate of 0.6 g.cm^{-2} per 1000 years (Lisitzin 1972; Worsley & Davies 1979), and area-age distribution of ocean basement according to Sclater *et al.* (1980). Each age province (top row) is represented by a column (1–13) showing the mass-age distribution of its sediments. For instance, basement between 20 and 35 Ma old (Col. 4) will be covered by sediments of the first four age groups given in Col. 15, whose respective masses are products of time interval, area and accumulation rate. (This product is halved for the first time interval, during which the province is being created and has a mean exposure to sedimentation of only half the interval.) The masses for the different age groups are summed in Col. 14, where their distribution between Jurassic, Cretaceous and Tertiary Systems is indicated.

TABLE 3 Mass-age distribution of surviving Phanerozoic sediments

1	2	3	4	5	6	7	8	9	10	11	12
			surviving mass, units of 10^{15} metric tons								
System	duration of Period, Ma	mean age, Ma	continents	passive margins	marginal basins	pelagic	abyssal plains	deep-sea fans	total	survival rate, S 10^9 t/y	log S
Tert-P1	66	33	151.8	140.8	121.8	76.7	73	13.2	576.9	8.74	9.94
Cretaceous	66	99	225.6	114.3	22.2	26.7	39	2.2	430.0	6.51	9.81
Jurassic	53	159	161.8	65.3	—	1.1	—	—	228.2	4.31	9.63
Triassic	50	210	124.0	16.8	—	—	—	—	140.8	2.82	9.45
Permian	45	258	108.8	—	—	—	—	—	108.8	2.42	9.38
Carbonif	65	313	137.7	—	—	—	—	—	137.7	2.12	9.33
Devonian	55	373	181.6	—	—	—	—	—	181.6	3.30	9.52
Silurian	35	418	67.8	—	—	—	—	—	67.8	1.94	9.29
Ordovician	55	463	97.8	—	—	—	—	—	97.8	1.78	9.25
Cambrian	80	530	157.4	—	—	—	—	—	157.4	1.97	9.29
			1414.3	336.8	144.0	104.5	112	15.4	2127.0		

Col. (2) from Afanasseyev & Zykov (1975); (4) Ronov *et al.* (1980); (5) this paper, Table 1; (6) this paper (see text); (7) this paper, Table 2; (9) this paper (see text); (11) surviving mass of System divided by duration of corresponding Period.

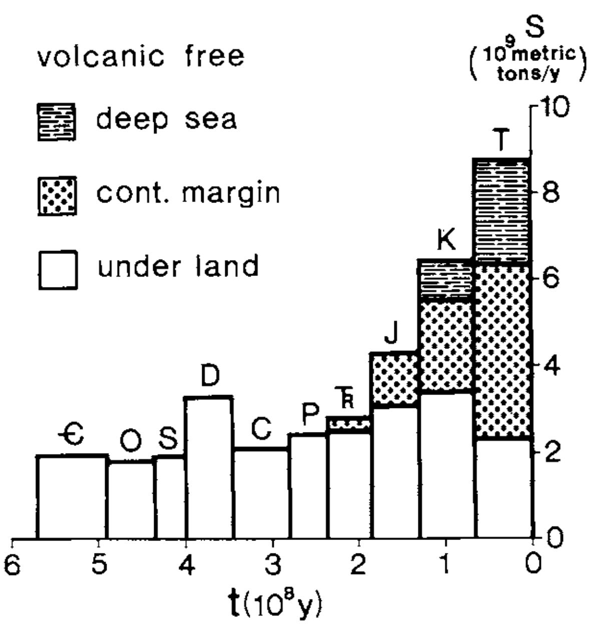

FIG. 1. Mass-age distribution of the Phanerozoic Systems. S, survival rate (mass of System divided by duration of Period).

Sea of Japan to the Sunda arc and includes part of the Andaman Sea. This survey, covering 6.3 million km^2, represents nearly a quarter of the world's marginal basins in the form of an isopach map, which for the present study was transferred onto an equal-area projection and analysed planimetrically. The resulting mean sediment thickness was 2.43 km. Perusal of world-wide data on marginal basins (e.g. Biju-Duval *et al.* 1979; Ladd & Watkins 1979) suggests that, though highly variable, the thicknesses in basins such as the Mediterranean and the Caribbean on the one hand, and in the narrow margins of the east Pacific on the other, fairly bracket the west Pacific average, which is adopted here for the whole area. Assuming a mean density (averaged from DSDP reports) of 2.2 g.cm^{-3}, the resulting sediment mass is $26.9 \times 2.43 \times 2.2 \times 10^{15}$ metric tons, or 144×10^{15} metric tons. This is distributed according to the area-age distribution of the basins (Sclater *et al.* 1980) assuming a constant rate of deposition per unit area, resulting in the allocation of 122×10^{15} tons to the Tertiary and 22×10^{15} tons to the Cretaceous. The Cretaceous may be overestimated by this method; but as its allocation is small anyway, the error is unlikely to be large. These results are shown in Table 3, Column 6.

The deep sea floor

Three types of accumulation need to be recognized here: (1) pelagic sediments; (2) non-pelagic sediments (mainly turbidites) on the proximal parts of the abyssal plains lying off passive margins, and (3) deep-sea fans such as the Bengal fan, the Indus cone and other large deltaic masses that have prograded from the continents onto the deep sea floor.

Pelagic sediments

Ideally, the mass-age distribution of pelagic sediments should be revealed by the numerous boreholes of the Deep Sea Drilling Project (DSDP Initial Reports, 1969–1982). In fact, the number of holes reaching basement is inadequate to give a reliable picture, granted the vagaries of sediment transport and basement topography. On the other hand, reliable data are available on the net accumulation rate of pelagic sediment over the past 50 Ma (Lisitzin 1972; Worsley & Davies 1979), and on the area-age distribution of oceanic basement (Sclater *et al.* 1980); and if the 50 Ma rate can be extrapolated to 180 Ma (age of the oldest known basement), a simple model can be made for finding the mass-age distribution. Such a model is embodied and explained in Table 2. It may err in exaggerating the masses of Jurassic and Lower Cretaceous sediment, formed when there were fewer pelagic foraminifera and known to exhibit unconformities; but granted the relatively small masses attributed to these parts of the column, the error is not likely to be large. The masses found in this way are reproduced in Table 3, Column 7.

Non-pelagic sediments of the abyssal plains

These make up a significant part of the deep-sea sediment mass. Sixteen of the seismic profiles studied for estimation of the marginal sediments afford data on thicknesses under the proximal abyssal plain, and fourteen of them show the proportions of Cretaceous and Tertiary sediments. Information about thicknesses, lithology and stratigraphy farther offshore is available in the DSDP reports. From the total thickness can be subtracted the thickness of pelagic sediment predicted by the pelagic sedimentation model, assuming a mean density of 2.0 $g.cm^{-2}$. From this it appears that about 35 million km^2 of the abyssal plains are underlain by an average thickness of 1.6 km of non-pelagic sediment, 65% of it Tertiary and 35% Cretaceous. Translated into mass this comes to 73×10^{15} metric tons of Tertiary sediment and 39×10^{15} Cretaceous. These figures are entered in Table 3, Column 8.

Deep-sea fans

The largest deep-sea fans are the Bengal fan and the Indus cone. Other bodies of terrigenous origin prograded on the deep sea floor, such as the Mississippi, Amazon, Niger, Congo and Orange River deltas, probably will not greatly affect the mass-age distribution attributed to this class of sediment from its occurrences in the north Indian Ocean. An isopach map of the Bengal fan by Curray *et al.* (in press) indicates a total volume of 3.6 million km^3; the Indus cone is a little smaller. Considering the sources of these two accumulations and the age of the oceanic basement underlying them, they must be largely of Tertiary age. The total volume world-wide is here estimated at 7 million km^3, 86% Tertiary and 14% Cretaceous. At a mean density of 2.2. $g.cm^{-3}$, this is equivalent to 13.2×10^{15} metric tons Tertiary and 2.2×10^{15} Cretaceous. These masses are shown in Table 3, Column 9.

The mass-age distribution

The overall mass-age distribution is shown in Table 3, Column 10. Column 11 gives the 'survival rate' of each System (surviving mass divided by the duration of the corresponding Period), and this is displayed graphically in Fig. 1, where it is broken down into subterranean (or cratonic), marginal and deep-sea components. The total Phanerozoic mass is about 2.1×10^{18} metric tons, and about 30% of it consists of sediment presently under the sea (on continental margins and the deep sea floor). The survival rate falls off steeply from Tertiary to Carboniferous; beyond that the tail of the distribution holds up and is rather irregular.

Former estimates of the mass-age distribution have shown the same general pattern as the present one. A plausible interpretation was given by Garrels & Mackenzie (1971a), who postulated a constant sediment mass and sedimentation rate with selective preservation in post-Devonian time of the Lower Palaeozoic Systems through entombment in the Palaeozoic orogenic belts; in other words, a steady-state model with age-dependent probability of destruction. Later studies (Garrels & Mackenzie 1971b; Veizer & Jansen 1979; Walker 1981; Dacey & Lerman, in press) have been concerned with identifying a secular component in the sediment flux or with avoiding the deterministic limitations of the earlier models (e.g. Gregor 1970). These valuable contributions have offered fresh viewpoints from which to look at the sedimentary cycle and more rigorous methods for analysing it; however (as Dacey & Lerman point out), the mass-age estimates available till now (including the present one) are too uneven and too uncertain to allow a unique choice of model. Granted the age of the earth, it is reasonable to assume that any secular change in the sediment mass must be slow in comparison to recycling; and if the Lower Palaeozoic Systems are set aside as having a partly 'fossilized' distribution (Garrels & Mackenzie 1971a; and see further below), the distribution over the past 300 Ma can be interpreted with as much precision as the data allow by using a constant-mass model with probability of destruction (per unit mass) evenly distributed in time. The rate of destruction of sediment (by erosion, metamorphism, subduction) is equal to the rate of sedimentation, and for a given formation of mass S, the rate of destruction is:

$$\frac{dS}{dt} = -\,k\cdot S \qquad (1)$$

so that the mass surviving at any time t after deposition of an initial mass S_o will be

$$S_t = S_o\cdot e^{-kt} \qquad (2)$$

or, in logarithmic terms,

$$\ln S_t = \ln S_o - kt \qquad (3)$$

If S_o is set equal to the mass deposited in a year, a plot of log S against age (t) should approximate a straight line whose intercept at $t = 0$ gives the mean sedimentation rate and whose slope is equal to $-$ k/2.3. Such a plot for the Phanerozoic is shown in Fig. 2. Neglecting the Lower Palaeozoic Systems (see above), the best fit is given by

$$\log S_t = 10.01 - 0.24t \qquad (4)$$

with t in hundreds of Ma. The correlation between log S and t ($r = -0.98$) is significant at the 95% confidence level. The intercept at log $S = 10.01 \pm 0.09$ (2 standard errors) indicates a mean sedimentation rate between 0.8×10^{10} and 1.3×10^{10} metric tons per year for the 300 Ma period, with a most likely value of 1.0×10^{10}; and the slope of $-$ 0.24 corresponds to k = 0.54 per 100 Ma, indicating a half-life for the post-Devonian mass of about 130 Ma. The rate agrees well with the immediately pre-human rate found by other methods (Judson 1968; Gregor 1970), but the agreement is probably fortuitous, because the instantaneous rate at any time is

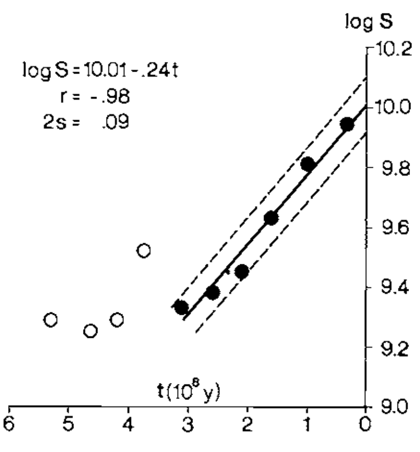

FIG. 2. Plot of log S against age, t.

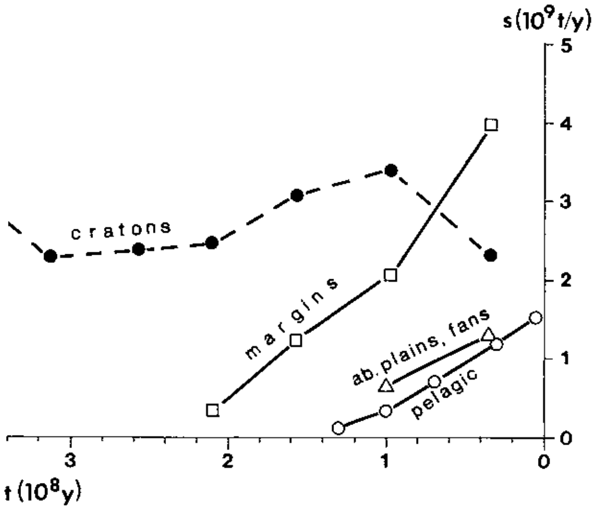

FIG. 3. Mass-age distributions for cratonic (subterranean), marginal and deep-sea sediments, Carboniferous through Tertiary.

unlikely to be near the average taken over several geologic Periods. The large fluctuations in preservation rate that become visible even on the scale of Ages and Epochs (see for example Livingstone 1963; Sloss 1979) show how widely the sedimentation rate must vary over short intervals, and this has to be borne in mind when making comparisons. For instance, the 'present-day' rate found by Judson (1968), 2.4×10^{10} metric tons a year, is large compared to the 'pre-human' rates of *c.* 1.0×10^{10} determined in uninhabited river basins or from chemical arguments; and there is no doubting the serious losses of soil caused by human activities such as deforestation and tillage. But without knowing the natural range of variation of the erosion rate, and the rate at which it can vary, we cannot tell how much of the difference has to be laid at our own door.

The foregoing is a reasonable interpretation of the total surviving mass as it is distributed in time, without regard to the behaviour of its different components. It is evident from Fig. 1, however, that the rapid decrease in survival rate over the first 300 Ma is almost entirely due to decay of the submarine (marginal and deep-sea) sediments. The subterranean (cratonic) mass shows no regular decrease with age on a time-scale of this length, but only a low-amplitude fluctuation. This pattern evidently expresses the durability of sediments incorporated into the cratonic interior, as recognized by Garrels & Mackenzie (1971a) in the case of the Lower Palaeozoic Systems. (As shown by Ronov *et al.* 1980, the survival rate of the cratonic sediments drops significantly after about 500 Ma and tails off slowly thereafter.)

A breakdown of the post-Devonian mass-age distribution into its different components, derived from Tables 2 and 3, is given in Fig. 3. Assuming uniform rates of pelagic sedimentation, sea-floor spreading and subduction, the survival rate of deep-sea sediments should be a linear function of age. Table 2 is based on the first of these assumptions, but assumes nothing about spreading or subduction rates. In fact, when the entries in Column 14 are divided by the corresponding time intervals in Column 15 and the quotients (survival rates, S) are plotted against median age, the resulting distribution is approximately linear with $S = 1.6 - 1.2t$ ($r = -0.98$), where S is in units of 10^9 metric tons per year and t in units of 10^8 years. A reduced version of this plot, showing five of the 13 points used, can be seen in the lower right corner of Fig. 3. The correlation coefficient r is significant at the 99.9% level. Submarine marginal sediments are removed to the crystalline rock reservoir by metamorphism and to the land by uplift (orogenesis, epeirogenesis). The majority leave by the latter route, and those exposed on mountain belts are quickly returned to the continental margins by erosion. Unfortunately, too few data points are available for a reliable analysis of the mass-age distribution of the marginal sediments. The possibility that it, too, is linear cannot be ruled out. The four points available from Table 3 (Columns 5 and 6) fit $S = 4.4 - 2.0t$ (same units as before), with $r = -0.99$, significant at the 99% level; the best exponential fit is $\log S = 9.85 - 0.57t$ ($r = -0.97$, significant at the 95% level). The mass-age distribution for abyssal plains and deep-sea fans is represented in Fig. 3 for the sake of completeness. As there are only two points, nothing can be said about the form of this distribution, whose contribution to the overall pattern is, however, not negligible.

It is thus apparent that the mass-age distribution of the Phanerozoic sediments is the sum of a number of discrete distributions, each with its own signature, about which there is insufficient information at present to characterize any one of them (with the possible exception of the deep-sea sediments) unambiguously. Before the overall distribution can be interpreted with confidence more work will have to be done on its different components, especially on the sediments of the continental margins, which dominate the later Phanerozoic picture.

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Sr isotopic constraints for the sedimentation rate of deep sea red clays in the southern Pacific Ocean

Norbert Clauer and Michel Hoffert

SUMMARY: Sedimentological, mineralogical, chemical and Rb-Sr isotopic studies were made on Fe-smectites (nontronites) from deep sea red clays taken from two cores located 100 km apart in the southern Pacific Ocean. They show that: (1) these sediments are autbigenic, (2) they contain variable but few amounts of volcanic-derived chemical components, and (3) they were transported from some place in the ocean and deposited at the present-day site. The transportation, which could occur at different periods, was likely induced by the submarine topography and bottom currents.

These observations necessitate a reconsideration of the traditional concept of sedimentation rate for the red clay facies. Although exhibiting considerable similarities, these sediments may correspond to different stages of a submarine sedimentary cycle. A given horizon can therefore reflect a submarine pedological event, a sudden and massive accumulation, or a very low sedimentation of material formed elsewhere. Genetic information is necessary to determine the local sedimentation rates of red clay, and the inconstancy of the sedimentation prevents any generalization on an ocean-wide scale.

Sedimentation rates represent an important parameter in the reconstruction of the geodynamic evolution of sedimentary basins. Since the pioneer studies of Ewing *et al.* (1940), they can be reasonably estimated from sedimentary thickness measurements based on seismic data. In addition sedimentation rates may also be estimated through the use of the ^{230}Th or the ^{10}Be methods (Tanaka & Inoue 1979; Brown *et al.* 1981; Krishnaswami *et al.*. 1982).

Sedimentation rates must, however, be used with caution for modelling purposes because they depend upon various parameters such as: (1) the nature of the sediments, which varies with the distance to the nearby continent and with the water depth and (2) the configuration of the sedimentation basin, which may be shallow, epicontinental and with little, if any, submarine topography or deep, oceanic and with a significant topography. Sedimentation rates may also be influenced by local parameters such as bottom currents, which can produce turbidites (Heezen & Ewing 1952) or gaps in the sedimentary sequence (Arrhenius 1952).

In the present paper, sedimentological, mineralogical, geochemical and Rb-Sr isotopic data on deep sea red clays from the southern Pacific Ocean are used to provide additional information on the sedimentary environments of the Pacific Ocean, and to allow reconsideration of the sedimentation rate for these deep sea red clays.

The concept of deep sea red clays

The first sediment called 'red clay' was sampled in the South Pacific at a depth of 4755 m and was described by Thompson (1874). Ten years later, Murray & Renard (1884) proposed the first classification of deep oceanic sediments including the red clay facies. They considered red clays as occurring in regions characterized by the *absence of detrital minerals* such as in the central Pacific Ocean, and consisting of minor amounts of clay minerals mixed up together with zeolites and Fe-Mn oxides. Later, Murray & Renard (1891) were even more explicit: they ascribed the origin of the red clays to the weathering of volcanic material deposited in deep oceans, at great distances from continents to avoid any detrital supply, and in sufficiently deep water to favour the dissolution of calcareous organisms.

Murray & Renard's suggestion (1891) remained unchallenged until Revelle (1936) who basically did not reconsider their classification. However, taking in account a certain number of 'weaknesses', he proposed modifications which, nevertheless, lead to a completely different interpretation for the origin of most oceanic clays. A continental origin was favoured, while clay minerals neoformed from volcanic rocks were considered as scarce. Thus, the term 'red clay' previously defined as a specific sediment mainly composed of neoformed clays in oceanic environments, was used especially by Sverdrupt *et al.* (1942) to designate *any* fine pelagic sediment.

Between 1944 and 1968, numerous studies led to a progressive understanding of the structural evolution of the oceans and to the sea floor spreading hypothesis. Concerning the mineralogy of the deep sea sediments, a continental origin of clay minerals remains generally admitted; those which are neoformed are rarely demonstrated as they are considered as scarce and *exclusively* located within volcanic active regions. Thus, the term 'red clay' was progressively given up, as it entirely lost its original meaning and, therefore, became ambiguous.

Since 1968 new sampling possibilities with the drilling oceanographic vessel 'Glomar Challenger', as well as new *in situ* investigations with diving vessels have allowed a systematic exploration of the oceans. In the related publications the term 'red clays' is used again in a general sense to describe sediments of different origins (i.e. Kukal 1971; Riley & Chester 1971; Lisitzin 1972; Leclaire *et al.* 1976, 1977; Giannesini 1977; Jacobs 1978; Berger 1978).

Recently, Hoffert (1980), among others, gave a detailed account of the mineralogy and chemistry of red clays, together with their geological environment in order to eluci-

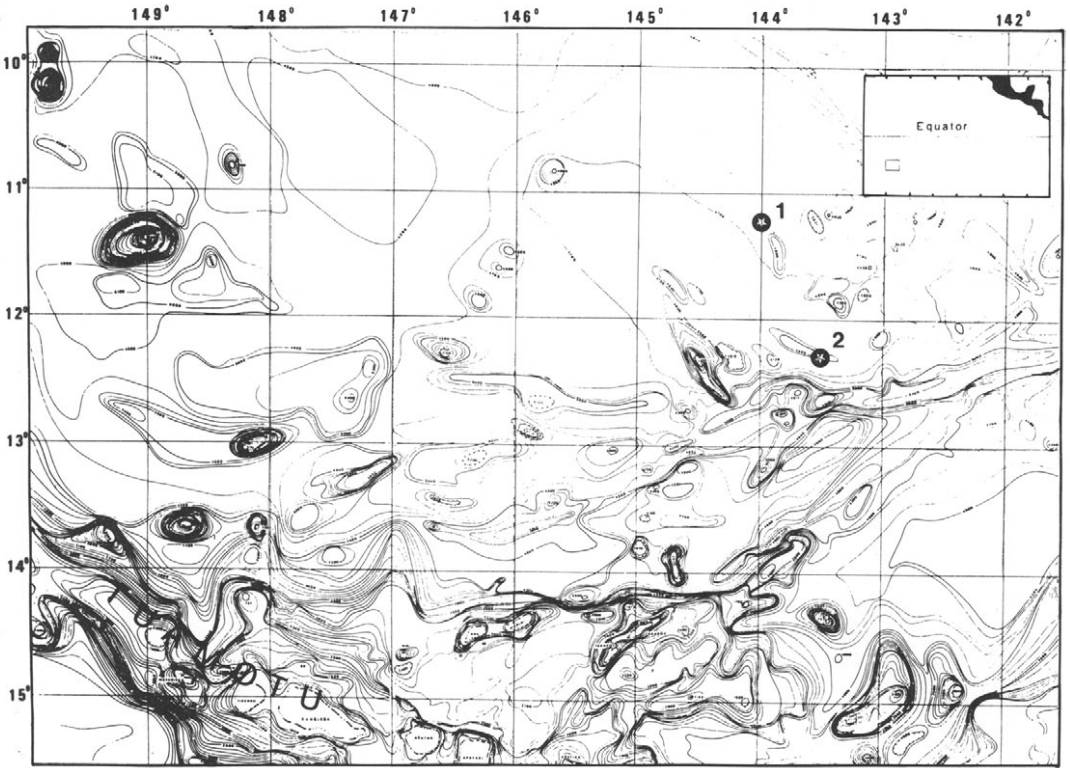

FIG. 1. Location of the two cores in the southern Pacific Ocean. The insert shows the location of the map.

date their origin and distribution in the central and eastern Pacific Ocean. An important conclusion of his study was to underscore that these sediments, which occur in the quietest and deepest zones of the ocean with only scarce organic remains, obviously represent material fashioned at a different place than that of deposition. They, therefore, consist of *authigenic* components at the ocean scale but were *reworked* with respect to the place of deposition and then reequilibrated chemically with their new sedimentary environment. A consequence of this observation is the consideration and recognition of a submarine sedimentary cycle in the oceans including erosion-weathering, dissolution-transport and sedimentation-diagenesis stages (Hoffert *et al.* 1978; Hoffert 1980).

The purpose of the present paper is not to generalize a new meaning of the term 'red clay' or to modify the present-day sense. The term is used here to describe sediments which seem authigenic, but may have been transported to and also chemically modified in their present environment. It corresponds more to Murray & Renard's definition (1884, 1891) than to the contentions proposed during the last decade.

Sediment location and description

The present study deals with two cores sampled during the Transpac I oceanographic cruise organized by the Centre National pour l'Exploitation des Océans in the south Pacific Ocean at 11°28.6'S; 144°00.5'W and at 12°32.0'S; 143°52.2'W, respectively (Fig. 1). These sites are located about a 100 km apart north of the Marquisas fracture zone and west of the Marquisas Islands in a region characterized by important troughs surrounded by volcanic seamounts (Fig. 2). The water depth is 4725 m for the first drill site and 5029 m for the second; the sediments lie below the carbonate dissolution limit estimated at a depth of 4200 m in this region (Berger *et al.* 1976). Furthermore, strong bottom currents move in this part of the Pacific Ocean from west to east (Pautot & Melguen 1975).

Both cores are about 380 cm long; they represent the upper part of a sedimentary horizon which is about 60 m thick (Hoffert 1980) and is topped at the water-sediment interface by concentrations of metalliferous oxides. The sediment of both cores is a dark-brown, fine, homogeneous ooze; no components of continental origin could be found. It contains,

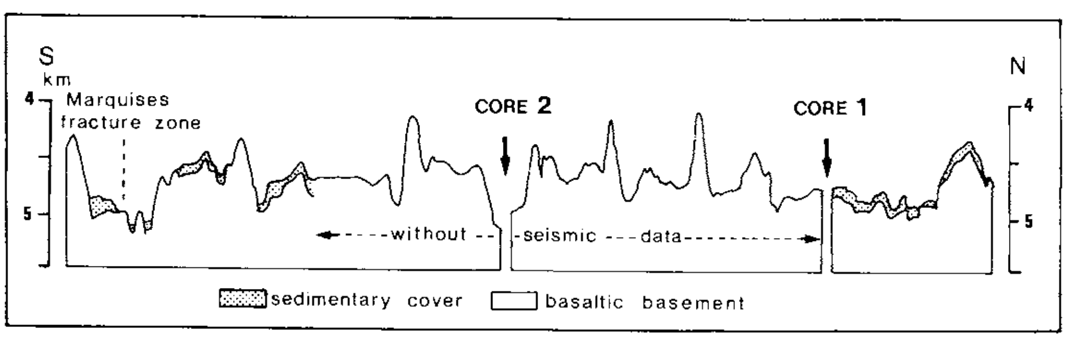

FIG. 2. Bathymetric and seismic profiles near the sampling stations.

however, a small amount of coarse material (> 63 μm): about 1.7% at the top of core 1,decreasing progressively downward to about 0.3%; in core 2, the amount of the coarse fraction is 4.6% at surface and varies between 0.1 and 0.5% lower down. The amount of calcium carbonate is small: in core 1 it comprises 8% of the whole sediment at the surface and 3% lower down; in core 2 it ranges from 5% at the top to nothing at the bottom.

The age of these sediments could not be determined. No fossils were observed below the seven uppermost centimetres of core 1 which contain scarce radiolarian of Plio-Quaternary age (Schaaf, pers. com.).

Analytical methods

The mineralogy of the ooze was determined by microscope examination, XRD, DTA and laser granulometric analyses (Cornillault 1974). The clay minerals concentrated in the size fraction smaller than 2 μm were separated in distilled water using the classical sedimentation technique without any prior dissagregation. Morphology of the clays was studied using electron microscopy, and the chemical data were obtained using spark spectrometry.

The Sr abundances of the clay minerals and their $^{87}Sr/^{86}Sr$ ratios were measured on a mass-spectrometer with a 30 cm radius and a 60° deviation angle, after sample dissolution in $HF + HClO_4$ and Sr purification by resin chromatography. The Rb abundances were determined on a different mass-spectrometer with a 15 cm radius and a 90° deviation angle. Several clay fractions were leached with 1N HCl following a technique proposed by Clauer (1982) and separate measurements were made on the leachates (with subscript L), the residues after leaching (with subscript R) and the untreated sample (with subscript U). The small differences between measured Rb and Sr contents of the untreated samples and the corresponding calculated amounts points out the reliability of these leaching experiments. During the course of the study, four independent measurements of the standard NBS-987 gave an average value of 0.70999 ± 0.00017 (2σ) for the $^{87}Sr/^{86}Sr$ ratio. The individual error of each $^{87}Sr/^{86}Sr$ ratio in Table I is quoted at the 2 σ of the mean level; the error of the Rb/Sr ratios is estimated as better than ± 1.0%. The values were corrected with the usual constants $^{85}Rb/^{87}Rb = 2.591$ and $^{86}Sr/^{88}Sr = 0.1194$; each $^{87}Sr/^{86}Sr$ ratio was also adjusted to 0.71022 which is the value generally admitted for the $^{87}Sr/^{86}Sr$ ratio of the NBS-987 standard. Further details of the analytical procedure are available in Clauer (1976).

Results

The results obtained on *core 1* have already been published (Hoffert 1980; Clauer *et al.* 1982); they are summarized with the new data obtained in this study (Figs 3 and 4).

The ooze mainly consists of fine particles: 70% of the material is smaller than 8 μm and the clay fraction (< 2 μm) represents 25 to 30% of the volume of the whole-rocks. Palagonite is the major component of the coarse fraction (> 63 μm) at the top of the core, while phillipsite predominates to a depth of 160 cm. Below this limit, phillipsite is replaced by 'brown aggregates' which already have been described by Revelle (1944). These aggregates are composed of smectites with minor amounts of phillipsite, quartz and Fe-oxides. The coarse fraction, in addition to these components, may also contain Fe-Mn micronodules with accessory amounts of organic debris and scarce crystals of dolomite, ankerite or quartz. The water content calculated with the following formula

$$\frac{\text{weight of wet sediment} - \text{weight of dry sediment}}{\text{weight of dry sediment}} \times 100$$

is about 104% of the whole sediment in the uppermost 10 cm of the core. It remains constant at 80% until a depth of 140 cm, then decreases quickly to 50%.

The clay fraction consists exclusively of smectites, with the exception of the surface fraction which also contains traces of illite and kaolinite. The DTA curves indicate that the smectites are very similar throughout the core; their chemical data are typical of Fe-rich minerals also called nontronites. The particles of 1–2 μm appear circular on the electron microscope photographs with sharp borders often rolled up and/or prolonged by delicate fibre-like prolongations of 0.5 μm length which seem to grow from the particles (Fig. 3 in Clauer *et al.* 1982). The chemical compositions of these smectites vary with depth; they show a decrease in Mg, Ca, Ti and Na contents and an increase in Fe and K contents (Fig. 3). These changes are interpreted as being results of diagenetic modifications.

TABLE 1 Rb-Sr isotopic data for the nontronites (< 2 μm) of core 2.

Samples	% of each fraction	Depth in core (cm)	Rb measured	Rb calculated	Sr measured	Sr calculated	$^{87}Sr^{+}/Rb$ (10^{-3})	$^{87}Rb/^{86}Sr$ measured	$^{87}Rb/^{86}Sr$ calculated	$^{87}Sr/^{86}Sr$ measured ($\pm 2\sigma/\sqrt{N} \times 10^{-5}$)	$^{87}Sr/^{86}Sr$ adjusted
2 – 1 U	—	0– 2 cm	43.6	—	524	—	0	0.241	—	0.70855 ± 10	0.70878
L	13.5		12.1	1.6	2790	377	—	0.013	—	0.70889 ± 10	0.70912
R	86.5		48.4	41.9	150.2	129.9	—	0.933	—	0.70695 ± 14	0.70718
L + R	100		—	43.5	—	506.9	—	—	0.249	—	—
2 – 2 U	—	110–117 cm	49.8	—	460	—	0	0.341	—	0.70844 ± 8	0.70867
2 – 3 U	—	170–175 cm	47.0	—	458	—	0	0.297	—	0.70846 ± 7	0.70869
2 – 4 U	—	204–209 cm	48.8	—	452	—	0	0.313	—	0.70842 ± 12	0.70865
2 – 6 U	—	305–310 cm	49.2	—	458	—	0	0.312	—	0.70841 ± 9	0.70864
2 – 7 U	—	372–378 cm	54.3	—	446	—	0	0.353	—	0.70850 ± 9	0.70873
L	10.5		19.7	2.1	2936	308	—	0.019	—	0.70888 ± 3	0.70911
R	89.5		57.5	51.5	152.9	136.8	—	1.091	—	0.70771 ± 8	0.70794
L + R	100		—	53.6	—	444.8	—	—	0.349	—	—

$^{87}Sr/^{86}Sr$ adjusted = measured $^{87}Sr/^{86}Sr$ adjusted to 0.71022 for the $^{87}Sr/^{86}Sr$ ratio of the NBS 987 standard. U = untreated; L = leachate; R = residue. $^{87}Sr^{+}$ = radiogenic ^{87}Sr.

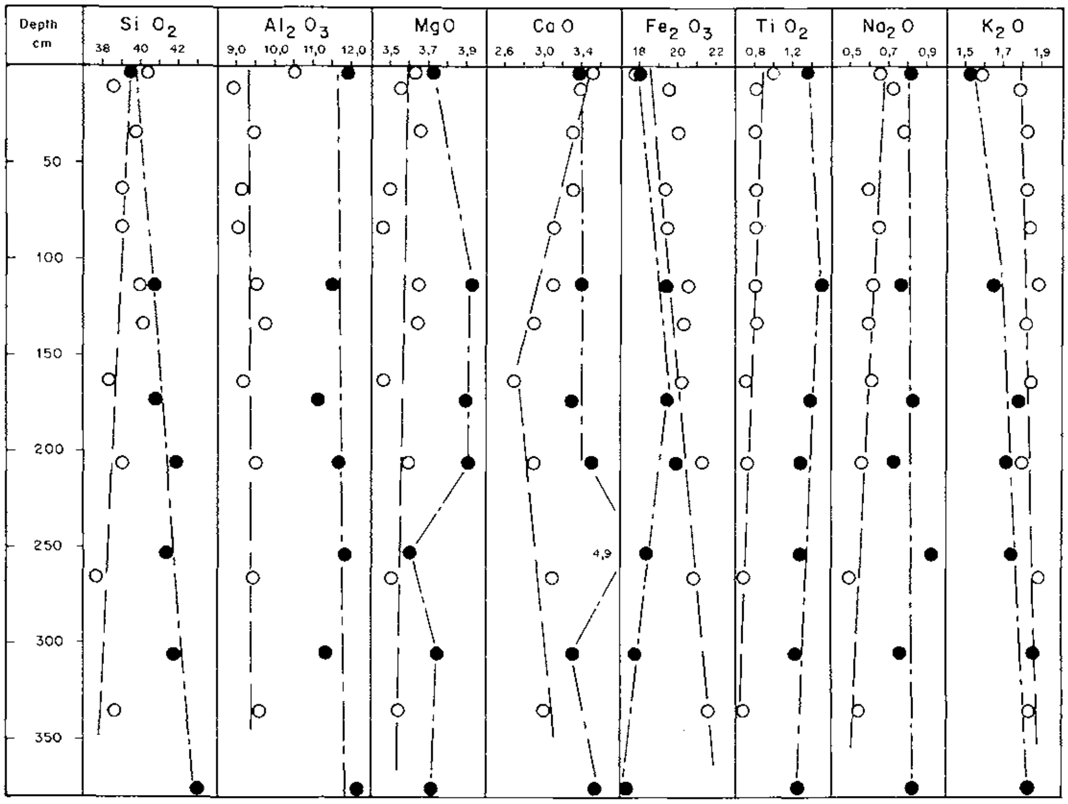

FIG. 3. Chemical data of the nontronites (< 2 μm) of core 1 (open circles) and core 2 (full circles) vs. depth. Amounts are given in percent.

The Sr isotopic determinations made on the clay minerals and their comparison with similar data obtained earlier on the associated interstitial waters (Clauer *et al.* 1975) reveal that the smectites at the sediment surface are not entirely in isotopic equilibrium with their environment. A volcanic influence during the formation of these minerals has to be taken in account to explain the relative depletion of the $^{87}Sr/^{86}Sr$ ratio of the clays when compared to that of the fluids. This interpretation is supported by the presence of palagonite in the coarse fraction. The Sr isotopic study also shows the migration of radiogenic ^{87}Sr from the silicate lattices into the nearby interstitial waters. This migration, which starts very early in the sediment, is interpreted together with the presence of brown aggregates in the coarse fraction and the progressive chemical variation in the clay from top to the bottom of the core, as a consequence of diagenetic transformations of the smectites after deposition. The morphology of the clay particles also argues in favour of an evolutionary history in three stages: (1) formation of clay minerals with sharp borders in a volcanic environment (by weathering of submarine basalts or during hydrothermal activity near mid-oceanic ridges) different from the present-day environment; (2) transportation by bottom currents to the present-day site where they are consequently *inherited*; and (3) re-equilibration with their new environment by modification of their chemical composition, readjustment of their Sr-isotopic composition, and formation of delicate fibre-like prolongations at their borders.

The sediments of *core 2* are also very fine; 70% of the volume of the whole-rocks consist of particles smaller than 2 μm which often agglomerate into aggregats of 2–8 μm. Palagonite predominates in the coarse fraction (> 63 μm); at the surface it represents 80% of the volume of this fraction then it decreases progressively to 10% near the bottom of the core. Micronodules, brown aggregates and phillipsite are scarce in the upper 90 cm; below, the brown aggregates become more abundant until they represent 80% of the coarse fraction. Some fragments of organisms were observed in the surface level, together with debris of shark teeth; a few grains of quartz and dolomite could also be recognized. In summary, palagonite is more abundant in core 2 than in core 1; it requires a greater volcanic influence. Phillipsite, which is scarce in the coarse fraction, is rather abundant in the silty fraction. The water content is relatively low in the upper 152 cm of the core ranging between 60 and 85% of the weight of the sediment. At greater depth, it increases to as much as 160–200% of the weight of the bulk sediment.

The mineralogical composition of the clay fraction in core 2 is identical to that of core 1 located a 100 km apart. This similarity is either the result of an unlikely chance or an argument of a mineralogical uniformity on a regional scale. The smectites are Fe-rich with occasional impurities of goethite, quartz and volcanic glass. Their chemical compositions (Table 2) show, from top to bottom, an increase of the Si and K contents, a constancy of the Al, Ti and Na contents, and an irregular variation of the Mg, Ca and Fe contents.

The clay fraction from surface sediment (sample 2–1) has an $^{87}Sr/^{86}Sr$ ratio of 0.70878 ± 0.00010, which is slightly but significantly lower than that of the marine Sr which is 0.70906 ± 0.00003 (2σ) (compilation in Faure 1982). This difference is even larger if one considers the $^{87}Sr/^{86}Sr$ ratio of 0.70718 ± 0.00014 for the residue after 1 N HCl leaching. This leaching experiment removes the Sr which is adsorbed on the mineral surfaces or trapped in neoformed marine components; the $^{87}Sr/^{86}Sr$ ratio of the leachate should therefore be identical to that of the marine Sr. The low $^{87}Sr/^{86}Sr$ ratio of the smectites from surface sediment, relative to sea-water, corroborates the presence of a volcanic derived component already found in

TABLE 2 Chemical data for the nontronites (< 2 μm) of core 2.

Samples	Depth in core	SiO_2	Al_2O_3	MgO	CaO	Fe_2O_3	Mn_3O_4	TiO_2	Na_2O	K_2O	BaO	Loss at 1000°C	Total
2–1	0– 2 cm	39.4	11.9	3.72	3.4	17.9	5.66	1.33	0.82	1.53	0.23	10.45	96.39
2–2	110–117 cm	40.7	11.5	3.93	3.4	19.6	5.07	1.51	0.76	1.65	0.21	9.73	98.06
2–3	170–175 cm	40.6	11.1	3.89	3.3	19.4	5.10	1.39	0.83	1.79	0.17	9.69	97.20
2–4	204–209 cm	41.9	11.7	3.90	3.5	19.9	5.10	1.27	0.72	1.72	0.18	9.63	99.45
2–5	250–255 cm	41.2	11.8	3.60	4.9	18.3	5.69	1.28	0.91	1.74	0.25	9.41	98.98
2–6	305–310 cm	41.7	11.3	3.74	3.3	17.8	5.03	1.23	0.76	1.87	0.20	9.44	96.39
2–7	372–378 cm	42.8	12.0	3.70	3.5	17.1	4.50	1.23	0.84	1.84	0.19	9.17	96.88

The amounts are in percent

that of core 1 (Clauer *et al.* 1982). As the $^{87}Sr/^{86}Sr$ ratio of this clay residue is lower than that of the corresponding fraction of core 1 (0.70858 ± 0.00012), the amount of volcanic derived material in the former is slightly greater; this is in accord with the greater amount of palagonite in the coarse fraction.

The $^{87}Sr/^{86}Sr$ ratio of the nontronites varies differently with depth in the two cores. In core 1 the ratio $^{87}S/^{86}Sr$ increases with depth, whereas it remains nearly constant between 0.70864 ± 0.00009 and 0.70878 ± 0.00010 along the entire length of core 2 (Fig. 4). Apparently, all the clays of core 2 have the same age as no radiogenic ^{87}Sr accumulation is detectable toward the bottom of the core. The amount of Rb increases only slightly in the nontronites from the top to the bottom of core 2, while the Sr content first decreases from the surface to a depth of 110 cm and then remains nearly constant (Table 1 and Fig. 4).

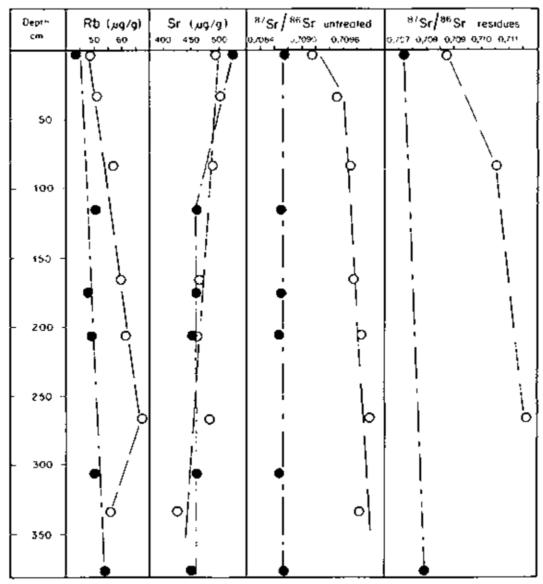

FIG. 4. Rb and Sr concentrations, $^{87}Sr/^{86}Sr$ ratios of the nontronites before 1N HCl leaching (untreated) and after (residues) of core 1 (open circles) and core 2 (full circles).

Discussion

The similarity, within the analytical errors, of the $^{87}Sr/^{86}Sr$ ratios of six nontronites sampled along the 380 cm of core 2, compared with a variation of 65.10^{-5} units for the corresponding clay minerals of core 1 located a 100 km to the north, requires that all clay minerals of core 2 have the same age, since their $^{87}Sr/^{86}Sr$ ratios, which are normally subordinate to their age, do not increase toward the bottom of the core like in example 1. This result has two other consequences: (1) the sediment was deposited recently and rapidly, as a low sedimentation rate should have produced an accumulation of radiogenic ^{87}Sr which is not detectable in the clay minerals, and (2) it was probably from the same source and mixed up during transportation. The predominance of palagonite in the coarse fraction (> 63 μm) of core 2 represents an argument in favour of a recent deposition. In core 1, it is abundant in the upper-most 7 cm of core 1, which have a Plio-Quaternary age; below it is replaced by phillipsite. Therefore, the time-span since deposition was certainly too short in core 2 to allow its recrystallization into zeolites. The isotopic results also argue in favour of a recent deposition: the $^{87}Sr/^{86}Sr$ ratios of the smectite-residues from core 2, are lower than those of core 1. This difference requires a larger amount of volcanic-derived material relative to sea-water authigenic neoformed fibres. This delicate neoformed material, which resulted from chemical reequilibration with respect to the environment, obviously formed only at the present-day site; fibres definitely would break down during transportation. Furthermore the $^{87}Sr/^{86}Sr$ ratios of the leachates from the clay minerals of core 2 are higher than the corresponding ratios of core 1. If one assumes that the $^{87}Sr/^{86}Sr$ ratios of the leachates reflect the values of the depositional environment (Clauer *et al.* 1982) and that the $^{87}Sr/^{86}Sr$ ratio of the sea-water progressively increased during Tertiary and Quaternary times (compilation in Faure 1982), then the deposition of the sediment of core 2 was clearly later than that of core 1. Moreover, a very young age for the sediments in core 2 is expected by their near uniform $^{87}Sr/^{86}Sr$ ratios despite some variations in their Rb/Sr ratios.

Several geochemical data, such as the random variations of Mg and Fe, the progressive increase of Si, and the near constancy of Na and Ti values with depth, do not suggest any crystallographic reorganization of the clay minerals; it is, therefore, obvious that the sediment of core 2 consists of only minor amounts of reorganized minerals. This is corroborated by the surprising variations of the water content of these sediments, even if the alkali-elements seem to concentrate with depth.

In summary, different segments of apparently the same sedimentary facies may occur in the southern Pacific Ocean within distances of 100 km. These sediments may have been deposited at different geological periods, even recently, although exhibiting considerable similarities in mineralogical, morphological and chemical make-ups. Inheritance at a given site of authigenic sediments formed elsewhere precludes the traditional use of a sedimentation rate uniformly applied to the entire basin.

Conclusions

Sedimentological, mineralogical, chemical and Rb-Sr isotopic studies of clay minerals from deep sea red clays from the southern Pacific Ocean clearly show:

1. that this sediment is authigenic on a very wide (ocean basin) scale;
2. that it contains variable but small amounts of volcanic-derived chemical components;
3. that it was transported from some place in the ocean basin and deposited at the present site, where it is inherited.

This removal from one site to another was likely induced by topography and bottom currents. The existence, within a 100 km distance in the southern Pacific Ocean, of two sedimentary horizons, similar in facies but different in age, induces a strong reconsideration of the traditional concept of sedimentation rate derived by dividing the thickness of sedimentary piles by their maximum age. The following aspects should, henceforth, be considered in the evolutionary models of sedimentary basins.

1. The sedimentation rate, which is commonly used, is calculated from the rate of accumulation of particles by time-unit. It may be easily obtained for sediments of continental or biogenic origin. We believe that it can only be used for red clays which contain clay minerals of continental origin, that is to say those which do not fulfill Renard & Murray's definition. The sedimentation rates measured for these red clays are always small, in the millimetre range for 1000 years (see i.e. Arrhenius 1963).
2. The commonly used concept of sedimentation rate can no longer be applied to red clays, from the southern Pacific Ocean at least, as they are of authigenic origin. The different components of this oceanic facies formed at the water-sediment interface, or slightly below in the uppermost part of the sedimentary level by modification of unstable pre-existing particles or by ionic precipitation from sea-water. The thickness of the red clays depends, therefore, on the amount of authigenic sediments formed from a preexisting material which weathered and modified after deposition. Subsequently, the appropriate concept is *formation and accumulation of authigenic material by time-unit* rather than sedimentation rate.
3. The results of the present study show that even for authigenic sediments the accumulation which includes deposition may be discontinuous at a given site of the ocean floor. Authigenic particles may indeed be transported by bottom currents and deposited again somewhere else, but this does not lead to the commonly used sedimentation rate.

Furthermore, the existence of red clay horizons several metres thick, which are in chemical desequilibrium with their environment, suggests powerful and sudden submarine transport processes. Massive sedimentation phenomena due to strong currents may, therefore, be proposed. They favour catastrophic events on geologic scales which in turn induce a 'poles-apart' view to the classical one which predicts low sedimentation rates. In addition to the isotopic arguments, the hypothesis of sudden displacement and of submarine erosion and transportation of authigenic particles, are supported by the known thickness of the red clay facies. In the southern Pacific Ocean, it never exceeds 8 m and may fluctuate within several metres at a same sampling station (Hoffert 1980).

Thus, the traditional premises for the calculation of sedimentation rate cannot be directly applied to the specific red clay facies. At a given site of the ocean, its thickness corresponds to a complex geological history of a *submarine sedimentary cycle*. The particles, at the water-sediment interface in the oceans, originate partly from parent-rocks which experienced weathering and dissolution to produce authigenic minerals. After erosion, transportation and deposition of these newly formed minerals, diagenetic modifications lead to the red clay facies. Thus, a given red clay horizon may correspond to the action of a submarine pedological event, or to a sudden and massive accumulation of sediments, or to a very low accumulation of particles formed elsewhere in the deepest and quietest part of the ocean. It is suggested therefore that, before attempting to determine the sedimentation rate for the red clay facies, genetic information is acquired. It seems also prudent to avoid any generalization over the entire facies, and to admit that this sedimentation is inconstant at an ocean-wide scale.

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Rates of Neogene depositional and deformational processes, north-west Himalayan foredeep margin, Pakistan

Robert G. H. Raynolds and Gary D. Johnson

SUMMARY: The dynamics of molasse accumulation and the subsequent structural displacement of a foredeep basin during orogenesis have rarely been chronometrically defined. The Himalayan foothill belt of northern Pakistan and India constitutes a detailed record of molasse sedimentation in a classical foredeep setting. Subsequent deformation of this sedimentary sequence has resulted in the exposure of over 3000 metres of the Neogene and Quaternary aged Siwalik Group. Recently, the development of a magnetic polarity stratigraphy constrained by the fission-track dating of numerous intercalated volcanic ashes has permitted a precise chronostratigraphic statement to be made concerning the accumulation and deformational history of this non-marine sedimentary sequence. Data from numerous localities in northern Pakistan and India document the onset, development, and termination of various events in this sedimentation history with a precision of from ~20000 to ~50000 yrs. Sediment accumulation rates associated with the progressive migration of the foredeep depocentre have been determined to range from ~20 to ~50 cm/1000 yrs. The southward progradation of lateral facies changes at rates of up to 30 m/1000 yrs together with the southward advance of the basin depocentre at rates of over 20 m/1000 yrs suggest a rough equilibrium between the modelled northward plate motion of the Indian subcontinent and the southward displacement of depositional processes within the foredeep. This illustration of the dynamic involvement of proximal foredeep terraine in the continuing Himalayan orogenesis yields a model potentially useful in similar tectonic settings.

Late orogenic fluvial molasse sediments comprising the Neogene- and Quaternary-aged Siwalik Group are exposed in the southern foothills of the Himalayan mountains in northern Pakistan and India. This folded and thrusted terrain has uplifted and exposed over 3000 metres of Miocene and younger sediments in the northern Punjab and an adjacent portion of south-western Kashmir, Pakistan and India. Particularly good exposures result from the interference of north-east and north-west oriented structural trends associated with the eastern termination of the Salt Range and the Pir Panjal foothills, respectively (Fig. 1). This conjunction of structural trends, the Jhelum re-entrant (Visser & Johnson 1978), tectonically constrains the entry of the modern Jhelum river, a structurally antecedent stream, onto the Indo-Gangetic plain.

The Himalayan foredeep is an ideal setting for the examination of the dynamic processes associated with molasse basin evolution. Because depositional and deformational processes are continuous to the present, modern analogues exist both for the depositional environments responsible for

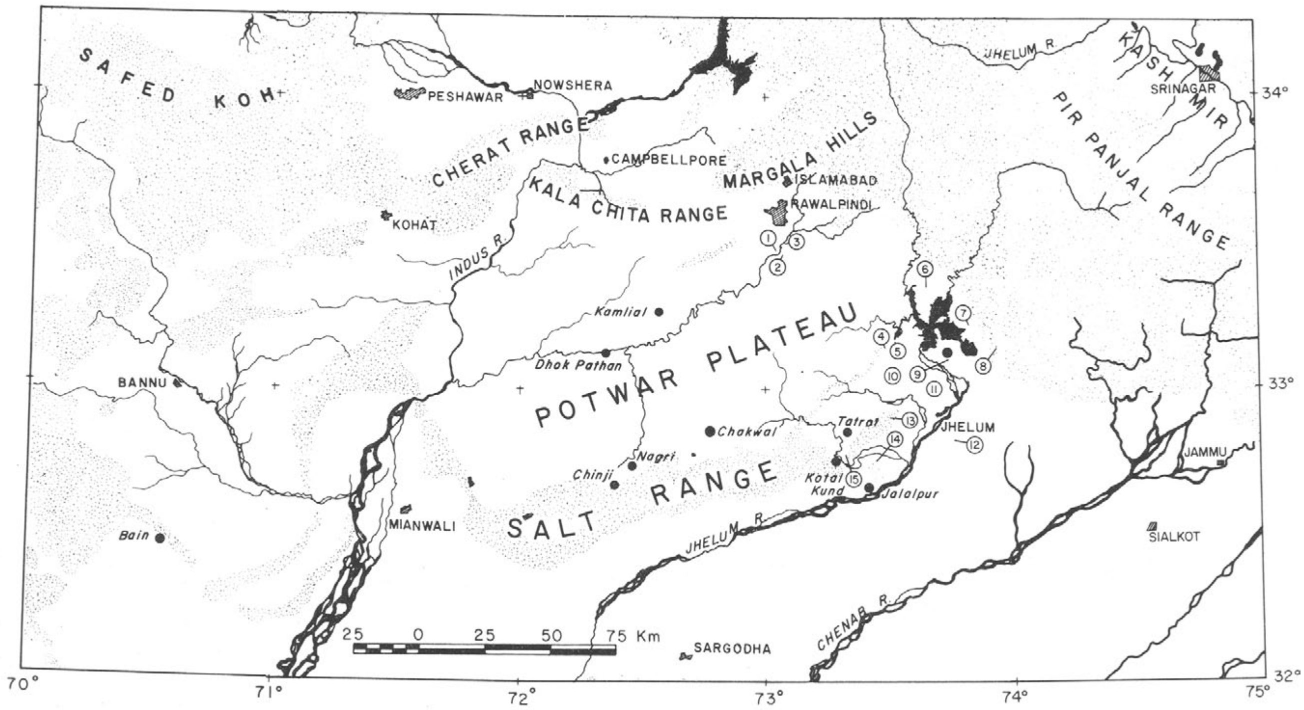

FIG. 1. Location of dated stratigraphic sections from northern Punjab and adjacent south-western Kashmir, Pakistan, which are discussed in this study. Numbers are keyed to Figs 2, 5, 9, 10, and 11.

the majority of tbese late orogenic sediments, and for the onset and development of tbe compressional tectonic style that has deformed the proximal, northern margin of the foredeep basin. To the above assets has been added an absolute chronostratigraphy developed through a synergistic application of palaeomagnetic stratigraphy and fission track dating.

Magnetic polarity stratigraphy

Fine-grained sediments, common as overbank or floodplain facies in non-marine rocks like the Siwalik Group, usually preserve a composite record of the orientation of the Earth's magnetic field prevailing at the time of deposition together

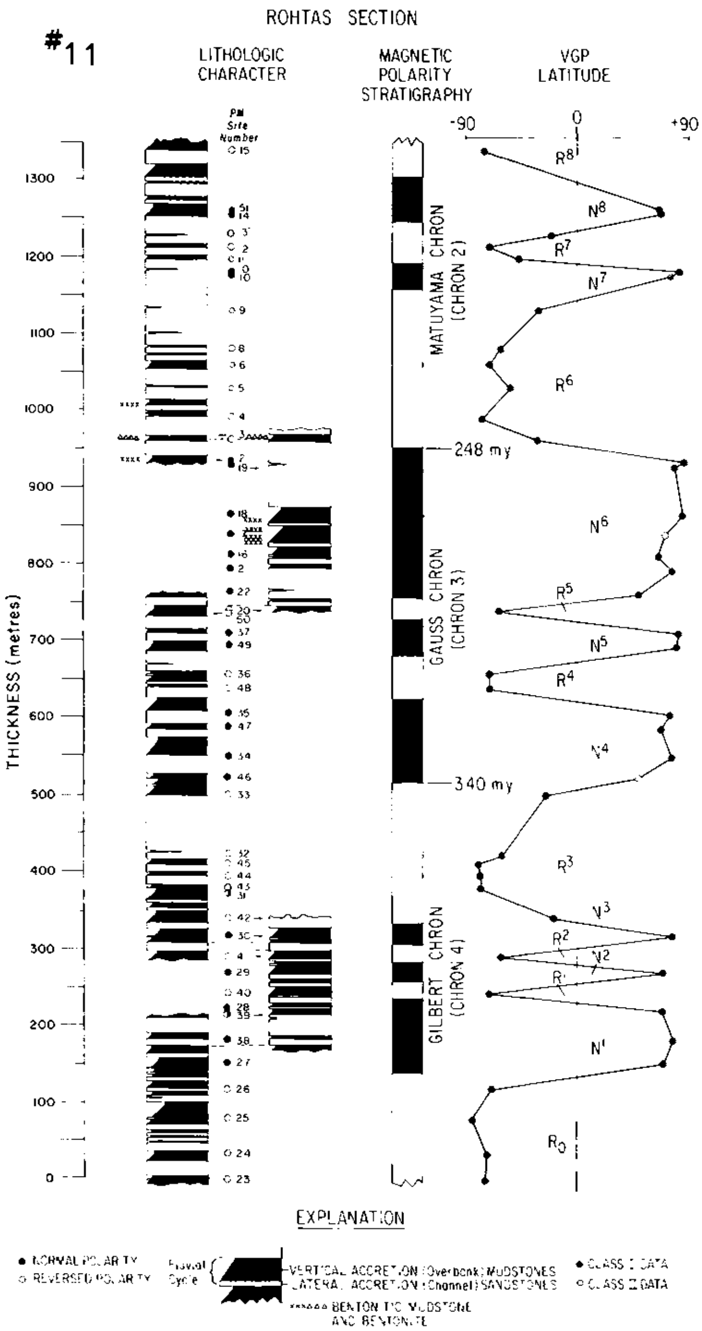

FIG. 2. Schematic lithostratigraphy and magnetic polarity stratigraphy of Siwalik Group sediments exposed at north-eastern nose of Rohtas anticline (locality 11, Fig. 1), vicinity Jhelum, District Gujarat, Punjab, Pakistan.

with that subsequently acquired during burial diagenesis. This natural remanent magnetization (NRM) can be modified by a variety of demagnetization techniques which remove post-depositionally acquired magnetic overprints, revealing the original character of the depositional remanent magnetization (DRM).

Manifest in such procedures is the ability to recognize intervals of unique normal and reversed polarity in the DRM data and to develop a stratigraphic sequence of superposed magnetozones. This stratigraphic sequence, developed at a given site, results in the establishment of a local magnetic polarity stratigraphy (MPS) which is a history of magnetic field reversal at the site. Local MPS may then be correlated with an established world-wide magnetic polarity time-scale (MPTS) allowing for a chronometric interpretation of the local MPS.

The following discussion results from the establishment of 15 local MPSs in the north-western Himalayan foredeep margin and their correlation with the presently accepted MPTS for the late Neogene and Quaternary.

From three to five carefully oriented magnetic samples were collected at regular stratigraphic intervals within each of the 15 localities. Fine-grained siltstones, typical of overbank deposits, were selected for sampling according to procedures outlined by Johnson *et al.* (1975). Alternating field and/or thermal demagnetization of all samples was carried out. In general, younger samples required alternating field demagnetization at field strengths of up to 150–200 oersteds in order to discern the original orientation of the DRM. In cases of sedimentary sequences where diagenetic factors are likely to have strengthened the post-depositional component of the NRM, thermal demagnetization at temperatures of up to 500°C was employed in addition to the Af demagnetization. Samples were spun on a commercially available spinner magnetometer following procedures and techniques described in Opdyke *et al.* (1979) and N. M. Johnson *et al.* (1982).

A virtual geomagnetic pole (VGP) was calculated for each sample set, and statistics describing dispersion on a sphere (Fisher 1953) were used to classify the data into Class I (Fisher statistic $k > 10$) or Class II (Fisher statistic $k < 10$) in order to quantity the confidence of the VGP construction (N. M. Johnson *et al.* 1982, Opdyke *et al.* 1977).

An example of a local MPS is illustrated in Fig. 2. Here, at the north-eastern plunge of the Rohtas anticline of the eastern Salt Range (Fig. 1, locality 11; Fig. 3) the interpreted magnetic polarity stratigraphy resulting from 47 stratigraphically superposed sample sites is comprised of eight normal and nine reversed polarity zones. This polarity zonation, based on 22 normal (20 Class I and 2 Class II) and 25 reversed (all Class I) sites is correlated with the MPTS in Fig. 4.

This type of interpretation can be equivocal unless constrained by additional data such as is provided by palaeontological or radiometric control. Most of the sections discussed herein contain fauna representative of the Tatrot and Pinjor faunal zones of South Asia (Opdyke *et al.* 1979) and radiometrically dateable volcanic ashes (G. D. Johnson *et al.* 1982). These two controls suggest that the Siwalik rocks under study occur in the upper portion of the Group, having been deposited in Pliocene and Pleistocene time.

In another example, two dated volcanic ashes occur in the Jhel Kas section on the flanks of the Mangla-Samwal anticline (Fig. 1, locality 9; Figs 5 and 6). In this case, the local MPS is correlated to the MPTS on the basis of both the MPS pattern

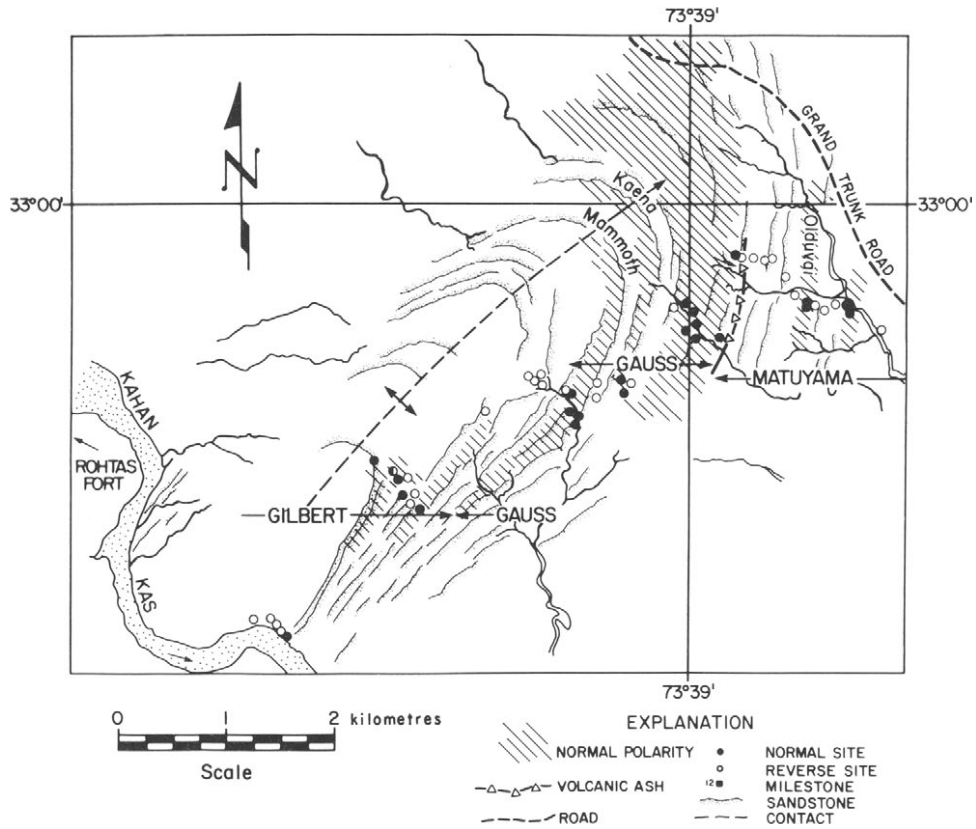

FIG. 3. Location of magnetic sample sites, north-eastern nose, Rohtas anticline, 4 km north-west of Jhelum, District Gujarat, Punjab, Pakistan.

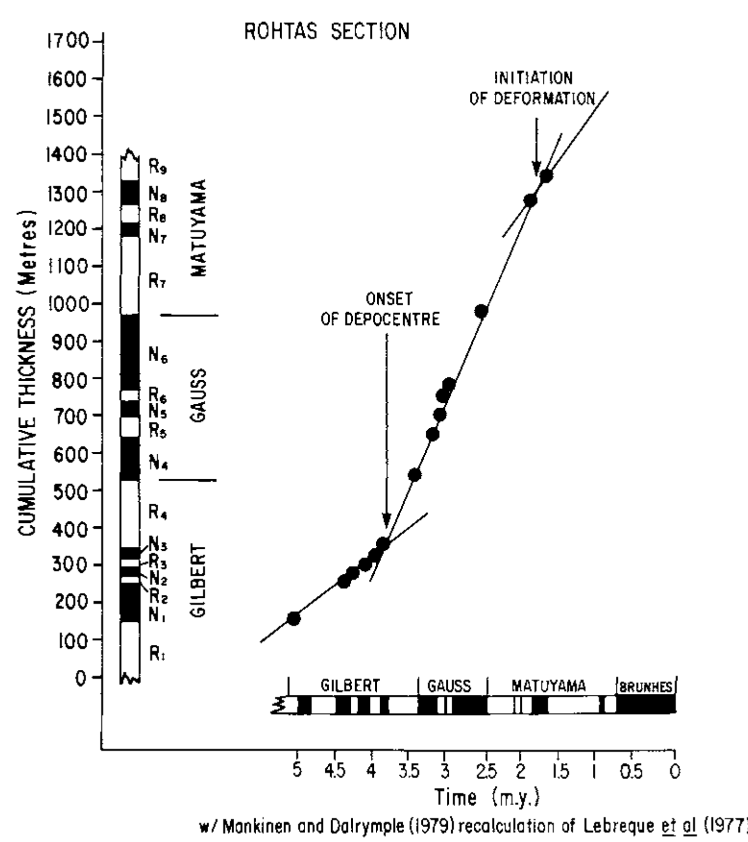

FIG. 4. Correlation of the Rohtas local magnetic polarity stratigraphy to the magnetic polarity time-scale of Mankinen & Dalrymple (1979). This provides for a chronometric interpretation of the sequence of polarity reversals (R_1-R_9, N_1-N_8) at Rohtas.

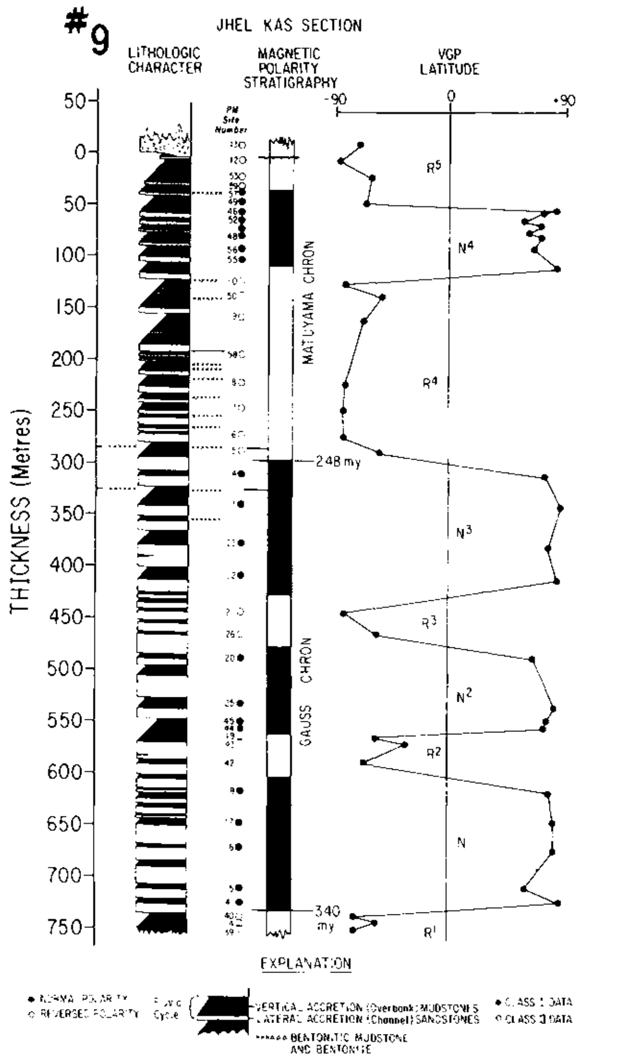

FIG. 5. Schematic lithostratigraphy and magnetic polarity stratigraphy of Siwalik Group sediments exposed in the Jehl Kas on the Mangla-Samwal anticline south of New Mirpur, Azad Kashmir, Pakistan. The pair of volcanic ashes astride the Gauss–Matuyama polarity boundary at 285 m and 325 m have been dated radiometrically by the fission-track method (zircon) at 2.40 ± 0.20 Ma and 2.56 ± 0.21 Ma respectively (G. D. Johnson *et al.* 1982). Modified from Opdyke *et al.* (1979) and G. D. Johnson *et al.* (1979).

and the radiometric constraints derived from fission-track ages on zircon microphenocrysts in two intercalated volcanic ashes (Fig. 7) (Opdyke *et al.* 1979; G. D. Johnson *et al.* 1982).

This approach of using palaeontological, radiometric, and MPS patterns has been used throughout our correlations of local stratigraphic sequences with the magnetic polarity time-scale. Given this chronometric control, we have made the following observations:

Dynamics of molasse accumulation

The migration of a foreland basin (foredeep) away from an evolving orogen involves the progressive displacement of the locus of maximum sediment accumulation toward an external foreland (Bersier 1948; Bally *et al.* 1966; Jordan 1981). The dynamics of sediment accumulation contributing to an orogenically derived clastic package can be schematically illustrated as in Fig. 8. At a given site on a distal foreland margin, initial sediment supply derived from the distant orogenic belt is low: evidence the character of fluvial sedimentation on the Indo-Gangetic plain at Delhi, India, where the waters of the Jamuna are just onlapping the Precambrian rocks of the Aravalli Range of the Indian foreland. For an initial period of time the rate of sediment accumulation at a site such as this, which can be considered part of the distal foredeep, will remain low. Subsequently, the growing proximity of the orogenic margin results in an increased sediment supply coupled with increased rates of basin subsidence. These two factors combine to yield a maximum rate of sediment accumulation (Fig. 8).

With continued convergence of the orogenic belt, the site in question receives more source-proximal sedimentary facies and ultimately becomes involved in uplift and deformation. As a result, the rate of sediment accumulation tapers and eventually the site undergoes denudation.

As illustrated in Fig. 8, two inflection points are anticipated in the accumulation rate curve. The early inflection represents the increased sediment accumulation rate associated with the onset of the foredeep depocentre. The latter inflection represents the passage from depocentre to proximal foredeep involvement in the orogenic margin. Inflections such as these are illustrated by the data from Rohtas and Jhel Kas, respectively (Figs 4 and 7). The locus of maximum sediment accumulation in a subsiding foredeep basin then is defined by the interaction of subsidence rates and sediment supply rates. A medial longitudinal zone of maximum accumulation rates is expected, with more distal portions of the foredeep receiving less sediment, and more proximal regions being subject to the onset of uplift and ultimately erosion of the synorogenic sedimentary sequence.

Sediment accumulation history

The magnetic polarity stratigraphy of 15 localities in the northern Punjab and adjacent south-western Kashmir is illustrated in Figs 9, 10, and 11. The correlation of these MPSs with the MPTS has been established using the same criteria as that demonstrated above for Rohtas and Jhel Kas. The resultant sediment accumulation curves are illustrated in Fig. 12. These data allow us to discuss the temporal and spatial sediment accumulation history of this portion of the Himalayan foredeep basin. The magnetic polarity stratigraphy and resultant temporal interpretations are integrated histories of sediment accumulation for given localities. We are not reporting instantaneous rates of sedimentation, but rather rates of sediment accumulation integrated over chron boundaries of the MPTS. The sedimentary package, defined by enclosing isochrons, is a measure of the sediment accumulation history. This rate history may be suspect if one considers possible differential post-depositional compaction which theoretically may foreshorten the thickness of shale-rich intervals to a greater degree than sand-rich intervals. We have found this not to be a significant problem in the sites under review.

This differential compaction of high sand/shale ratio and low sand/shale ratio intervals between localities has been tested in the Jhelum area in Siwalik sediments bounded by two laterally extensive volcanic ashes (Visser & Johnson 1978). Their data suggest that compaction effects in these young sediments have been minimal.

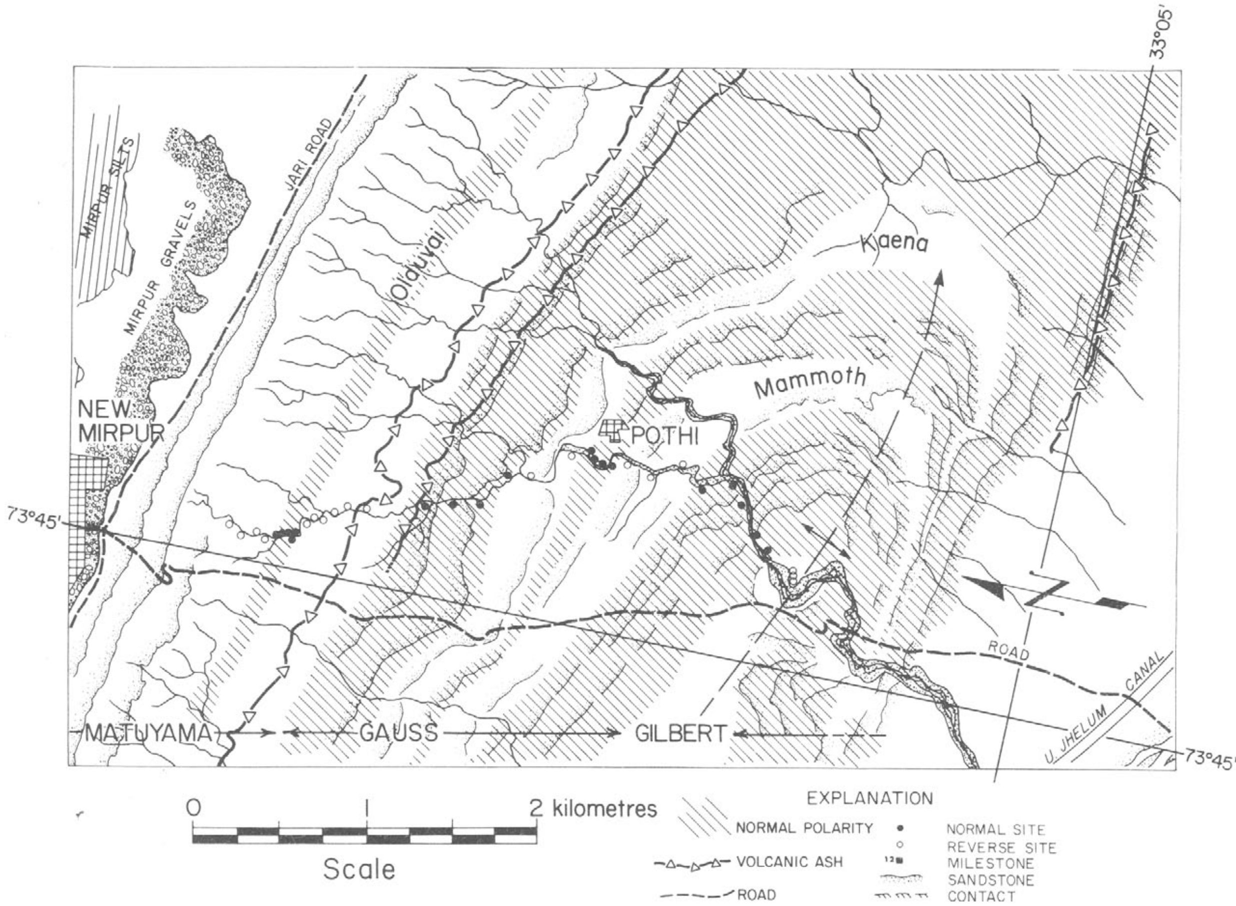

FIG. 6. Location of magnetic sample sites along the Jhel Kas, south of New Mirpur, Azad Kashmir, Pakistan. Modified from G. D. Johnson (1979).

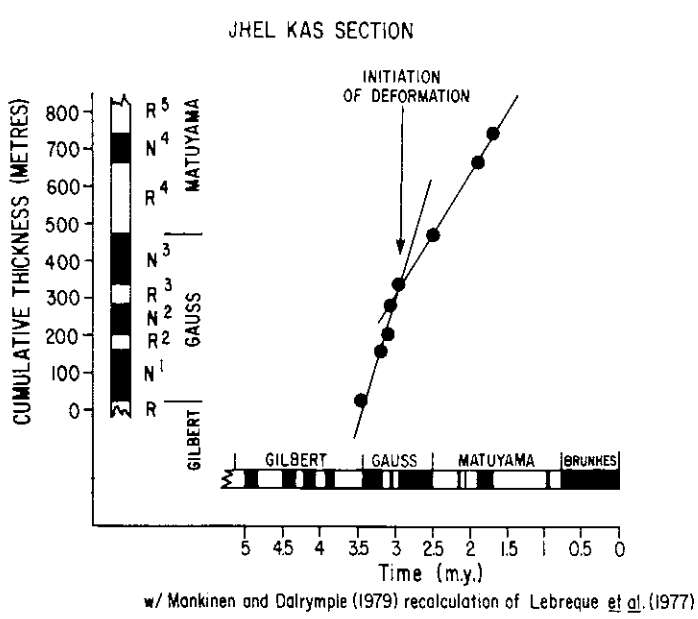

FIG. 7. Correlation of the Jhel Kas local magnetic polarity stratigraphy to the magnetic polarity time-scale of Mankinen & Dalrymple (1979).

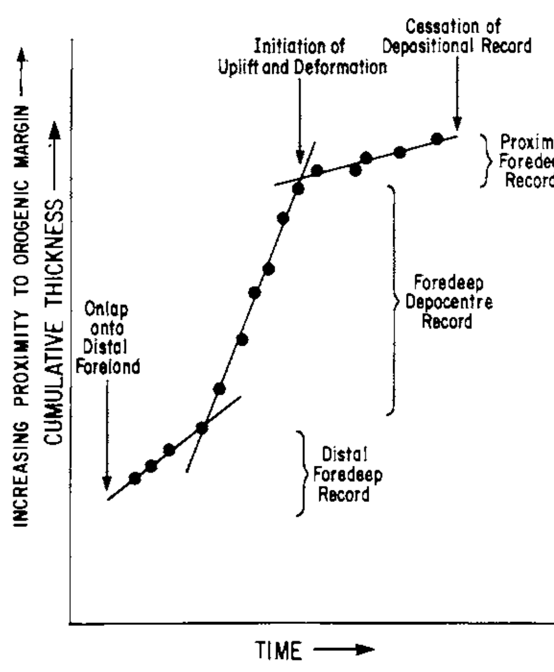

FIG. 8. Idealized cumulative sediment accumulation *vs* time illustrating the two principal inflection points related to: (1) rate increase associated with the onset of maximum sediment accumulation/foredeep depocenter conditions; and (2) rate decrease associated with the onset of uplift and deformation in the proximal foredeep.

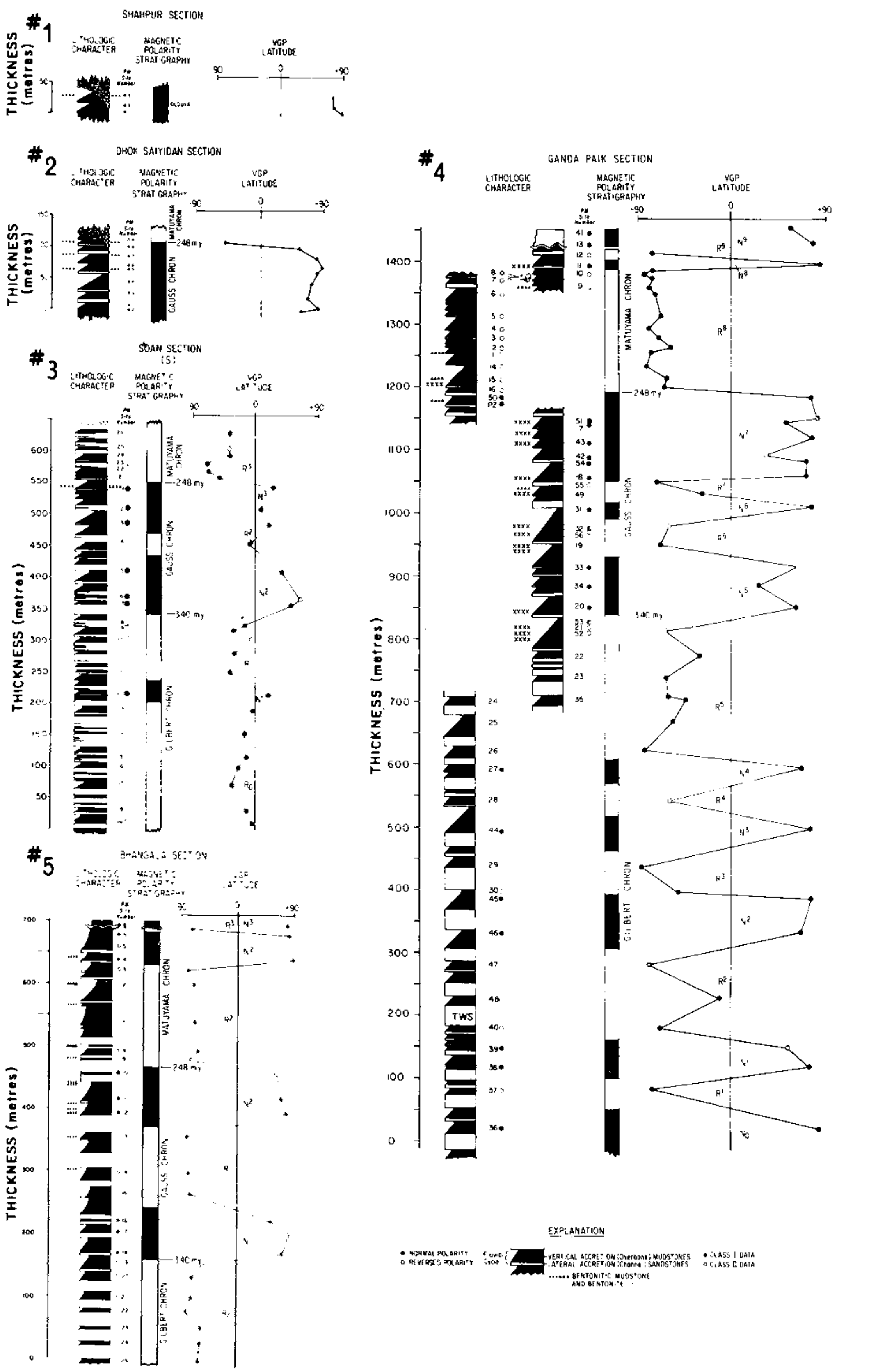

FIG. 9. Lithostratigraphy and magnetic polarity stratigraphy of selected localities in the eastern Potwar Plateau, northern Punjab, Pakistan. See Fig. 1 for section locations. Data for Shahpur (#1), Dhok Saiyidan (#2) and Soan (#3) modified from Moragne (1979), Raynolds (1980) and G. Johnson, unpublished. Data for Ganda Paik (#4) and Bangala (#5) from Raynolds (1980).

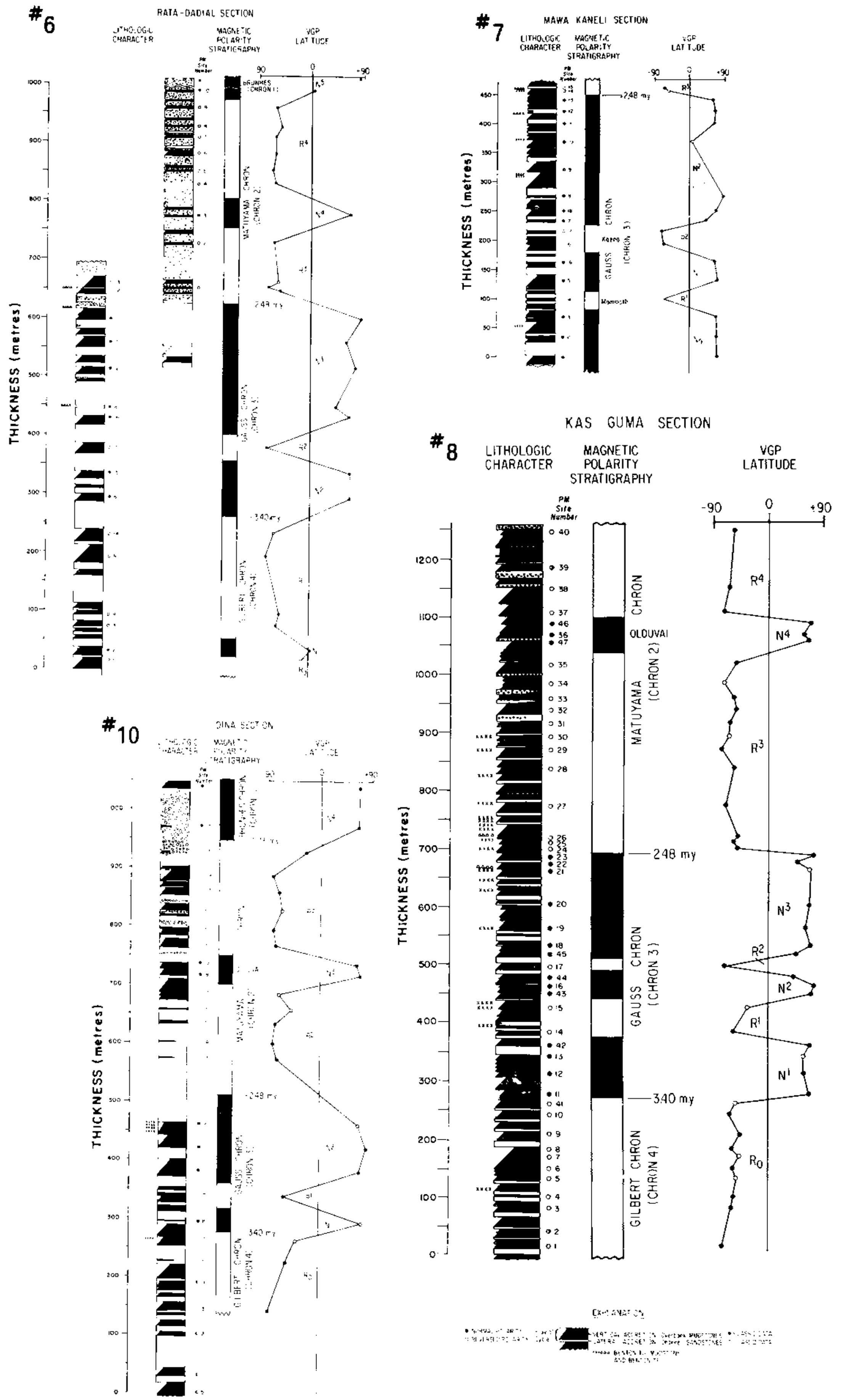

FIG. 10. Lithostratigraphy and magnetic polarity stratigraphy of selected localities in the Pir Panjal foothills, Azad Kashmir and portions of the eastern Salt Range, District Gujarat, Punjab, Pakistan. See Fig. 1 for locations. Data for all localities modified from Raynolds (1980).

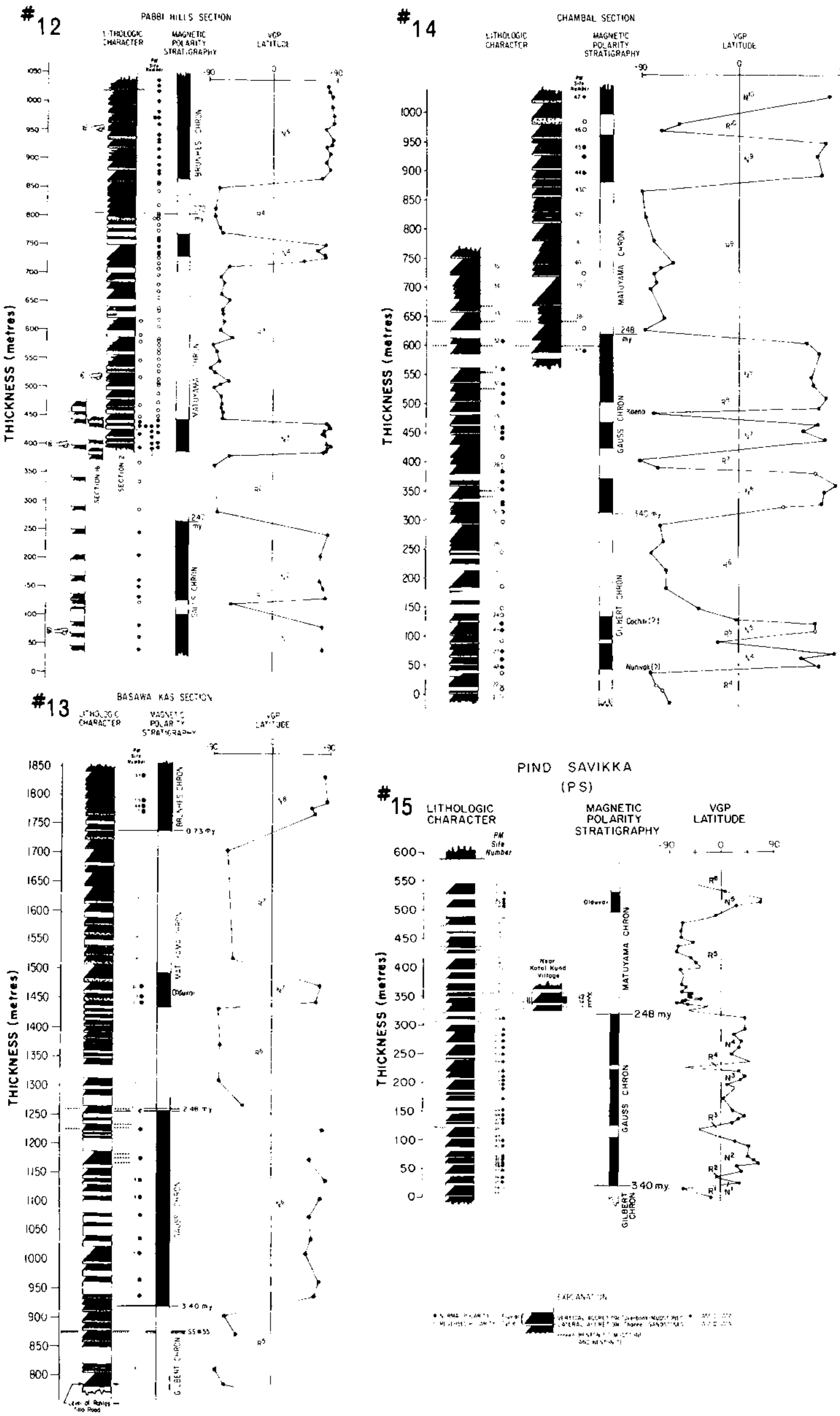

FIG. 11. Lithostratigraphy and magnetic polarity stratigraphy of selected localities in the eastern Salt Range, District Gujarat, Punjab, Pakistan. See Fig. 1 for locations. Data for Pabbi Hills (#12) modified from Keller *et al.* (1977) and G. D. Johnson *et al.* (1979). Data for Basawa Kas (#13) and Chambal (#14) modified from Opdyke *et al.* (1979) and G. D. Johnson *et al.* (1979). Locality data for Pind Savikka (#15) modified from Frost (1979) and G. D. Johnson *et al.* (1982).

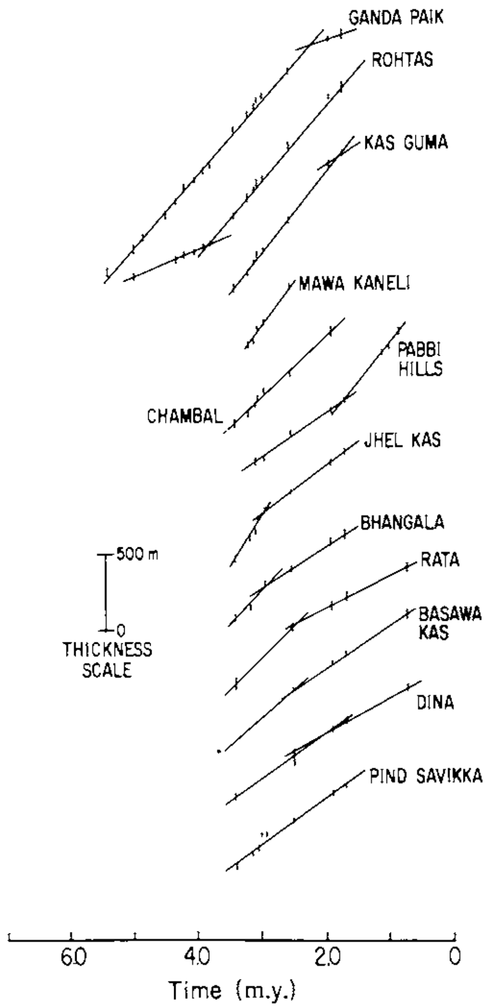

FIG. 12. Cumulative sediment accumulation for the Siwalik Group during the late Neogene–early Quaternary derived from the stratigraphic sections illustrated in Figs 2, 5, 9, 10, 11. See Fig. 8 for related interpretation.

Accumulation rate variation in space

The widespread distribution of accumulation rate data permits us to examine the spatial relationships of accumulation rates during selected intervals of time. Figure 13 illustrates the rate of sediment accumulation (in cm/1000 a) in the Jhelum area for time increments defined by magnetic polarity chrons and covering the span from 0.73 to 3.40 Ma.

During the Gauss chron (from 3.40 to 2.48 Ma ago), the depocentre, or locus of maximum sediment accumulation occurred in the north-eastern portion of the outer Jhelum re-entrant as defined in Fig. 13(a). During this interval, maximum sediment accumulation rates exceeded 50 cm/1000 a. Subsequently, the depocentre appears to have migrated south to within the vicinity of the modern town of Jhelum during the early Matuyama chron. During the late Matuyama chron, the depocentre was located to the south of the study area. These data document a minimum southward displacement of the depocentre of over 60 km in the 2.67 Ma of record at a rate of excess of 20 m/1000 a.

The pattern of sediment accumulation is modified by two factors. First, because the study area is located in the axis of the Jhelum re-entrant the supply of sediment is tectonically constrained to the axial portion of the structure along a line approximated by the course of the modern Jhelum river. This axial concentration of sediment is evident in the sediment accumulation data for the Gauss and early Matuyama chrons, both of which show elongate highs along the re-entrant axis. Second, the pattern of sediment accumulation is influenced by the fact that the sampled stratigraphic sections are located on the flanks of anticlinal structures. There may be a tendency for reduced rates of accumulation to be documented as the adjacent structure starts to deform or uplift. This is probably the cause for the isolated low rate of accumulation evident in the Gauss chron data in Fig. 13(a) (the Dina locality).

Accumulation rate variation in time

The data can also be examined in terms of the change in rate of sediment accumulation recorded at various localities over

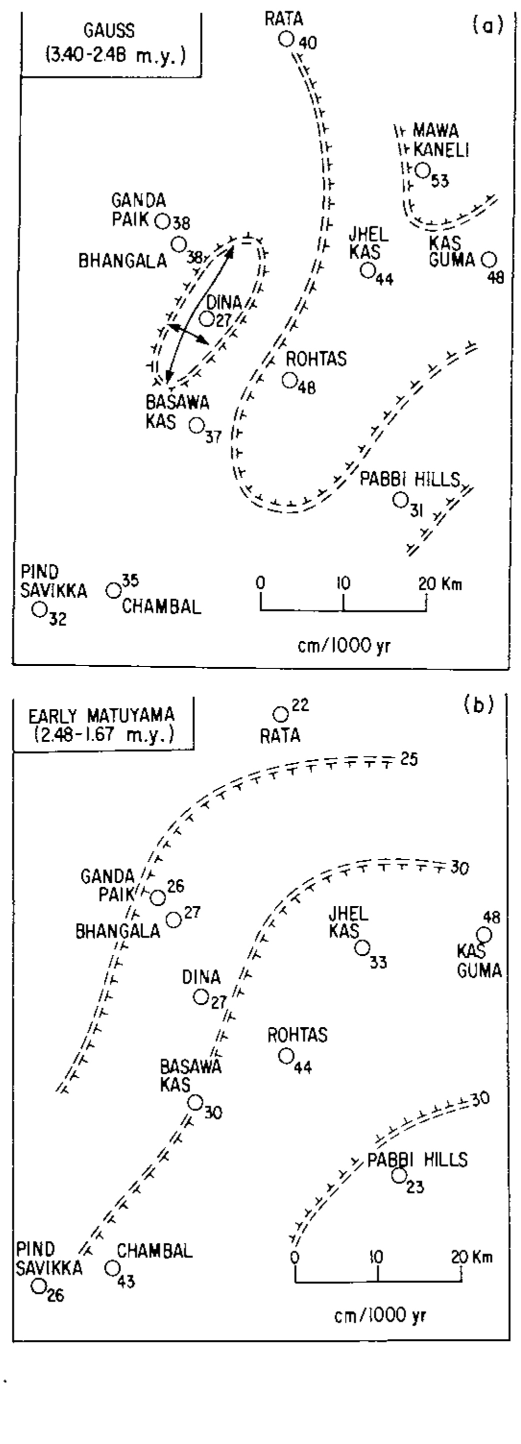

FIG. 13.

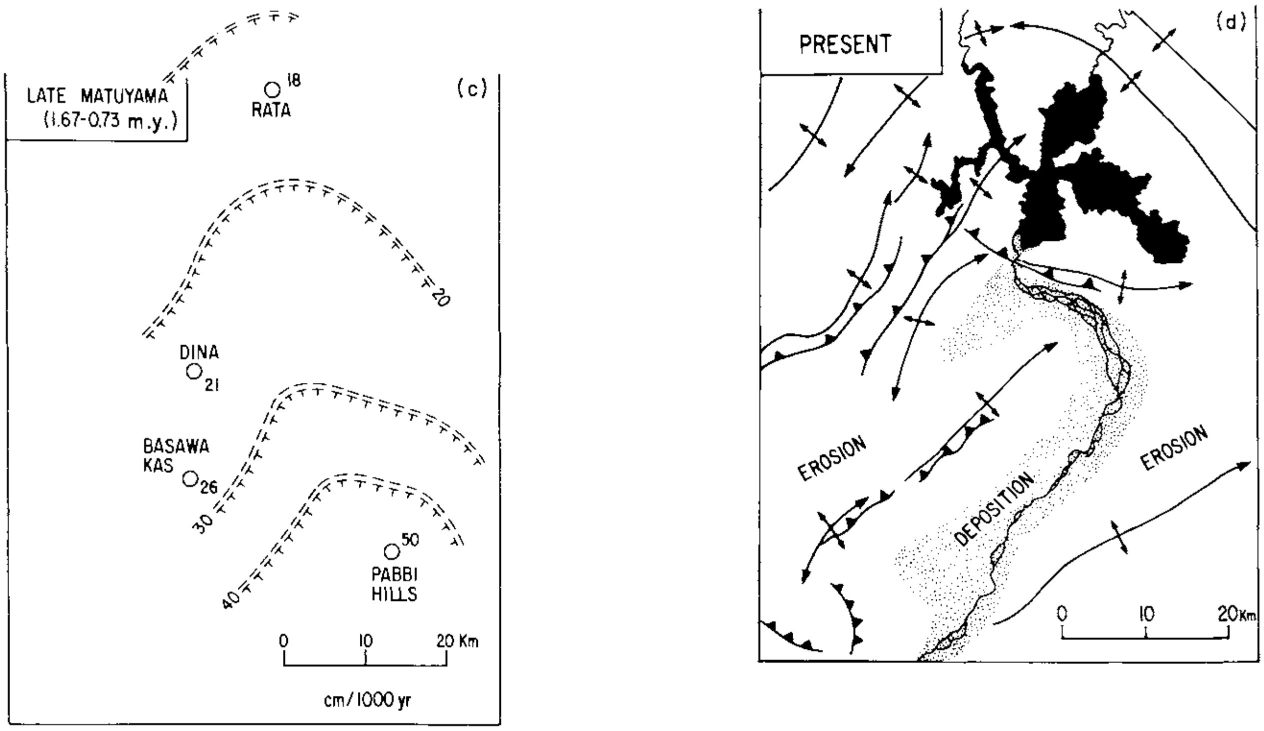

FIG. 13. Spatial distribution of Siwalik Group sediment accumulation (cm/1000 a) for four time increments (a)-(d) in portions of the eastern Potwar Plateau and Salt Range, Punjab, and south-western Kashmir, Pakistan. The progressive southward displacement of the basin depocentre over this time interval appears to be a manifestation of the southward migration of the foredeep deformational and depositional process.

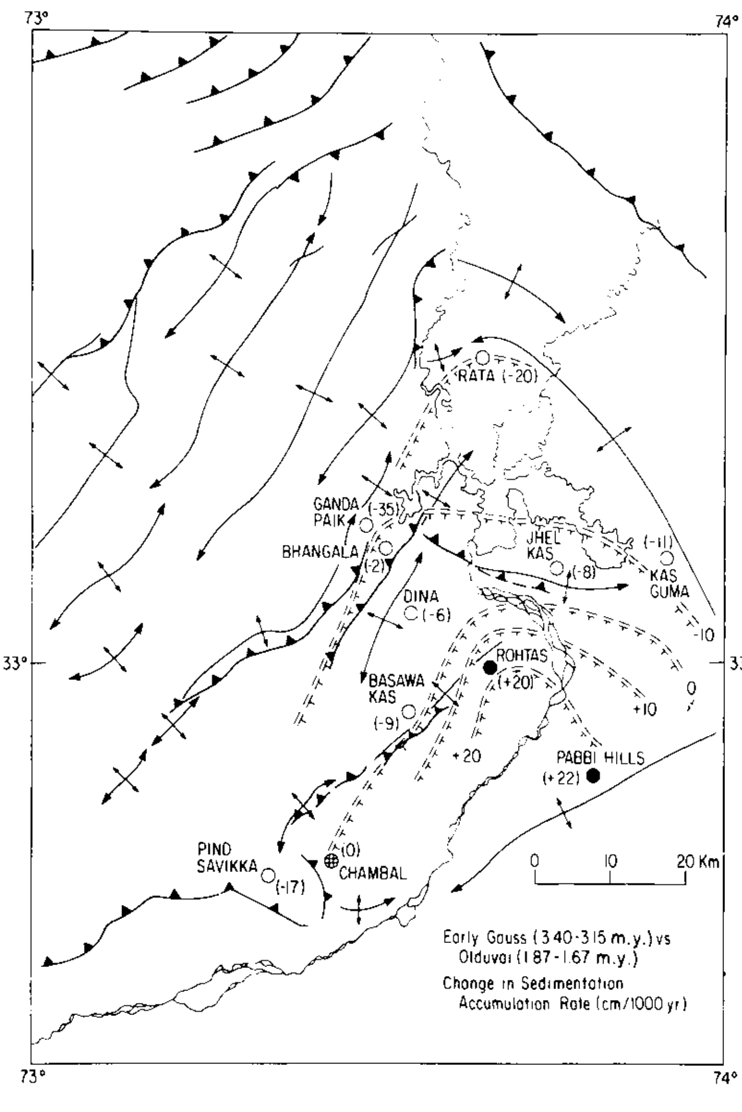

FIG. 14. Changes in rates of Siwalik Group sediment accumulation (in cm/1000 a) between 3.4 and 1.8 Ma BP in the Jhelum area. As predicted by the concept of foredeep migration, areas proximal to orogenic involvement record a decrease in accumulation rates while areas distal to this influence record an increase in rates over the same time interval. Localities exhibiting a decline in the rate of sediment accumulation, open circles; no apparent change, cross-hatchered circles; an increase in the rate of sediment accumulation, filled circles.

an interval of time. Figure 14 illustrates the change in sediment accumulation rates between the early Gauss chron and Olduvai subchron. As predicted by the model of a southward migrating foredeep taken in conjunction with the onset of structural growth along the northern periphery of the basin, these data document a decrease in accumulation rates in the northern portions of the area, while the southern regions experience an acceleration in the rates of sediment accumulation over this same interval of time.

The model developed above may be carried to the present, as most of the study area is currently undergoing denudation while the locus of maximum sediment accumulation is associated with the current thalweg of the Jhelum river where it flows across a low relief floodplain south of the study area (Fig. 13(d)).

An instantaneous sediment accumulation rate picture may be constructed for the Gauss-Matuyama boundary at 2.48 Ma as illustrated in Fig. 15. Structural control on sediment accumulation patterns can be identified within the re-entrant axis. The east-west trending Mangla Samwal anticline and the Dina anticline both exhibit lowered rates of accumulation suggesting syndepositional structural growth at these sites.

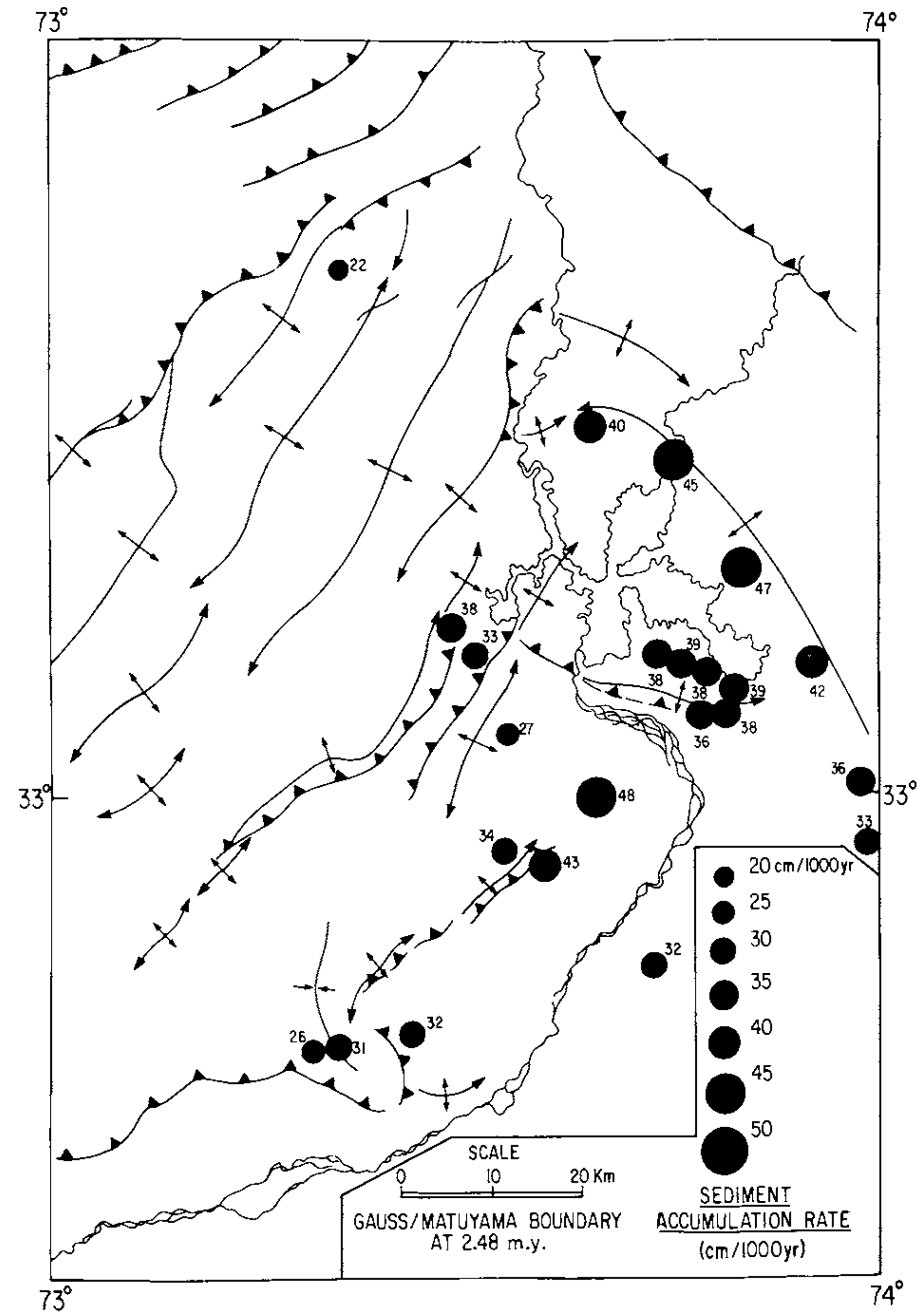

FIG. 15. Sediment accumulation rates for the Siwalik Group interpolated at the Gauss–Matuyama chron boundary at 2.48 Ma BP for the Jhelum area, Pakistan.

Rates of uplift and deformation

Two approaches have been used to quantify the rate of uplift in the north-west Himalayan foredeep. Both methods yield a value which represents minimum average rates of deformation.

First, a quantification of uplift can be arrived at by dating the youngest sediment deposited by a depositional system flowing across a site which subsequently becomes involved in the deformation. In the Himalayan context, the premise is that at the time aggrading rivers flowed across a site, there was no surface expression of the structure on the floodplain.

Second, when it is possible to establish the age of two stratigraphic sequences which bound an episode of deformation, the relief developed during the episode can be constrained to have been formed during the time interval delimited by the two dated sequences.

The first method has been applied to the Pabbi Hills, the southern-most anticlinal fold on the eastern side of the Jhelum re-entrant. The youngest fluvial facies on the flank of the anti cline were deposited on the order of 400 000 years ago (Keller *et al.* 1977). These beds do not now occur at the crest of the structure, but as shown in Fig. 16 there is structural relief of approximately 600 metres on this stratigraphic interval. The development of this relief within the available time indicates that the area underwent a minimum average uplift rate of 1.5 m/1000 a. This may be a conservative estimate.

The second approach can be demonstrated in the Soan syncline near Rawalpindi where a deformed sequence of Siwalik sediments is topped by a thin, essentially undeformed unit. The underlying deformed sequence is over 3000 metres thick with a MPS (Fig. 9) indicating that the youngest deformed strata in the structure are on the order of 2.1 Ma

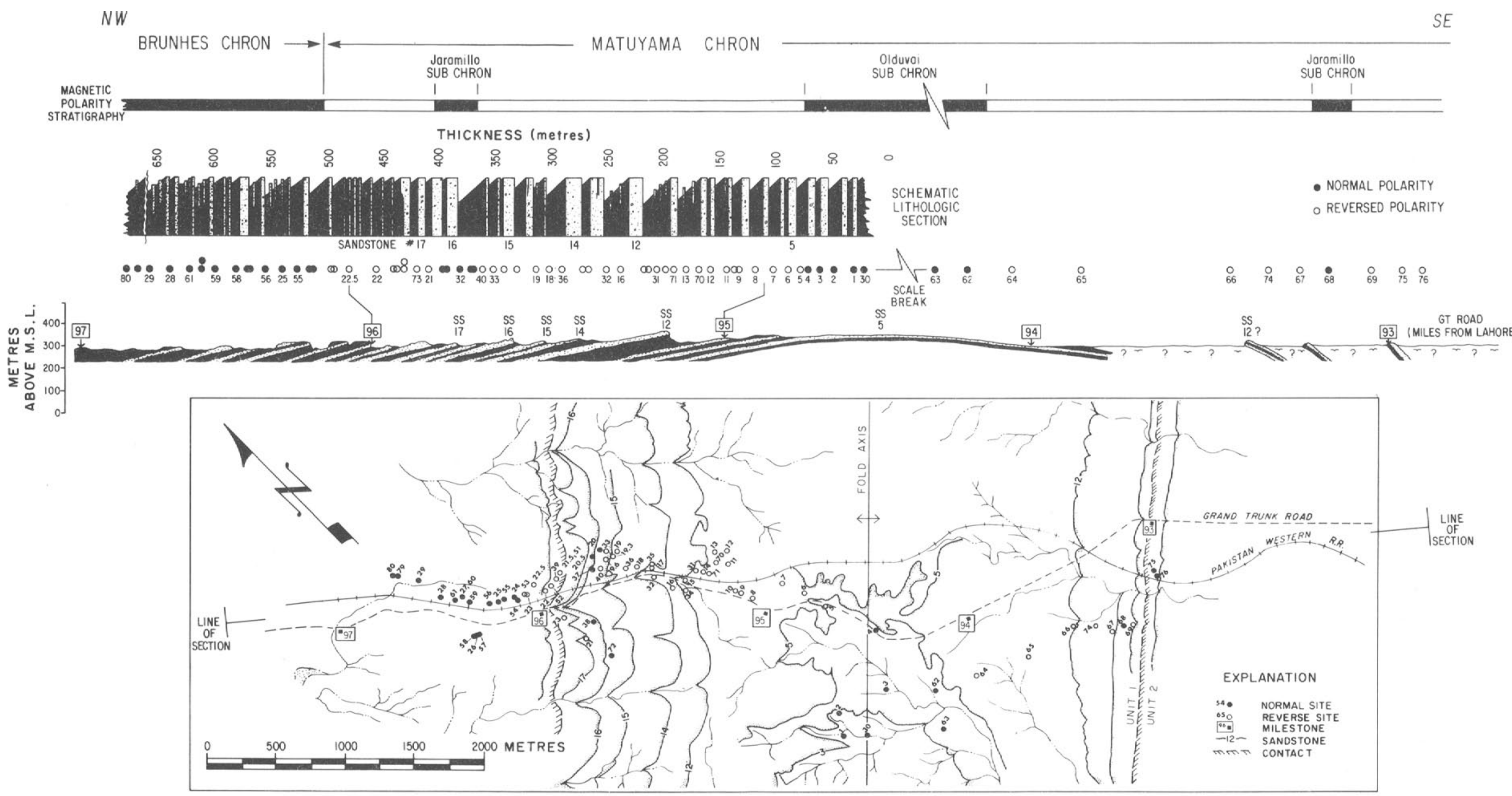

Fig. 16. Schematic structural section across the Pabbi (Kharian) Hills, 5 km south of Jhelum, showing some 600 m of relief developed since the deposition of the youngest exposed deformed strata on the structure. This horizon which occurs within the normally magnetized rocks of the Brunhes chron has an interpolated age of 400 000 yr BP and is represented by sample site 80 on the schematic lithologic section above. Data modified from Keller *et al.* (1977) and G. D. Johnson *et al.* (1979). Locality 12, Fig. 1.

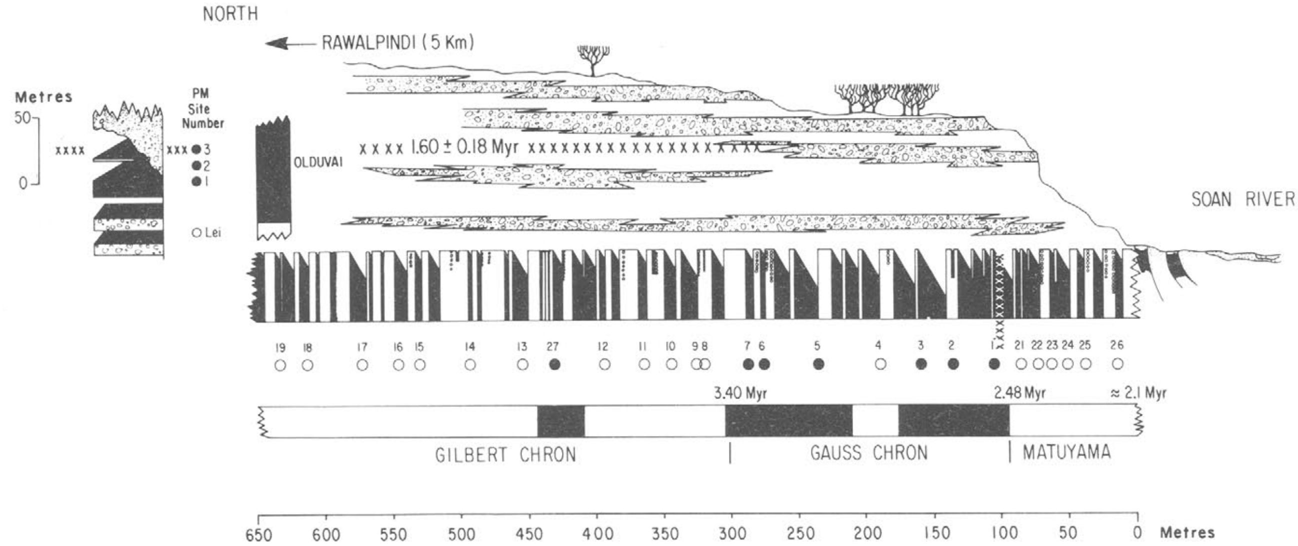

FIG. 17. Schematic structural section across the northern flank of the Soan Syncline, 5 km south of Rawalpindi, Pakistan, showing the disposition of two dated stratigraphic sequences which bound an episode of deformation responsible for the folding and peneplanation of over 3000 m of strata. Volcanic ash occurrence noted by (XXXXX). Modified from Moragne (1979) and G. D. Johnson and R. G. H. Raynolds, unpublished.

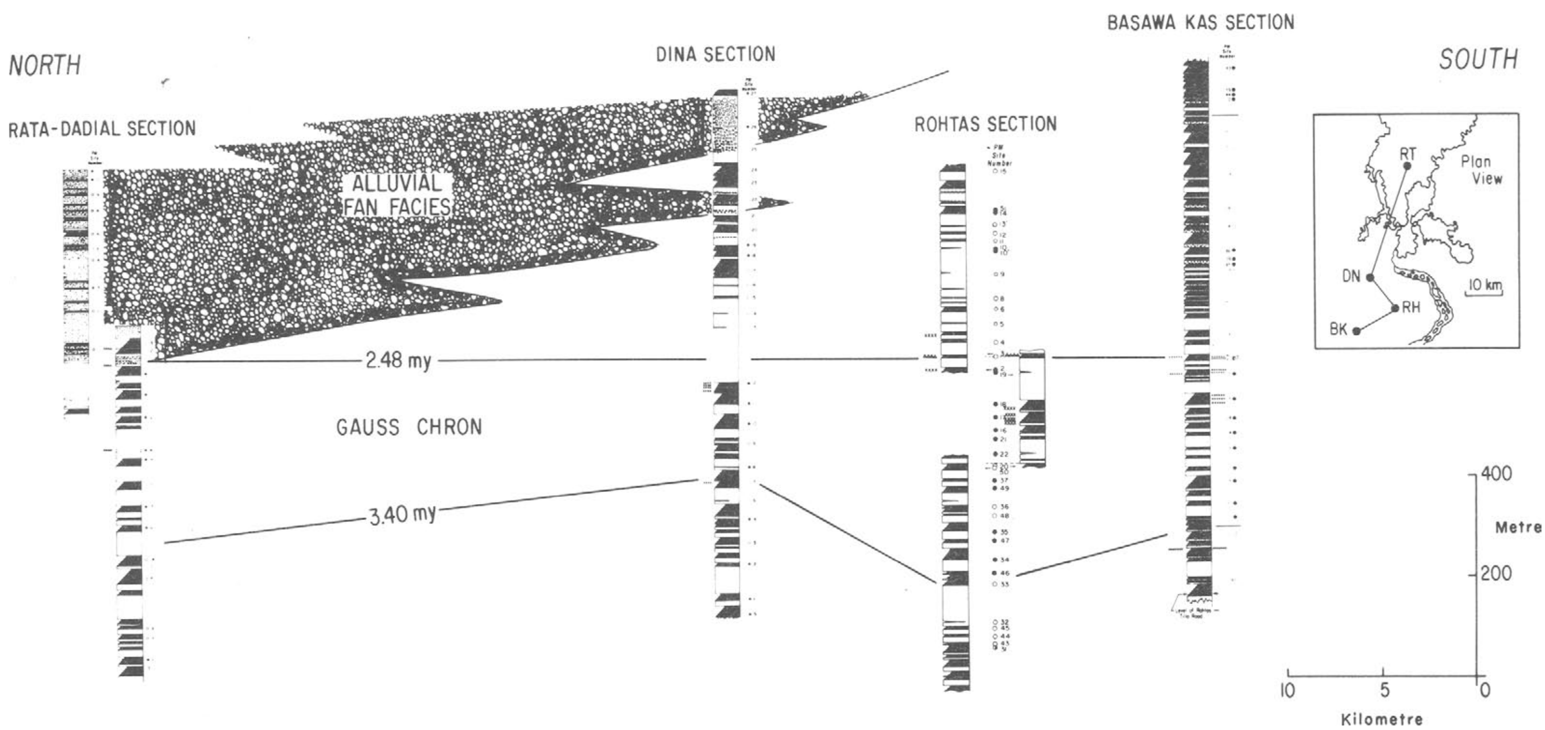

FIG. 18. Depositional dip section taken along trend of mean palaeo-current flow within the Jhelum structural re-entrant to illustrate the southward younging of the first stratigraphic appearance of a conglomeratic fluvial facies (the so-called 'Boulder Conglomerate' strata) within the upper portion of the Siwalik Group in northern Pakistan.

old. The overlying sequence has been dated on the basis of a MPS (Fig. 9) and the fission track dating of an enclosed volcanic ash (G. D. Johnson *et al.* 1982) with the oldest undeformed rocks shown to be a little less than 1.9 Ma old. This relationship is illustrated in Fig. 17. As over 3000 metres of structural relief were developed between 2.1 and 1.9 Ma ago, a minimum average uplift rate of 15 m/1000 a is indicated.

Rates of Facies Change

Facies changes within the Siwalik Group have long been the cause of correlation problems in regional mapping within the Himalayan molasse. Whereas the time transgressive nature of the rock facies boundaries has often been pointed out (Cotter 1933; De Terra & de Chardin 1936; Morris 1938), only relatively recently have workers used the necessary care to trace out individual horizons in order to yield valid regional lithofacies maps (Gill 1952; Behrensmeyer & Tauxe 1982).

Documentation of these regional facies trends is facilitated in part by the development of our chronostratigraphy for the Siwaliks of the eastern Potwar Plateau. Three examples of the application of chronostratigraphy to fluvial facies changes are cited below.

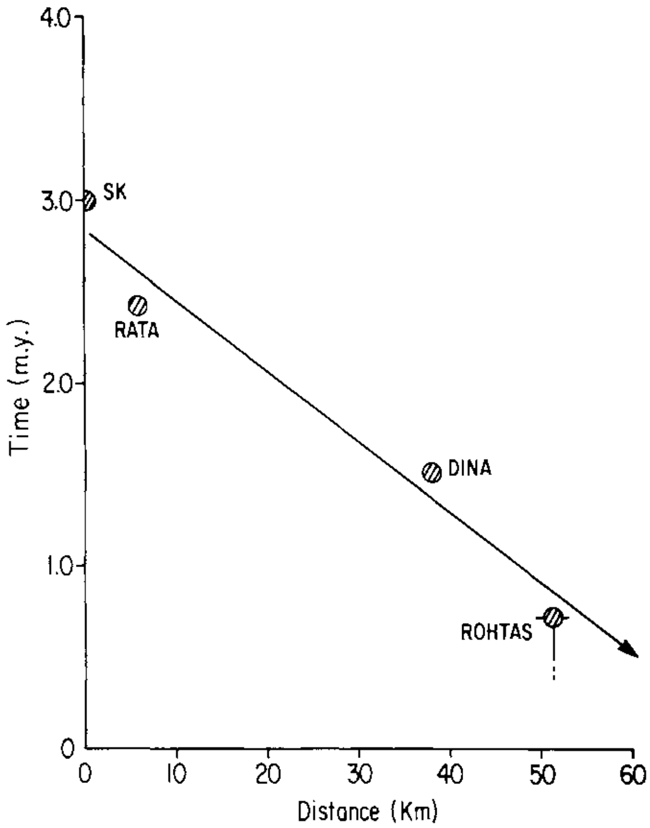

FIG. 19. The first stratigraphic appearance of conglomeratic fluvial facies *vs* time along the axis of the Jhelum structural re-entrant. Facies advance: about 3 cm/a.

In the region surrounding the modern town of Jhelum, there are pronounced changes in the colour and mineralogy of the sandstone bodies which make up the upper Siwalik Group. Our work shows that since approximately 4.5 Ma ago, a south flowing river system depositing brown coloured channel sands dominated by recycled sedimentary assemblages has characterized the study area. This river system displaced an older complex which had previously flowed in an easterly direction and which had deposited white coloured channel sands characterized by mineral assemblages derived from the crystalline Himalaya. It has been suggested on the basis of palaeocurrent studies and mineralogy (Raynolds 1981) that the white sand bearing rivers which flowed longitudinally down the axis of the foredeep basin may have represented the ancestral Indus River, and this river system was displaced from the study area by the transverse flowing Jhelum River system. This situation is comparable with that of the modern Ganges drainage in India. The Ganges flows longitudinally along the axis of the foredeep basin while its principal tributaries flow transverse across the basin as they debouch from the Himalayas. Similar patterns of fluvial systems have been reported from foredeep basins in Europe (Fuchtbaur 1967; van Houten 1974), and western Canada (Eisbacher *et al.* 1974).

Secondly, fluvial facies cyclicity can be demonstrated in all measured sections in the Upper Siwalik Group rocks (Figs 2, 5, 9, 10, 11). With time resolution based on the correlation of the local MPS with the MPTS it is possible to derive a fluvial cycle recurrence rate. This factor is a measure of the frequency of stream thalweg encounter at a given site through time, and ranges from 30000 to 50000 years for the Upper Siwalik sequences in the Jhelum area.

To vertebrate palaeontologists, this suggests that collections from a given locale are likely to sample fossil populations at 10^4 yr to 10^5 yr increments. The development of a biochronology for the Siwalik Group, therefore, faces the constraint of a non-continuous record at this level of resolution. Brief intervals of sedimentation history are stacked in a sequence in which channelling events may have removed significant portions of the time record.

In addition to the interplay of fluvial systems and fluvial cyclicity, the historically observed coarsening upwards of the Himalayan molasse is in actuality a time transgressive phenomenon resulting from the outward displacement of the orogenic front. Stratigraphic sections measured along a north-south transect within the Jhelum reentrant reveal the southward younging of the first stratigraphic appearance of a conglomerate sheet often termed the 'Boulder Conglomerate'. The onset of this lithofacies in the Siwalik Group has been mistakenly assumed by many workers to represent a synchronous pulse of clastics signalling a late stage orogenic event. Figure 18 illustrates the southward progradation of this facies change along the axis of the Jhelum re-entrant. In the northern stratigraphic sections which we considered, the first appearance of this distal braided stream facies occurs about 2.4 Ma ago, while to the south the earliest comparable facies is less than 0.7 Ma old. The southwards progression of the conglomeratic facies along the mean direction of palaeocurrent flow is illustrated in Fig. 19. From this it is apparent that the advance of the conglomeratic facies is taking place at a rate of approximately 30 m/1000 Ma (3 cm/a).

It is interesting to note the establishment of a form of equilibrium in this portion of the Himalayan foredeep, as estimates of the northward relative motion of the Indian subcontinent are on the order of 4 cm/yr (Minster *et al.* 1978), a value comparable in magnitude to this documented southward migration of both the foredeep basin depocentre and the first appearance of a widespread conglomeratic facies.

Summary

The above data provide a basis for understanding the dynamics of a portion of a modern foredeep basin. It is likely that patterns of geologic behaviour manifest in the Himalayan foredeep will be and have been repeated numerous times during the development and evolution of other tectonically comparable situations.

Patterns of variable rates of sediment accumulation, geologically rapid deformation, and abrupt time transgressiveness of lithofacies are thus to be anticipated in foredeep settings. Indeed, the understanding of the dynamic interaction of tectonics (both regional and local) and the associated deposition of sedimentary strata may make possible a clearer interpretation of the sedimentary record of orogenic activity provided by synorogenic clastic deposits. The Himalayan foredeep with its well constrained chronostratigraphy can serve as a model for the elucidation of less well preserved, or less well dated molasse basins and associated orogenic terranes.

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Geochronology in the Central Alps: dating of metamorphism, subsequent cooling with Oligocene glaciation, and the development of the recent landscape

E. Jäger

The interpretation of geochronological data

The first K-Ar and Rb-Sr measurements from the Central Alps immediately revealed that the measured age does not date the metamorphic mica formation but a later event. Samples from areas of simple geological relationship have been analysed to learn which parameters control the age results: biotites of post-metamorphic pegmatites from the Verzasca valley (Ticino) and from adjacent Verzasca gneiss yield identical Rb-Sr ages. The same is true for biotite flakes of 20 cm diameter from open fissures and fine grained hiotite from the nearby gneiss. This means that Rb-Sr biotite ages cannot date the climax of the metamorphism nor a compression phase.

When mica ages are related to the *Tertiary Lepontine metamorphism* of the Central Alps, certain patterns in the regional age distribution can be observed when rocks of granitic composition are analysed:
1. pre-Lepontine-metamorphic Rb-Sr and K-Ar ages on biotites are found only in areas of low grade metamorphism, below the stilpnomelane stability;
2. a zone of intermediate biotite ages (between pre- and post-metamorphic ages) follows metamorphic isograds; for rocks of granitic composition the intermediate age zone corresponds to the zone of Alpine stilpnomelane. In resistant rocks, more basic rocks which are poor in fluid phases, such as pre-Alpine eclogites and granulites, pre-Alpine biotite ages are found, even in rocks from the zone of Alpine chloritoid;
3. within the area of Alpine staurolite only Alpine biotite ages are found, independent of the type of rock and the biotite grain size; within a certain region, all samples from granites and gneisses to pegmatites and open fissures yield identical biotite ages. Over larger distances, the biotite age values vary smoothly, only the Insubric and the Simplon-Centovalli lines separating areas with different biotite ages. Biotite age boundaries, with the exception of the two faults, cut tectonic lines and nappe structures as well as metamorphic isograds, indicating once more their younger nature;
4. under lower grade metamorphism, micas and feldspars, even adularia from open fissures, may take up excess argon; this tendency can still be observed in the outer staurolite zone. In samples which did not reach Sr-homogenization in Alpine time, the Rb-Sr biotite age calculation might suffer from the unknown composition of the incorporated strontium; apart from these two exceptional cases, Alpine Rb-Sr and K-Ar ages on biotite are identical;
5. white micas, muscovites and phengites, show a much higher stability of the Rb-Sr system than biotite; the border between Alpine and pre-Alpine white mica Rb-Sr ages runs parallel, somewhat outside the staurolite-chloritoid reaction line. No intermediate age zone exists for white mica, because neoformation of white mica may occur without disturbing the Rb-Sr system of older white mica; Alpine and pre-Alpine white mica Rb-Sr ages can be found even in different regions of single mica crystals;
6. in zones of lower grade of Alpine metamorphism surrounding the high grade metamorphic area from west and north to the east, Rb-Sr ages of 35–38 Ma have been found on white micas, especially on phengites; this date coincides with the supposed time of the climax of the Lepontine orogenic phase;
7. within the zone of Alpine staurolite white micas generally give Alpine ages, which are higher than the biotite ages; their regional variation shows a similar trend to the biotite age distribution and also some dependence on the tectonic position, rocks from higher elevation yielding older white mica Rb-Sr ages, indicating early updoming which did not last to the time of biotite ages;
8. young Alpine K-Ar ages from white micas lie between the biotite and the Rb-Sr white mica ages, nearer to the biotite ages.

Based on these regularities in the regional mica age distribution and making the necessary assumption, that the process which opens the Rb-Sr and K-Ar systems during progressive metamorphism is the same for the systems closure in retrograde conditions, a model for the interpretation of mica age data has been established. Although this model holds for rocks of granitic composition, it neglects the possible local influence of other parameters, such as fluid phases and extreme stress conditions. However, compared to models derived from contact-metamorphic zones, this concept has the advantage of direct relation to conditions of regional metamorphism.

The main parameter which controls mica ages must be temperature; above a certain temperature range, the K-Ar and Rb-Sr clocks are reset, K-Ar and Rb-Sr ages are 'rejuvenated'. By comparing the zones of mica rejuvenation to temperatures of metamorphic reactions, the temperatures for mean rejuvenation, the 'blocking temperatures' for the K-Ar and Rb-Sr systems in micas, have been derived (Purdy & Jäger 1976). A blocking temperature of 300°C is proposed for the K-Ar and Rb-Sr system in biotite. Since biotite does not form below this temperature, the K-Ar and Rb-Sr biotite ages must always be considered as cooling ages to 300°C. The Rb-Sr system in white mica is more stable than in biotite, with a proposed blocking temperature of 500°C for Rb-Sr in muscovite and phengite. In areas of high grade metamorphism where metamorphic temperatures exceeded 500°C, white mica ages date the cooling to this temperature. Since white mica can form helow 500°C, white mica formation can be dated. For K-Ar white mica dating a blocking temperature of 350°C has been proposed. In most cases K-Ar white mica ages must therefore be interpreted as cooling ages. For all these mica blocking temperatures, uncertainties of ± 50°C have been estimated.

On the basis of temperature estimates from metamorphic reactions in sediments, and on ^{39}Ar-^{40}Ar measurements on illites from the Central Alps, Frank & Stettler (1979) support the interpretation of white mica K-Ar data and the proposed blocking temperature of 350°C. On many of the previously dated samples from the Central Alps, apatite fission track data have been reported (Wagner *et al.* 1977). Based on annealing experiments, Wagner *et al.* conclude a blocking temperature of 120°C for apatite fission track data and rates of 1–10°C/10^5a for Alpine cooling data from bore hole samples supporting this blocking temperature (Naeser 1979). Apatite fission track ages are in all cases lower than biotite ages, but the regional age distribution in the Central Alps is the same for both methods.

Alpine metamorphism and subsequent cooling

Rb-Sr whole rock analyses (Dal Piaz *et al.* 1978) show that early subduction and high pressure metamorphism has occurred already 130 Ma ago in the Western Alps. Many age results in the Eastern and Western Alps indicate Cretaceous metamorphism, evidence of high pressure metamorphism being well preserved in the Western Alps. Even in the deep Pennine area of the Central Alps with the intensive Tertiary (Lepontine) overprint, we assume Cretaceous thrusting and metamorphism; the many eclogite occurrences in the Ticino area might represent remnants of the Cretaceous high pressure metamorphism. Although many ages in the range from 60–120 Ma have been obtained in the Alps, their interpretation is still uncertain and it is therefore impossible to reconstruct the p,T- path for the early Alpine history in Cretaceous time.

In the Central Alps, the different stages of the Tertiary Lepontine metamorphism and the subsequent uplift and cooling may be reconstructed using different methods: the temperature maximum resulting in white mica recrystallization has been dated from low grade metamorphic rocks at 35–38 Ma (Rb-Sr phengite); the cooling history of several regions from the Central Alps has been reconstructed by relating all the available age data to the different blocking temperatures, indicating time and rate of cooling from highest to lowest blocking temperatures (i.e. from 500°C to 120°C). Apatite fission track ages show a positive correlation between sample elevation and age and thus the relation elevation- versus fission track age-difference is a measure of the uplift rate (more precisely the rate at which the 120°C isotherm passes the sample locations), or the erosion rate which brings the sample about 4 km below the surface, assuming a geothermal gradient of 30°C/km. Thus, by applying the different dating methods, it has been possible to compare the cooling time and the cooling rate for several regions of the Central Alps with the time and speed of erosion and uplift; high erosion rate must be caused to the main extent by high uplift rate (Wagner *et al.* 1977).

Consistent results have been found for the different regions of the Central Alps: early, quick cooling and uplift in the eastern part, the Bergell area, where biotite ages range from 18–26 Ma, the cooling rate and erosion speed decreasing with time; in contrast, in the western part, the Simplon area and the western part of the Aar massif, late cooling and uplifting is found, with biotite ages of 10–12 Ma, and an increasing cooling rate. In the Simplon area, the recent uplift rate of 1.6 mm/a is higher than at any time in the last 25 Ma (Pavoni 1979). Different cooling and erosion rates are also found in the north–south profile through the Central Alps, with earlier cooling in the south, near the Insubric line. This model, with the differential cooling and uplift history of the Central Alps, is self consistent: many observations can now be explained, several theories supported, some disproved — but nowhere does the model contradict the geological evidence. However, the common opinion, that the Alps formed a high mountain chain in Oligocene time, were then eroded in Miocene to a hilly landscape and started to grow again in Pliocene time, is not supported by our results. Recent palaeobotanical and geomorphological studies show that this simple geomorphological picture is not correct. Boulders in foreland sediments witness a complicated history of uplift and erosion, quite different for the various regions of the Central Alps.

A key to the uplift history of the Central Alps is the Bergell intrusion, which presents a good example of the fine time resolution of geochronological data. The Bergell magmatic rocks, granites, granodiorites and tonalites, cut the Pennine nappe structures. Boulders of Bergell tonalites and granodiorites are embedded in Upper Oligocene–Lower Miocene conglomerates of the molasse near Chiasso–Como, 80 km to the south-west of the Bergell Alps. Near the Bergell plutonic rocks, the climax of the Lepontine metamorphism has been dated at 35–38 Ma. U-Pb ages of 30 Ma are reported for granites, granodiorites and tonalites, as well as granodiorite boulders in the molasse, (Gulson & Krogh 1973). This means that the Bergell plutonic rocks were intruded after the climax of the Lepontine metamorphism, when uplift and cooling had already started. Wagner *et al.* (1979), measured apatite fission track and mica K-Ar ages on Bergell granodiorites from the Alps and from molasse boulders. The boulders have yielded K-Ar ages of 29 Ma, within the error limit of the U-Pb zircon data of 30 Ma, indicating a very quick cooling rate during and shortly after the Bergell intrusion. The magmatic rocks in the Bergell area crop out within an area surrounded by three faults: the Insubric, the Muretto and the Engadine lines. Apatite fission track data from the area within the faults show a positive correlation with sample elevation and higher ages, 10–17 Ma, than those from rocks outside the triangle. This shows that the area bounded by these faults was squeezed up very quickly, earlier than the neighbourhood, within the framework of large-scale tectonics. The quick uplift rate must have caused the melting of the Bergell magmatic rocks. In contrast apatite fission track ages from the three analysed boulders are much higher- 23, 24 and 26 Ma, indicating their origin from a much higher level than the present outcrops, viz 7800 m, 8300 m and 9600 m respectively. Although these elevations are relative to the present surface, they indicate an original overburden in the Bergell area of about 10 km. The fission track data from the boulders give the time when rocks within the Bergell pluton cooled to 120°C, still at a depth of a few kilometres. The best time estimate for the molasse sedimentation is also 23 Ma (Rögl *et al.* 1975), which is identical with the lowest fission track age of the molasse boulders. This demonstrates the very short time involved from erosion of several kilometres of overburden to sedimentation. In the Bergell area, high mountains in Oligocene–Miocene time must have been created in association with this enormous uplift rate.

Amongst the molasse boulders near Chiasso, the Bergell rocks dominate; more than 90% of the giant blocks greater than 1 m^3 are Bergell derived rocks, predominantly granodiorites. These enormous boulders must have been transported, mainly by glacier, over a distance of 80 km. With palaeobotanical studies Hantke (1982), demonstrated that in Upper Oligocene time, the temperature must have decreased considerably. At this time, the high Bergell Alps with elevations of more than 6000 m, must have been covered by extensive glaciers. Hantke & Jäger (in press), give evidence for a glacial transport of the Bergell boulders of about 50 km. The further transport to the site of chaotic oceanic sedimentation (Gunzenhauser 1982) cannot have been by rivers, but rather floods, which presumably originated in the catastrophic bursting of moraine dams. Whilst the upper part of the Bergell intrusion was uncovered, eroded and transported to the south-west, the rocks forming the present day outcrops were maintained at temperatures above 300°C.

Outside the Bergell area, there is also strong sedimentological evidence for differential uplift. In Middle Miocene time, the main watershed in the Central Alps was much farther to the south, near the Insubric line and at this time, the high Western Alps with the Monte Rosa formed rather low hills (Hantke & Jäger, in press). This supports the evidence from the geochronological data which suggest the recent quick uplift of the Aar massif, the Simplon and the adjacent Western Alps, which resulted in the repositioning of the watershed to the north.

These examples show, that geochronological data, when supported by petrological and palaeobotanical studies in the foreland sediments, not only permit the unravelling of the different stages of metamorphism, magma formation and subsequent cooling, but are also essential in the identification of rocks now deposited in sediments. Further the present study shows how geochronological data, especially fission track results can be used to trace the paths of eroded rocks from their origin to their subsequent deposition. It thus provides an important contribution to the history of the Alpine landscape.

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Isotopic and palaeomagnetic evidence for rates of cooling, uplift and erosion

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SUMMARY: The use of isotopic and magnetic blocking temperatures to infer cooling history is reviewed in the light of kinetic models of the blocking process. Published isotopic estimates of cooling and uplift (erosion) rates are based upon (a) vertical gradients of fission track ages, (b) age differences amongst isotopic-mineral systems with different blocking temperatures and (c) kinetic analysis of ^{39}Ar-^{40}Ar step-heating experiments. The palaeomagnetic approach to cooling history depends upon thermal demagnetization experiments upon individual samples, from which different blocking temperatures can be assigned to different directions of magnetization having known age relationships. Successive magnetic reversals might in principle be used but only in areas of rapid postorogenic cooling. Where cooling is very slow it may be possible to use apparent polar wander paths to infer cooling rates, as in the mid-Proterozoic of West Greenland. Conversely, in the Grenville Province distinctive components of magnetization have been dated by reference to a cooling history based upon argon-39. In Scotland a study of imposed magnetizations in thermal aureoles of Tertiary dykes provided estimates of ambient temperature at the times at which they were intruded. Cooling (erosion) rates so derived vary bimodally from $\sim 30°C$ (~ 1 km) per million years over periods of $\sim 2 \times 10^7$ years immediately following orogeny, to $\sim 1°C$ (~ 30 m) per million years in cratonic areas over periods of $\sim 2 \times 10^8$ years, and broadly agree with sedimentological estimates.

Central to any discussion of thermal history, and how it affects the palaeomagnetic and isotopic records, is the concept of blocking temperature. In palaeomagnetic work this concept has played a fundamental role since the classic work of Néel (1949, 1955). It was introduced into geochronology somewhat later, to help to interpret mineral age patterns in regionally metamorphosed terrain (Armstrong 1966; Harper 1967; Jäger *et al.* 1967). At its simplest, the concept can be stated thus: for every palaeomagnetic or isotopic system, there is a critical temperature above which all evidence of its earlier history is continuously erased; as the system cools through that temperature it becomes stable (i.e. 'blocked') against thermal disturbance; a magnetic system records the prevailing magnetic direction at that temperature and a geochronological system records the time of cooling through it. Analysis of various geochronological systems with different critical 'blocking' temperatures should directly give time and temperature co-ordinates on a cooling curve; analysis of palaeomagnetic directions should give a temperature-direction relationship, which can be converted into a cooling curve if the history of the magnetic field is known.

Confusion sometimes arises from the use of the term 'blocking temperature' for two quite distinct situations: (1) the closure of a new system during initial cooling; and (2) the opening of an established system during a heating event. The two temperatures need not necessarily be the same. It is therefore convenient to adopt the term 'closure temperature' in our discussion of cooling isotopic systems. In our palaeomagnetic discussion, however, we retain 'blocking' because of its wide currency amongst palaeomagnetists.

Definitions of blocking (closure) temperature

Kinetic models of the closure process for geochronological systems will be adopted here; for some systems they are firmly established, and for others they are at least plausible. In such models the closure process cannot be instantaneous (Fig. 1). The hot, fully-open system, in which the daughter product is lost as soon as it is formed, changes gradually to a fully closed system, passing through a transitional temperature range in which the daughter product is partly lost, partly retained. Extrapolation of the low-temperature (fully closed) part of the growth curve back to the time axis, as shown by the dotted line, corresponds to a calculation of apparent age. A simple, unambiguous definition of closure temperature is therefore 'the temperature of the system at the time represented by its apparent age' (Dodson 1973). In principle, therefore, a knowledge of closure temperature for a particular isotopic system enables us to determine the temperature of the host rock at one moment in its history. Using several systems, with different closure temperatures, a more complete picture may be constructed.

Magnetic blocking temperatures of a slowly cooling system may be similarly defined (Fig. 2). At high temperatures the magnetization is in equilibrium and follows any change in applied field. In other words, the thermal fluctuations of the system can readily surmount the energy barrier to the change of direction of magnetization (the activation energy). In the transitional range of temperature, which is narrower than for isotopic systems, the magnetic system lags behind changes in the direction of the external field, though partial remagnetization occurs. Finally thermal fluctuations are diminished to a level at which the energy barrier can no longer be surmounted, and the mean direction of magnetization acquired during the transition is frozen in. Extrapolation back to the equilibrium curve gives the blocking temperatures as indicated.

In contrast to isotopic systems, a single magnetic mineral can vary widely in its blocking temperature, which depends, most importantly, on the shape and size of the grains. Most rocks contain a wide spectrum of blocking temperatures reaching up to the Curie Point of the magnetic mineral concerned (above the Curie Point atomic spin alignment breaks down and no ferromagnetism is possible). This spectrum of blocking temperatures means that the magnetization of a mineral phase may be composed of directions of magnetization from more than one geological event. For example, a rock which is cooling from high temperatures during a continuous change in direction of the ambient field due to plate movement may record this directional variation

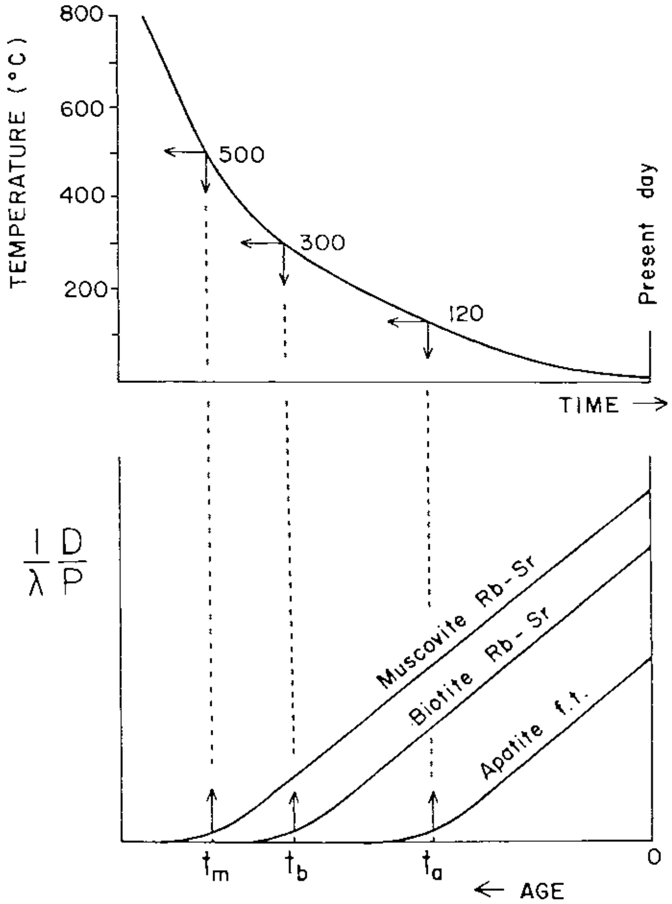

FIG. 1. Definition of isotopic closure temperature. Extrapolation of the straight-line portions of the growth lines back to the age axis corresponds to the process of calculation of the apparent age.

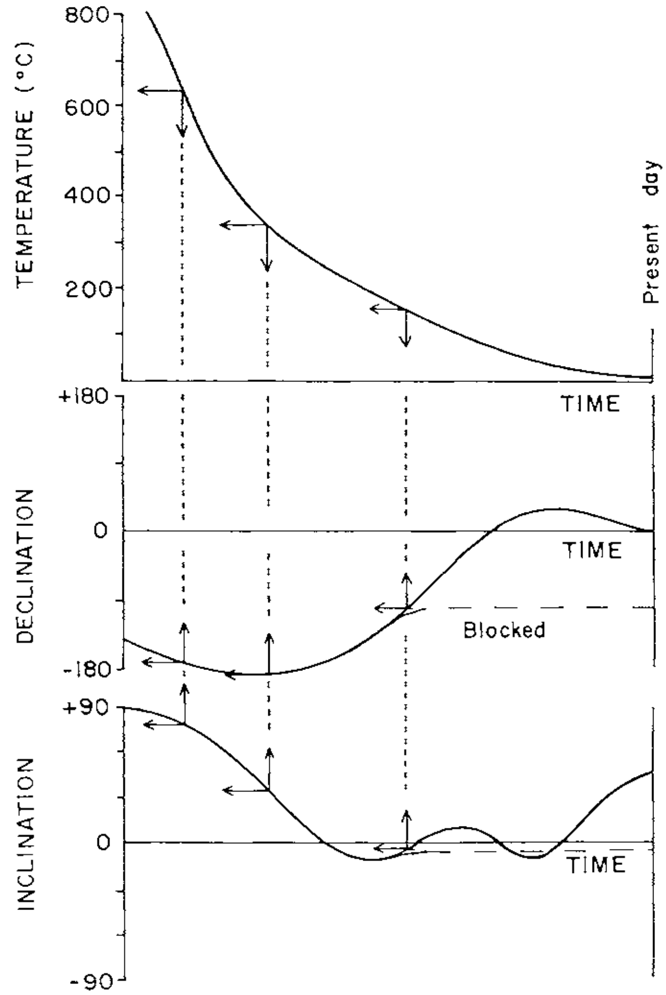

FIG. 2. Idealized behaviour of a magnetic system in a changing magnetic field in response to the same cooling curve as in Fig. 1. As the temperature falls, grains with successively lower blocking temperatures become blocked and record the two components (declination and inclination) of the direction of the ambient field at the time of blocking; three possible directions are shown. At first all grains follow the direction of the changing field shown by the solid curves: the horizontal broken lines show, for one blocking temperature, how the direction of magnetization diverges from the direction of the ambient field after blocking has occurred.

within a single sample. Alternatively, successive lower grade metamorphic events may be recorded as discrete directions, each relating to a single event, within the magnetization record of a rock; these are commonly termed 'multicomponent remanences'. If these changes of direction, discrete or continuous, can be disentangled from each other, and from non-thermal magnetizations, a single sample containing one magnetic mineral can provide information about a relatively complex thermal history (Fig. 2).

Theoretical background

Thermally activated processes such as magnetization of a ferromagnetic mineral, diffusion of a daughter isotope, or annealing of fission tracks, tend to follow some form of the Arrhenius relationship.

$$k(T) = k_0 \exp(-E/RT) \qquad (1)$$

where $k(T)$ is the rate parameter at absolute temperature T, E is the activation energy, and R the gas constant. The pre-exponential factor k_0 (sometimes known as the frequency factor) is the value of k at infinitely high temperatures.

E corresponds at least in part to the energy barrier which has to be surmounted for any discrete change to take place in the system. E/RT is generally much larger than 1, and the quantity $\exp(-E/RT)$ corresponds, crudely, to the very small probability that any particle of the system is able to cross the energy barrier during a time interval $1/k_0$.

Large values of E/RT mean also that thermally activated processes are very steeply dependent on temperature (commonly log(k) is plotted agianst T^{-1}). Thus the temperature interval for the transition from a fully open to a fully closed (blocked) system can be expected to be fairly small.

Theoretical work on the blocking or closure process (Dodson 1973, 1976, 1979; York 1978a, b; Dodson & McClelland-Brown 1980) has shown that, with certain approximations, a single equation can be used for blocking or closure temperature (T_c) in a wide range of kinetically controlled cooling systems, namely

$$E/RT_c = \ln (A\tau k_0) \qquad (2)$$

Here A is a geometry-dependent parameter, k_0 is the frequency factor from eq. (1), and τ is the so-called 'cooling time constant' — the time required for the value of $\exp(-E/RT)$ to diminish by a factor of e (2.73..) in the transitional temperature range. The underlying physical significance of this relationship was discussed by Dodson (1976, 1979).

For the present, it is important to note that the cooling time constant is inversely proportional to cooling rate: the slower the cooling, the larger the value of τ. In fact we can derive from (2):

$$E/RT_c = \text{constant} - \ln\,(\text{cooling rate}) \qquad (3)$$

For isotopic systems the constant term depends upon the particular mineral species, grain size and the dating method employed. For magnetic systems it depends upon the mineral species and is extremely sensitive to the grain size. The logarithmic relationship implies, however, that T_c is not very sensitive to cooling rate. In geochronological systems the change by a factor of 10 will alter T_c, typically, by 10° to 80°C depending upon E and T_c (Fig. 3). Smaller changes are normal in palaeomagnetic systems.

Application of equation (2) to geochronological systems depends on the availability of reliable kinetic data, which are rather scarce, though the situation is improving. The quantity k_0 is equal to D_0/a^2 for diffusion of a radiogenic isotope, 'a' being the effective grain size. Uncertainty in the latter quantity is a rather serious problem since it is commonly much smaller than the physical size of the mineral grains (diffusion processes in minerals have, in several instances, been shown to be not significantly speeded up by crushing until very fine powders are produced). An interesting possible way round this problem was demonstrated by Berger & York (1981) who calculated the kinetic parameters, D_0/a^2 and E, from argon-39/argon-40 step-heating experiments; the same experiments also gave the age of the samples being investigated.

Palaeomagnetic application of (2) is more complex because the activation energy varies with temperature and may be one of two types, depending on the magnetic structure of the grain concerned. Small grains possess a uniform magnetization which is generally constrained to either of two antiparallel orientations controlled by shape or crystallographic effects; these are called single domains. For larger grains this is energetically unfavourable and the magnetic structure breaks up into domains of different magnetic orientation; such grains are termed multidomains. The activation energy for single domains is the energy required to reorient the whole magnetization of the grain, whereas activation energy for multidomains is the energy required to move the boundaries between domains. In general, single domain grains carry the stable remanence which is analysed in palaeomagnetic studies. Dodson & McClelland Brown (1980) showed how the variation of single-domain activation energy with temperature could be taken into account, and presented curves relating thermal demagnetization temperatures, determined in the laboratory, to natural blocking temperatures for the same components of magnetization, assuming single-domain grains. The difference between the experimental and natural blocking temperatures diminishes towards zero as the temperature approaches the Curie point. The difference is also very insensitive to cooling rate, changing typically by ~ 10°C or less for an order-of-magnitude change in the latter. The relative contributions to the total remanence of components for various blocking temperatures will depend on the size distribution of the domains, but there are theoretical indications that magnetization in blocking temperature intervals near the Curie point will be relatively more intense.

The theory is at best a rough approximation to the complexities of behaviour of real rocks and minerals.

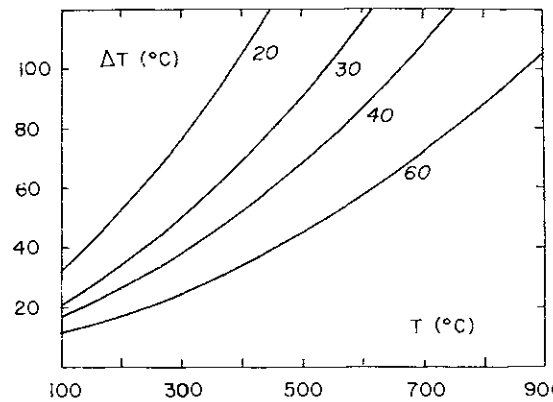

Fig. 3. Approximate change ΔT in isotopic blocking temperature for a × 10 change in cooling rate. ($\Delta T = 2.303\, RT^2/E$). Values of E on each curve are in kcal/mole.

However, from the evidence available, it seems at present to provide an adequate description of at least some natural systems.

Estimation of isotopic blocking temperatures

The 'Alpine' approach

Jäger *et al.* (1967) provided the first firm estimates of blocking temperature in micas dated by the Rb-Sr method. Their methods are summarized by Purdy & Jäger (1976), and by Jäger (this volume). In analysing the pattern of mica ages from the Central Alps they recognized that 'mixed' ages (from partially updated Hercynian micas) in the lower-grade metamorphic zones reflected the attainment of temperatures just high enough to open the system. From the association of mixed ages on biotites with stability of stilpnomelane they inferred a blocking (opening) temperature of 300° ± 50°C for Rb-Sr in biotite. Similarly, association of mixed ages on muscovite with the staurolite-chloritoid boundary led to an estimate of 500 ± 50°C for the Rb-Sr blocking temperature of muscovite. (They noted that for this mineral, unlike biotite, crystallization could in some circumstances occur at temperatures below the blocking temperature.)

Later, extending the work to potassium-argon, Purdy & Jäger (1976), from comparisons between the K-Ar and Rb-Sr age patterns, inferred K-Ar blocking temperatures of 300°C and 350°C for biotite and muscovite respectively.

While there seems no reason to believe that these figures are seriously misleading, since they give results for Alpine cooling history which are broadly consistent with other data, the assumed identity between opening and closure temperatures is not self-evident: kinetic models for loss of daughter product require that cooling rate and duration of thermal pulses be considered when relating closure temperatures to opening temperatures.

Kinetic approaches

Wagner & Reimer (1972) in their interpretation of fission track ages of Alpine apatites, pioneered the application of experimental kinetic data to calculation of geological closure temperatures. They treated fission track annealing as a first-order chemical reaction, and integrated numerically the differential equation for simultaneous generation and annealing of fission-tracks in a cooling system, showing that the resulting closure temperature is insensitive to large changes in cooling rate. Although the assumption of a first-order process is not really correct, and in fact no fully satisfactory mathematical treatment of the problem has been published, their results are supported by a calculation based upon physical arguments (Dodson 1979), and by the borehole measurements discussed below, all estimates being in the neighbourhood of 100°C.

The kinetic theory of closure temperature, given by Dodson (1973) for diffusing radiogenic isotopes, yielded for fine-grained Alpine biotites a result of about 300°C, in good agreement with the estimate of Jäger *et al.* However, an important objection to this result stems from the fact that very coarse (~10cm) Alpine biotites closed at the same time and temperature as the neighbouring fine-grained material. While there is no fundamental objection to believing the effective diffusion size to be independent of the physical size of the grains, it would be slightly surprising if it were identical with the grain size of the finer-grained samples.

The above-mentioned theory has been applied more widely in the last year or two. Berger & York (1981) used it extensively to calculate closure temperatures from ^{39}Ar-^{40}Ar step-heating experiments. Humphries & Cliff (1982) used it to estimate T_c for Nd in garnet (480–600°C, depending on garnet composition) and for several related geothermometers. Harrison (1981) applied it to argon in hornblende, and Harrison & McDougall (1982) to K-feldspar. The results obtained are broadly consistent with other observations when applied to cooling history (see below).

The Eielson borehole

Measurements of the age-depth relationship in a borehole drilled for geothermal investigation at the Eielson Air Force Base, Alaska, (Naeser & Forbes 1976: Naeser 1979), provide an (at present) unique approach to the direct measurement of closure temperature. A uniform age gradient for the apatite fission track ages, which decrease with increasing depth, suggests that the terrain has been eroded at a steady rate of 0.02 mm/a for the past hundred million years or so. Extrapolation of the line suggests that zero age would be obtained just below the bottom of the hole, at a temperature of 105°C, which must be the value of the closure temperature. The result is consistent with the slightly higher T_c calculated for Alpine apatites for higher cooling rates. Biotite ages also show a linear trend: extrapolation to zero age gives 220°C, again consistent with Alpine blocking temperatures when the slower erosion rate is taken into account.

It is to be hoped that similar measurements can be made on other boreholes in areas of steady uplift and erosion.

Summary of isotopic closure temperatures

Some current estimates of closure temperatures are displayed in Fig. 4. Published figures have been adjusted where necessary to a cooling rate of 30°C/Ma, in cases where the activation energy is known. Uncertainties in diffusion size (e.g. for the exsolution lamellae of microcline feldspar) and strong compositional dependence (e.g. for garnet) mean that much of the data should be regarded as, at best, tentative. Other estimates, however, such as that for fission tracks in apatite, can be considered to be reasonably well established. The numerous values of T_c for K-Ar systems found by Berger

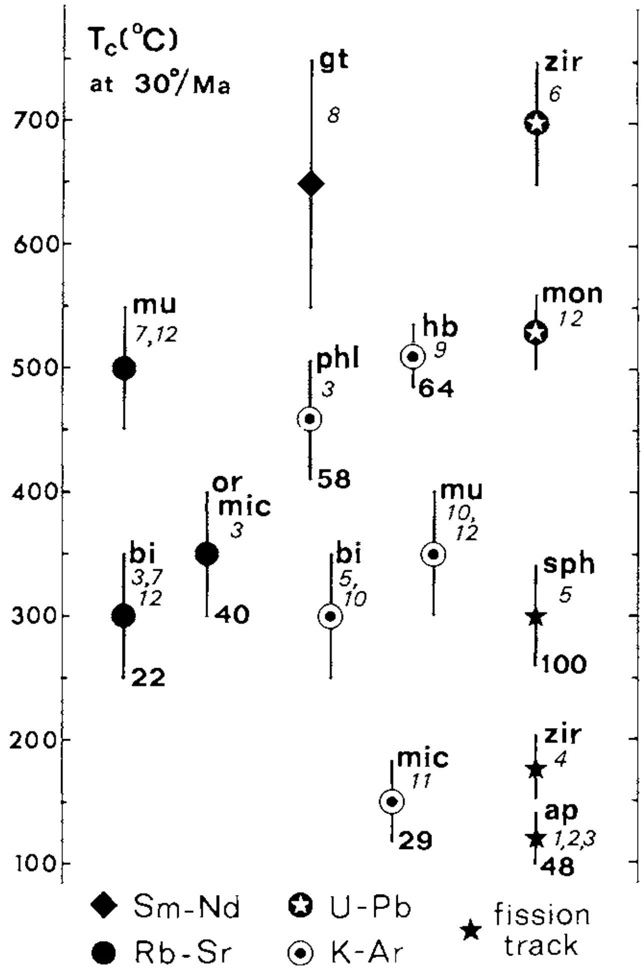

FIG. 4. Some estimates of closure temperature, standardized where possible to a cooling rate of 30°C/Ma using equation (3). Bold figures are activation energies in kcal/mole (at 50% annealing level for fission tracks). References (italic numbers), with diffusion size where diffusion models have been employed, are: 1. Wagner & Reimer 1972; 2. Naeser & Forbes 1976; 3. Dodson 1979 (a = 0.5 mm); 4. Harrison *et al.* 1979; 5. Harrison & McDougall 1980a (a not known); 6. Mattinson 1978; 7. Jäger *et al.* 1967; 8. Humphries & Cliff 1982 (a = 2.5 mm); 9. Harrison 1981 (a = 40 to 100 μm); 10. Purdy & Jäger 1976; 11. Harrison & McDougall 1982 (a = 3 μm); 12. Wagner *et al.* 1977; 13. Cliff & Cohen 1980.

& York (1981) have been omitted from the figure, for simplicity, but may be seen in Fig. 5. They seem to be generally rather higher than the corresponding results in Fig. 4.

Estimation of magnetic blocking temperatures

Laboratory estimates of the blocking temperature for a particular component of magnetization can be made by thermal demagnetization. In this procedure, rock samples are heated to increasingly higher temperatures and allowed to cool in zero magnetic field. During the heating, all magnetic grains which have blocking temperatures less than the maximum temperature reached will attain an equilibrium magnetization, i.e. they will be demagnetized. In the case of

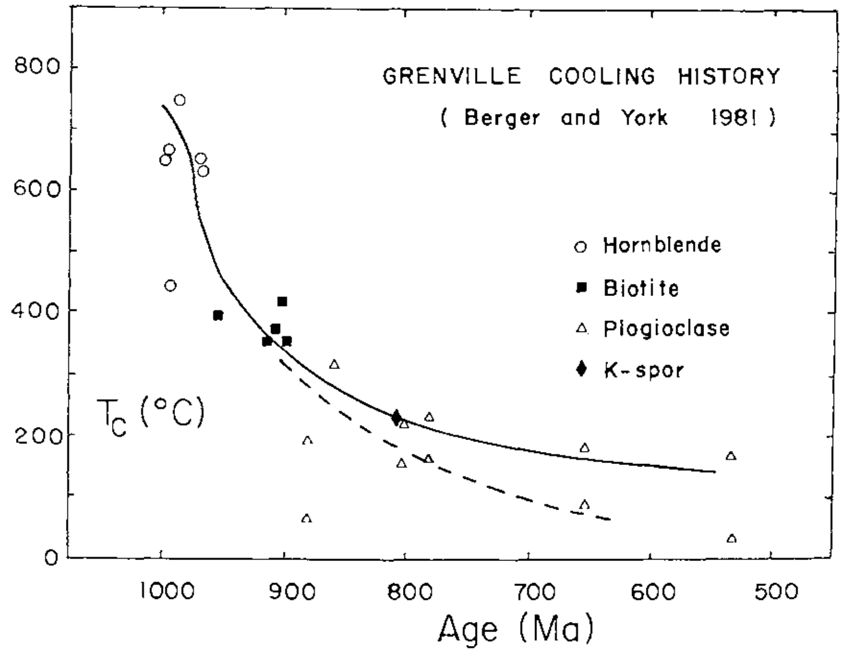

FIG. 5. Cooling curves for the Haliburton Highlands, Grenville Province, after Berger & York (1981). The solid curve is the preferred alternative. Cooling rates of 5°C/Ma (ages greater than 850 Ma) and 0.5°C/Ma (lower ages) were assumed, so closure temperatures need to be revised upwards for comparison with Fig. 4.

thermo-remanent magnetizations the original sequence of magnetization of the rock is reversed, and progressively older magnetizations are uncovered as the demagnetization temperatures are increased. At each heating step the magnetic vector which has been removed is found from the difference between the remanences before and after that step. The blocking temperatures thus obtained for various components of magnetization can be related to the corresponding acquisition temperatures by the theory discussed.

The highest magnetic blocking temperatures that could be recorded are the Curie points of pure magnetite and pure haematite, 575°C and 675°C respectively. In spite of theoretical suggestions that the dominant intensity will be acquired close to the Curie point, components having a wide range of blocking temperatures are commonly observed during thermal demagnetization of samples which contain a single magnetic mineral.

Isotopic estimates of uplift and erosion rates

For the present purpose we make the simplifying assumption that in the regions under consideration erosion and uplift rates are roughly constant and in balance. In such a thermal 'steady state' we have the relationship:

$$\text{Cooling rate} = \text{erosion rate} \times \text{thermal gradient}$$

Estimated thermal gradients can thus be used to relate cooling rates to erosion rates.

In some instances a more direct approach to erosion rates is possible. In the Eielson borehole, and in some mountainous regions, notably the Alps, ages determined by a single method show strong linear correlations with altitude. These linear age gradients are plausibly interpreted as reciprocal erosion rates; higher samples in the succession evidently passed through the closure temperature earlier than lower ones. In Alaska the inferred rate is 0.04 mm/a; in the central Alps Wagner *et al.* (1977) found 0.2 to 2 mm/a. In Colorado (Mt. Evans) the same kind of data suggests a sudden acceleration about 65 million years age from 0.02 mm/a to about 1 mm/a. The sharp change reflects the Laramide orogeny, and the higher rate is consistent with stratigraphic data (Naeser 1979).

The alternative approach, which is often more practicable, is to date from a single sample, or from a limited number of neighbouring samples, a range of mineral-isotopic systems with different blocking temperatures. The Alpine data of Jäger and co-workers is the most complete application of this approach to recent orogenic history, with results in the range 0.2 to 2 mm/a during the past 25 million years. Important differences amongst the uplift histories of different regions are observed, and the inferred recent rates are broadly consistent with the vertical gradients of apatite fission track ages.

This multi-method approach has been used by several groups to study post-orogenic cooling history over much longer periods. The results point to a quasi-exponential decay of temperature, implying that cooling rates diminish from very high values immediately following orogeny to much lower, barely measurable values, 10^8 to 10^9 years later. Berger & York (1981), applying the $^{39}Ar/^{40}Ar$ method to a variety of minerals from the Haliburton basin intrusions within the Grenville metamorphic terrain, found ages from 500 to 1000 Ma which they correlated with closure temperatures, individually estimated for each sample, of 100 to 750°C. The temperatures they obtained are shown in Fig. 5. For biotite and hornblende they are mostly rather higher than those given in Fig. 4, perhaps because of the breakdown of the structure of these minerals on dehydration when heating them *in vacuo*. However, there seems no reason to doubt the broad correctness of the picture which they present.

Harrison *et al.* (1979) find a similar quasi-exponential pattern for the Quottoon pluton (early Tertiary) in British Columbia, using in their case a wide range of methods (K-Ar, Rb-Sr, U-Pb and fission track) as well as of minerals. A slightly older pluton (Ecstall, 80 Ma) appeared, however, to have had its cooling disrupted by an early Tertiary heating event.

A further noteworthy example of the multi-method approach is the work of Harrison & McDougall (1980a) on the early Cretaceous Separation Point batholith in northernmost South Island, New Zealand. Here the cooling history of the batholith was used as a basis for studies of argon loss from hornblende in its thermal aureole. The inferred cooling curve, based on a thermal model fitted to certain key points from the isotopic measurements, suggests that uplift was discontinuous, so that the cooling rate dropped almost to zero before increasing again.

Palaeomagnetic evidence and cooling history

Grenville Province, Canada.

The first experimental evidence that magnetic minerals could record a cooling history that spanned a hundred or more million years, came from the Grenville structural province in the Canadian Precambrian Shield. The Grenville Province is of generally high metamorphic grade, normally amphibolite to granulite facies, and contains a considerable number of mafic intrusive rocks. Several workers have found that these intrusive bodies often carry two and occasionally three components of magnetization which define what is known as the Grenville loop in the apparent polar wander (APW) track. Until recently, the dating of this track has been problematical, since all the magnetization directions that form the path were acquired during cooling, and are difficult to date accurately. It had been considered that the Grenville track might be synchronous with the interior Laurentian Logan Loop (dated at 1000–1200 Ma) and that disparity between the two tracks reflected separate histories before plate collision caused the Grenville orogeny (Ueno *et al.* 1975; Buchan & Dunlop 1976). Alternatively, if the Grenville loop is younger than the Logan loop and both record movement of the same plate, then the sense of polar movement along the Grenville track and therefore the relative ages along the track were unclear (McWilliams & Dunlop 1978).

The joint isotopic and palaeomagnetic study by Berger *et al.* (1979) of which the isotopic aspects are discussed above, has calibrated palaeopoles from the Grenville Province and has enabled a good estimate to be made of rates of continental movement at the time of cooling after the Grenville orogeny. This study has also cleared up the controversy over two or one plate models of Grenvillia and interior Laurentia, allowing a single apparent polar wander track to be compiled (Fig. 6).

The study by Berger *et al.* focused on the Haliburton basic intrusions, whose palaeomagnetism had been studied in detail by Buchan & Dunlop (1976). It consists of three thermally acquired remanences, termed A, C and B, the directions of which change systematically with blocking temperatures, showing that they were acquired progressively during cooling. Buchan *et al.* (1977) have demonstrated that the A component was thermally acquired during cooling from the peak of the Grenville orogeny, and has laboratory blocking temperatures of 550–650°C carried by both magnetite and hemo-ilmenite. The B direction is also carried by two mineral fractions: one is a haematite fraction with blocking temperatures greater than 450°C (in some samples overlapping those of the A component) which is undoubtedly a chemical remanence produced below 450°C; the other is a titanomagnetite fraction, with laboratory blocking temperatures of 100–450°C. There is also a sparsely represented C direction

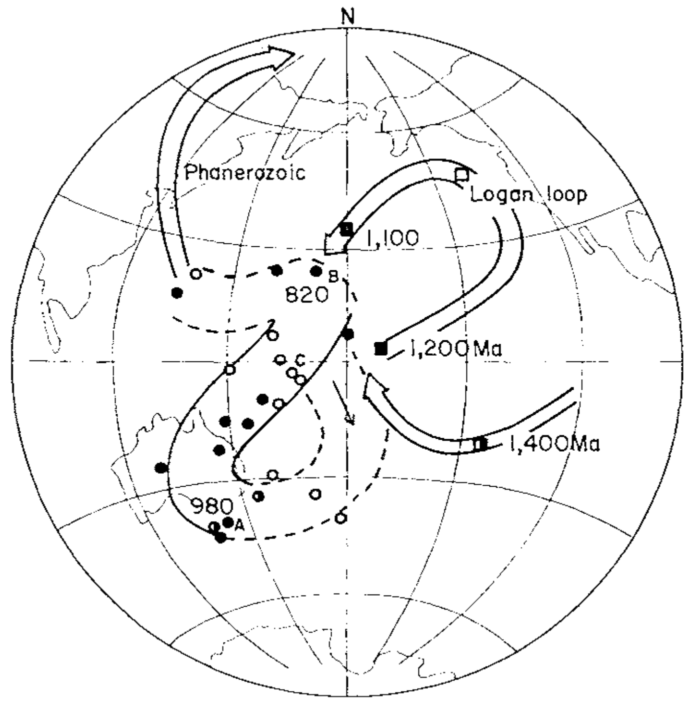

FIG. 6. Calibrated Grenville apparent polar wander path from Berger *et al.* (1979). All palaeopoles are from the Grenville Province, open and closed symbols representing opposite polarities.

which is opposite in polarity and intermediate in age between A and B.

To date directly the time of acquisition of the A and B components, the natural blocking temperatures at which the components were acquired must be estimated from the laboratory blocking temperatures. Theoretical relationships between natural and laboratory blocking temperatures have been given by Pullaiah *et al.* (1975) for static temperatures and by Dodson & McClelland-Brown (1980) for various cooling rates. The A component has a laboratory range of 550–650°C which, for a cooling rate of about 3°C/Ma obtained from the isotopically derived cooling curve, suggests (on either theory) that the magnetization was actually acquired between 520° and 650°C. Similarly natural blocking temperatures of up to 250° would fit the laboratory blocking temperatures of up to 450°C for the B component. Therefore Berger *et al.* assigned acquisition times of 980 ± 10 Ma for component A and 820 Ma for component B (see Fig. 5).

This produces a calibration of the Grenville APW path. In Fig. 6 the Laurentian APW path is continued into the Grenville loop which is well-dated from the assigned ages of components A and B. The other poles are all from the Grenville province, open and closed circles indicating opposite polarity. The rate of drift of 60° of arc in not more than 160 Ma gives a minimum average drift rate of 5 cm/a, which is plausible.

Given this calibration of the APW path for Grenvillia it should be possible to use palaeomagnetic evidence, similar to that discussed above, to infer cooling histories of other parts of the province. However, it should be borne in mind that only thermo-remanent components (TRM) can be used. Chemo-remanent magnetization has not in the past always been distinguished from TRM, giving misleading results (McClelland-Brown 1982).

The Nagssugtoquidian Mobile Belt, Greenland

A study by Morgan (1976) on a high grade Precambrian metamorphic terrain in West Greenland is probably the only study so far which has shown APW recorded continuously within a single mineral phase. The area is just within the early Proterozoic Nagssugtoquidian Mobile belt close to its gradational boundary with the Archaean craton to the south. The terrain comprises granulite and amphibolite facies gneisses cut by dolerite dykes. The change of ambient field direction during cooling from the peak of metamorphism is displayed as systematic movements of the magnetization vector in a southeasterly direction during progressive demagnetization (both alternating field and thermal). These cooling magnetizations are believed to have been acquired during a fairly late stage of uplift and cooling, long after the peak of metamorphism and perhaps close to the time at which K-Ar mineral clocks closed. K-Ar mineral ages for the Nagssugtoquidian belt as a whole range from 1650 to 1790 Ma.

A cooling rate for this metamorphic terrain has been estimated from Morgan's palaeomagnetic data by Dodson & McClelland-Brown (1980). The palaeomagnetic calibration necessary to obtain this cooling rate comes from the accepted Laurentian APW path for 1750–1500 Ma (e.g. Irving & McGlynn 1979) shown in Fig. 7. There is one well-dated and one less well-dated primary pole, supported by reversals and contact tests, which calibrate this track: the rest of the poles are overprinted magnetizations from the Canadian Shield and Greenland, whose ages are estimated from associated K-Ar dates of cooling. Morgan's linear trend of poles falls nicely on the Laurentian track when corrected for the drift of Greenland away from North America.

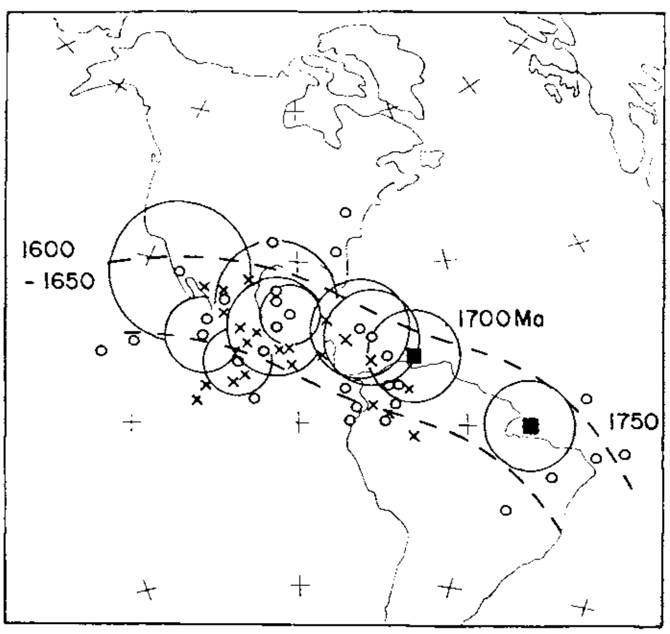

FIG. 7. Laurentian apparent polar wander path for the interval about 1750 to 1600 Ma from Irving & McGlynn (1979). Diagonal crosses represent the Nagssugtoquidian palaeopoles from Morgan (1976), open circles denote overprinted magnetizations from the Canadian Shield and Greenland and closed squares denote primary poles dated at 1700 Ma for the Sparrow dykes (McGlynn *et al.* 1974) and at about 1750 Ma for the Et-Then Group (Irving *et al.* 1972).

Before determining this cooling rate, it was essential to show that the systematic movement of the magnetization direction during progressive demagnetization is due to the recording of APW within the blocking temperature range of each specimen, since such movement might be due to the co-existence of two components of magnetization with different directions and overlapping blocking temperature spectra, which are removed unequally on demagnetization, causing the resultant vector to swing. There is little evidence for such a coexistence, where one component must have been chemically formed below its blocking temperature at a later time than the first component. Morgan & Smith (1981) have shown, however, that the stable magnetization is carried by a single mineral phase of fine-grained magnetite exsolved in the feldspars; furthermore, Morgan found a correlation between geographic position of a site and position of site mean pole, which would be unlikely if each pole were the combination of two components.

The method of calculation of cooling rate from the thermal demagnetization data was described by Dodson & McClelland-Brown (1980). A slight increase of the cooling rate, to 1°C/Ma, is now proposed, using a more accurate estimate of APW rates (40° of arc in 100 Ma) from Irving & McGlynn (1979).

Other studies of metamorphic terrains discussed above have shown that the effect of slow cooling is generally recorded in several discrete directions rather than in the continuous fashion inferred from Morgan's work. There are several possible reasons for this. One is that the temperature may well not decrease smoothly, but uplift and cooling may take place intermittently. Secondly, some components may reflect chemical changes to the magnetic minerals, altering their blocking temperatures, rather than thermo-remanence. Such a change may occur as a kind of retrogressive metamorphism which may not be accompanied by obvious changes in other mineral assemblages. There may also be a problem in interpretation, since palaeomagnetists, given a continuum of directions, may be tempted to split it into several discrete directions: the rather wide temperature steps which are usual in thermal demagnetization make it easy to quantize the results in this fashion.

Magnetic reversals and slow cooling

The examples from the Grenville Province and south-west Greenland, discussed above, show very long periods of a single polarity. Many comparable observations have been made elsewhere in the Proterozoic of the Canadian Shield and Greenland (Irving & McGlynn 1979; McWilliams & Dunlop 1978), of the Baltic Shield (Poorter 1975), and of Southern Africa (Morgan & Briden 1981). Either the reversal frequency was much slower at that time than in the Phanerozoic, or one polarity was dominant, periods of the opposite polarity being too brief to be recorded. The question remains, what would be the effect of more frequent polarity reversal on the magnetic record acquired during slow cooling? According to Beckmann (1976) repeated reversals might cancel out the recorded magnetization. From the theoretical work of Dodson & McClelland-Brown (1980) it is now possible to give a semi-quantitative answer in terms of the cooling time constant, τ.

For typical magnetic systems 100° to 200° below their Curie point, τ is the time taken for the system to cool through about 4°C. The transition interval from the 95% 'open' (superparamagnetic) to the 95% blocked condition is 3.5 τ corresponding to about 15°C drop in temperature*: if, in this interval, the field passes through several reversal cycles, one

*The figure given by Dodson (1973), 7.2 τ, is incorrect.

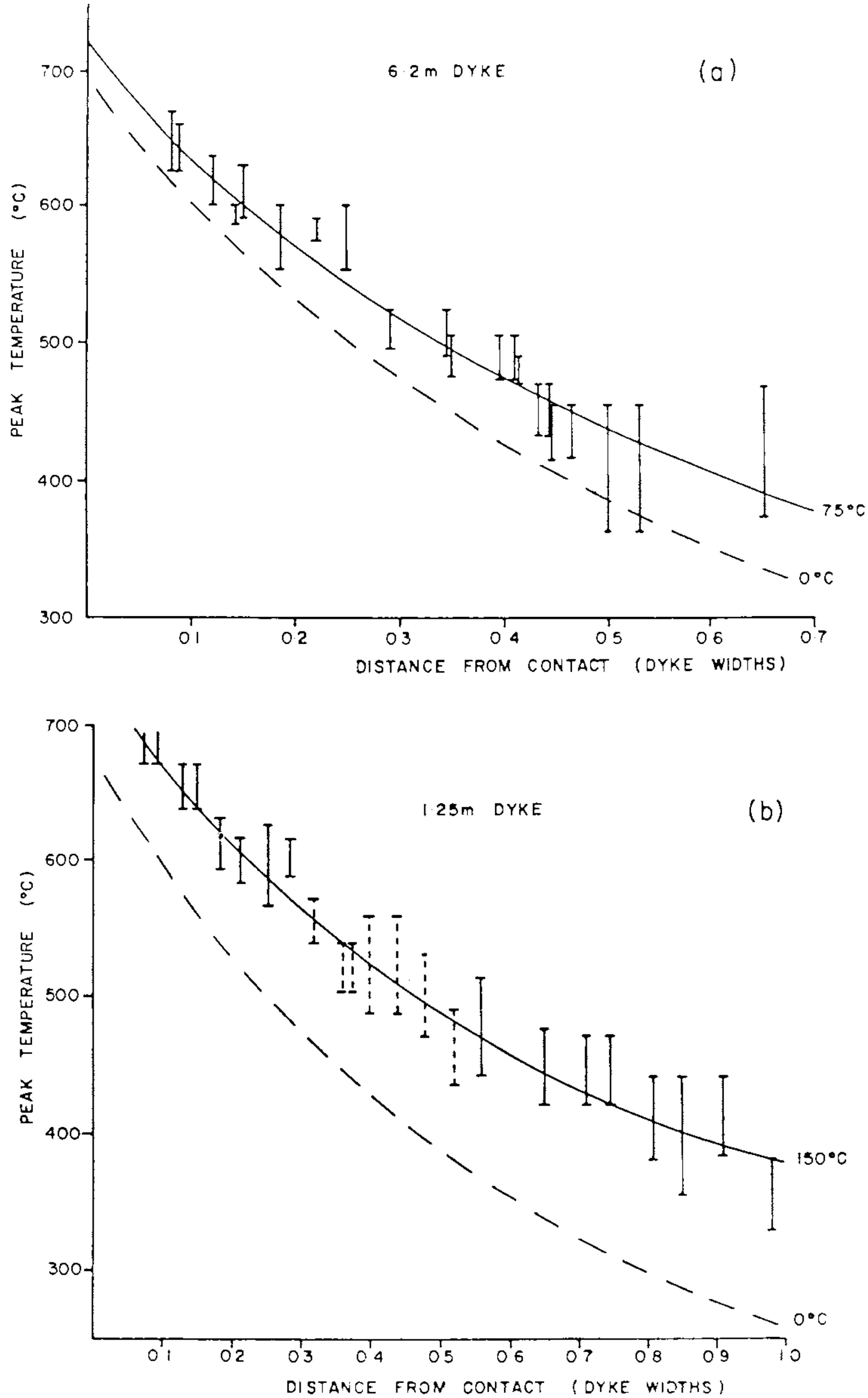

FIG. 8. Experimentally determined peak temperatures reached in contact heating due to Tertiary dykes. (a) 6.2 m wide and (b) 1.26 m wide. Experimental errors are shown as vertical bars. The dashed curve is the peak temperature calculated from conductive heat transfer assuming the country rock was at 0°C immediately prior to the intrusion. The solid curve is the peak temperature calculated for (a) 75°C and (b) 150°C country rock ambient temperature levels. From McClelland Brown (1981).

TABLE 1 Estimates of regional erosion rates

Region	*Rate* mm/a = km/Ma	*Period* Ma BP	*Approach*	*Reference*
Cratonic areas				
Alaska (Eielson well)	0.04	100 to 0	Isotopic	Naeser 1979
Colorado (Mt. Evans)	0.02	130 to 65	Isotopic	Naeser 1979
Scourian, NW Scotland	0.05* to 0.15	2660 to 2490	Isotopic	Humphries & Cliff 1982
SW Greenland	0.03*	1750–1650	Palaeomag.	This paper
Great Plains	0.046	150 to 0	Sedimentol.	Menard 1961
Orogenic areas				
New Zealand (Separation Point)	2*	114–108	Isotopic	Harrison & McDougall 1981
Central Alps	0.3 to 2	30 to 0	Isotopic	Wagner *et al.* 1977
Himalayas	0.2	40 to 0	Sedimentol.	Menard 1961
Himalayas	1.0	0	Sedimentol.	Menard 1961
W. Himalayas	2 to 15	5 to 0	Isotopic/ Palaeomag.	Raynolds & Johnson, this volume.

*Thermal gradient assumed to be 30°C/km.

would expect the two polarities to roughly cancel out; if, on the other hand, a single polarity is maintained, it should at the end be blocked against demagnetization by a subsequent polarity reversal. Given that reversal periods are typically of the order of a million years, one might therefore hope to uncover successive reversals by thermal demagnetization of rocks which have cooled at typical orogenic rates (30° to 60° per million years), but not at the very slow cooling rates found in Greenland and the Grenville province.

Progression of volcanic activity, Mull

The previous examples have concerned the relationship between rates of cooling and apparent polar wander recorded in remanent magnetism. An alternative approach utilizes the effect of an igneous intrusion on the magnetism of its host rock to estimate the ambient temperature of the host rock immediately prior to intrusion. This gives us an opportunity to look at much lower temperatures and much shorter time-scales than does the recording of apparent polar wander.

The intrusion will locally re-heat the host rock and this will partially reset the remanence of the host rock (or completely reset it if the Curie temperature has been exceeded). The peak temperature of the resetting will decrease with increasing distance from the contact and will reach the ambient temperature at a large distance from the intrusion.

A palaeomagnetic study of two intrusions of slightly different ages which indicate a rapid change in ambient temperature between their intrusion times is discussed by McClelland-Brown (1981), where the ambient temperature levels at the time of intrusion of two Tertiary basalt dykes from the Mull swarm in Scotland are presented. The two dykes are 25 km from the nearby Mull igneous complex and are intruded into Siluro-Devonian andesites which have a very stable magnetization. They are situated within a kilometre of each other and are intruded into the same structural level of the andesite pile. However, they are not contemporaneous since they are of opposite polarity. The Mull igneous activity has been dated at 58 ± 1 Ma by Sankey & Mussett (1979). Mussett *et al.* (1980) determined the duration of the activity to be less than 3.5 Ma, spanning a reversed to normal polarity transition followed by a normal to reversed transition. The contact aureoles of both these dykes show very nice partial remagnetization of the country rock; the maximum laboratory demagnetization temperature required to remove this remagnetized component decreases with increasing distance from the dykes.

Peak temperature profiles can be drawn up for both dykes. The two profiles in Fig. 8 show peak temperatures reached; they lie between the limits of the bars and are corrected for the difference between the duration of laboratory demagnetization and that of the Tertiary remagnetization. These are actual temperatures of reheating.

These results can be compared with profiles calculated from conductive heat transfer models for various levels of ambient temperature in the country rock prior to the intrusion. The experimental data evidently fits the conduction model best when the ambient temperature level is assumed to be 75°C ± 25°C for the normally magnetized dyke, and 150°C ± 25°C for the reversely magnetized dyke. The relative ages of the two dykes are not known but since the duration of the Reversed – Normal – Reversed sequence is less than 3.5 Ma the time between the intrusion of Reversed and Normal dykes will probably be less than 2 Ma. The two temperature estimates therefore indicate a minimum cooling (or heating) rate of about 30°C/Ma close to the Mull igneous intrusion. It is unlikely that this rate is solely an expression of uplift since rates of 1 km/Ma (1 mm/a) are highly improbable 25 km distance from the volcanic centre. The difference between the two estimates is interpreted to be due to the change in heat flux from the Mull centre as the igneous activity progressed.

Review of erosion rates

In Table 1 are collected the various values of erosion rates already discussed, together with other isotopic results and some sedimentological estimates. To convert cooling rates to erosion rates, a thermal gradient of 30°C/km has been assumed.

The estimate for the Archaean of NW Scotland is derived from the work of Humphries & Cliff (1982) on garnet Sm-Nd systematics. These authors' data suggest a time gap of about 170 Ma between the peak granulite facies metamorphism (drated by U-Pb on zirons) at a temperature estimated variously at 820°C to 1250°C, and the closure of garnets at perhaps 600°C. Thus we have a cooling rate in the range 1.4° to 4°/Ma.

The rate for the Cretaceous orogeny in New Zealand comes from the ages and closure temperatures given by Harrison & McDougall (1980a) for minerals from the Separation Point batholith. The zircon ^{206}Pb – ^{238}U age of 114 Ma, together with the K-Ar biotite age of 108 Ma, suggest a cooling from 700 ± 50° to 280 ± 40° in 6 Ma, a rate of about 70°C/Ma.

The sedimentological estimates, due to Menard (1961), are based upon observations of the total volume of marine sediments deposited in a given period by, for example, the Mississippi River draining the Great Plains. The low value for the Himalayas for the past 40 million years is probably distorted by palaeogeographic changes due to continental collision (Menard's paper predates plate tectonic theory). On the other hand, the very high value of 15 mm/a due to Raynolds & Johnson (this volume) is probably only of local significance, though 2 mm/a or more may be typical for this extremely active region.

Discussion and conclusions

The consistency in Table 1 amongst the various approaches to estimating rates of erosion is both striking and encouraging. Otherwise the most notable feature of the Table is the marked contrast between the low values for the 'stable' areas, and the one to two orders of magnitude higher values for orogenic areas, with no obvious intermediate estimates. This bimodality may be misleading, based as it is on rather limited data. However, it will be of great interest to discover, as more data of this nature becomes available, whether orogenic uplift normally slows down in a quasi-exponential manner, or whether some kind of 'braking' occurs leading to a sharp reduction in the uplift rate. For the stable areas the further question arises, are the results in Table 1 really typical? If so, we have to face the major problem, what mechanism could sustain such rates over periods of 100 to 200 million years, resulting in the removal of several kilometres of crustal material?

In conclusion, useful estimates of rates of cooling, uplift and erosion may be obtained from both isotopic and, to a lesser extent, palaeomagnetic data, giving results in accordance with sedimentological estimates. The most reliable results are probably those based upon steady-state vertical age gradients, since they are independent of kinetic models and estimates of past geothermal gradients. More measurements of this kind ought to be made in locations where there are reasonable grounds to suppose that a steady-state has existed for sufficient time. Magnetic methods have shown sufficient promise to justify further application, using rates of polar wander in regions of slow cooling and magnetic reversals in regions of very fast cooling. Finally, both ^{40}Ar-^{39}Ar step-heating experiments and the 'multi-method' isotopic approach have a continuing role to play in elucidating cooling and uplift history in at least a semi-quantitative manner.

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Processes and problems in the continental lithosphere: geological history and physical implications

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SUMMARY: Systematic changes in surface continental patterns of geology, tectonics and magma types reflect the unidirectional non-uniformitarian evolution of the continental lithosphere upon which the dominant constraint has been progressively declining terrestrial heat production. Widespread shallow asthenospheric mantle precluded the early stabilization of continental lithosphere, and the subsequent rapid growth of late Archaean continental crust implies considerable differences in the long-term thermal histories of oceanic and continental lithospheres. Proterozoic continental lithosphere rapidly became thick and refractory, though internally ductile, whilst oceanic lithosphere, the focus of terrestrial heat flow, remained relatively thin and plastic. This physical contrast probably has decreased in more recent times, a trend which may be coupled with the increased evidence for continental fragmentation in the Phanerozoic record. Finally, it is argued from thermal considerations, coupled with freeboard, isotopic and geochemical data, that continental growth has continued at a decreased and decreasing rate throughout post-Archaean time.

The widespread acceptance during the last 15 years that most contemporary geological patterns and processes are a natural consequence of lithospheric plate motions has hastened attempts to record and interpret ancient crustal patterns with the aim of seeking a fuller understanding of crustal evolution. Several major volumes of collected works (e.g. Kröner 1981; Moorbath & Windley 1981) represent significant advances in documenting the geology and geochemistry of Precambrian terrains, yet debates continue over interpretive aspects, such as the lateral extent, thickness and mobility of ancient lithosphere, the nature and timing of continental growth and recycling processes, and changes in the rates and sources of magma production. Whilst there is good agreement that systematic temporal variations in the geological patterns reflect long-term changes in causative processes, attempts to modify the modern plate-tectonic approach to account for these variations have met with less accord. This is largely because geological evidence for lateral accretion and collisional events, for example, in the form of linear calc-alkaline magmatic arcs and ophiolites in crust older than about 1000 Ma, has been difficult to find (see Gilkson 1980 for a review). It is to the 78% of Earth history before this time that contrasting evolutionary models apply. Against this background, the brief provided for this contribution to '*The Chronology of the Geological Record*' was that of evaluating evidence for the nature and rates of crustal processes, past and present, with particular reference to the links between magmatism and continental growth. Enough has been said already to indicate the uncertainties in tackling this problem; nevertheless, by combining information from several interdependent sources, some useful constraints can be applied.

Thus, in this paper, a summary of the key geological elements observed in continental crust of different ages is coupled with analytical, experimental and isotopic geochemical data, with evidence for the constancy of continental freeboard and, finally, with important implications derived from analysis of the Earth's thermal evolution to comment, largely qualitatively, on aspects of continental evolution. It will be concluded that, since about 3800 Ma ago, *decreases* in the rate of magma production and in the internal temperatures and ductility of the continental lithosphere have been accompanied by *increases* in the lateral and vertical dimensions of lithospheric fragments and possibly in the mobility of those fragments. These conclusions require (cf. Chapman & Pollack 1977; Bickle 1978) that the creation and destruction of oceanic lithosphere has provided the dominant mechanism for heat loss throughout the Earth's history.

Some implications of continental geology

From radiometric studies of the oldest provinces it is now clear that no continental crust older than c.3800 Ma is likely to be found (Moorbath 1977; Moorbath & Taylor 1981). This lack of direct evidence has frustrated attempts to do more than speculate about conditions in the pre-Archaean lithosphere for, almost certainly, such a solid 'crust' must have formed soon after accretion of the Earth. Nevertheless, by extrapolating the Earth's heat production-heat flow regime back to pre-Archaean times (an argument developed more fully below) and by applying to the Earth the well-documented impact stratigraphy of other terrestrial planets and the Moon, a few points of consensus have emerged.

The *average* thickness of rigid lithosphere, the outer heat-conducting layer, is inversely related to surface heat flow, thence to heat production, which was *at least* three times (Lambert 1980) and probably close to four and a half times its present average value (McKenzie & Weiss 1975) 4500 Ma ago. Thus the primitive lithosphere probably resembled, in *average* composition and thickness, modern lithosphere of ocean ridge zones. Such an unstable lithosphere would founder rapidly, or be subducted on cooling to higher density, into low viscosity shallow asthenospheric mantle, a process aided dramatically by the many thousands of major impact events at the Earth's surface during its first 700 Ma (Wetherill 1977; Grieve 1980). The process would be accompanied by degassing of volatiles, rapid re-equilibration and, almost certainly, complete re-homogenization of basic composition 'crust' with the mantle. Stable thermal conditions required to produce silicic partial melts in the pre-Archaean Earth (cf. Smith 1981) would occur only rarely within this primitive lithosphere. The lack of any geological or radiometric evidence probably indicates, therefore, that such pre-Archaean silicic material as was differentiated, must have been recycled through the mantle and, possibly, also rehomogenized: there was no permanent pre-Archaean continental crust.

Although, today, Archaean rocks (3800 to *c.*2500 Ma in age) are exposed over 20–30% of the continental area, in volumetric terms they may be much more significant. It is agreed that 70–100% of the present continental crustal volume had appeared by the end of Archaean times; at least some part of the younger crust represents recycled Archaean material (see Armstrong 1981 and discussion below). It is interesting to note that, despite the almost exponential growth in isotopic studies of Archaean crust, the 3800 Ma Amitsoq gneisses of Greenland have, since 1975 (Moorbath 1975) remained the oldest dated terrestrial rocks. Ages of 3600 Ma are known also from Africa, N. America and Scandinavia and, by 3000 Ma ago, every continent was recording geological activity, with a major 'peak' of Archaean ages in the 2600–2900 Ma time interval. Two major tectonic environments and petrological regimes are recognized (Windley & Smith 1976; Windley 1977): low-grade linear greenstone-granite belts and high-grade granulite-gneiss complexes. The latter are usually dome-like and range from equidimensional to linear on a scale from 10s to 100s of kilometres; they are surrounded by and occasionally grade into relatively undeformed homogeneous strata of the low-grade belts, but the features of adjacent high and low grade zones are so different that they must have formed in different tectonic environments (Windley 1981).

Occurring within the greenstone belts are both mafic to ultramafic mantle-derived volcanics and intermediate to acid calc-alkaline 'arc-type' intrusive and extrusive members. This blend of igneous rock types, together with their derived associations of immature clastic greywackes and conglomerates, led to the recognition of linear, synformal greenstone belts as probable Archaean ensialic marginal basins, developed on thinned lithosphere undergoing extension and later closure, above zones of active mantle diapirism (Tarney & Windley 1981). An alternative view of their origin, which dispenses with the need for lateral 'plate-like' migrations, also requires that the greenstone sequence formed as thick mafic lava piles were erupted over zones of mantle diapirism but, in this case, the basin would have sagged isostatically and remelted to form more silicic magmas (summary by Hargraves 1981). The latter view developed from the proposal (Fyfe 1978) that the *early* Archaean lithosphere was fairly uniform, globe-encircling and, therefore, (from considerations of conductive heat transfer) only 10–20 km thick. Yet it is now clear from studies of the mineral assemblages in *late* Archaean high grade complexes that silicic crust alone was between 30 and 80 km thick by 2800 Ma ago (Wells 1980). For this and other reasons discussed later, the present author (cf. Brown & Mussett 1981 Ch. 10) considers that, in relation to the average thickness of 10–20 km, a thinner oceanic–thicker continental lithosphere model provides the best fit for evidence from at least the *later* part of the Archaean record.

High-grade Archaean crust is dominated by 'grey' amphibolitic and granulitic gneisses, of tonalite-trondhjemite composition, together with layered basic assemblages and high-grade meta-sediments (see summaries in Barker 1979). Windley & Smith (1976) drew attention to the many similarities between these Archaean calc-alkaline orthogneisses and deeply-eroded modern continental arc batholiths; the main geochemical contrasts are the higher Na/K and Eu/Eu* of Archaean suites and the scarcity of intermediate andesite-diorite compositions (Barker & Peterman 1974). The source of tonalite-trondhjemite magmas with these characteristics was by partial melting of an amphibolitized mafic source, probably 'oceanic' crust sinking into the mantle in a relatively high temperature regime (Tarney & Saunders 1979) possibly almost vertically over down-going mantle convection cells as the andesite contribution from the 'mantle wedge' is absent (Barker *et al.* 1981). As with the interpretation of greenstone belts, again there is a note of controversy in determining the source of magmas represented in Archaean high-grade zones: some authors clearly see a need for lateral motions of small continental plates which therefore over-ride subducted fusible oceanic lithosphere whilst others prefer near-vertical sinking of mafic crust followed by partial melting.

Continental crust of Proterozoic age (*c.*2500 Ma to 600 Ma in age), surrounding and covering ancient Archaean cratonic nuclei, is widely preserved in every continent and records a diachronous change from vigorously-active to more stable tectonic conditions. From the geological record, the lateral dimensions of continental masses had increased from hundreds to thousands of kilometres, but these extensive stable platforms are traversed by linear zones along which Proterozoic tectonic activity was focused. These zones may comprise transcontinental basic dyke swarms (e.g. that of the N. Atlantic region, Escher *et al.* 1976), anorthosites, alkaline intrusive complexes and kimberlites where the crust was clearly under tension (Emslie 1978; Windley 1981), and the more controversial geosynclines and mobile belts containing thick sequences of highly deformed and metamorphosed trough sediments, calc-alkaline intrusives and, at high levels of exposure, andesite-rhyolite lavas. The origin of Proterozoic geosynclines and mobile belts has been widely debated: models range from those involving ensialic basement reactivation associated with incipient continental rifting or thinning (e.g. the mid-Proterozoic mobile belts of South and Central Africa, Kroner 1979; the early Proterozoic Labrador geosyncline, Dimroth 1981) through those involving localized continental 'plate-jams' due to collision in the absence of subduction (e.g. Baer 1981 for the Grenville province) to models with major horizontal plate displacements (e.g. evidence from the 2000 Ma old Coronation geosyncline, Hoffman 1973, and the much younger Pan-African belts of Afro-Arabia and S. America which developed on Proterozoic continental margins, Gass 1982; Shackleton 1979). Palaeomagnetic evidence bearing on this problem (e.g. Piper 1976; Irving & McGlynn 1976, 1981) indicates that any openings and closings across Proterozoic mobile belts must either have been small or rapid; significantly, however, movements of the major continental blocks were taking place during Proterozoic times (Irving 1979, and Dunlop 1981 estimated 4–6 cm a^{-1} for N. America). Thus, whilst the Proterozoic continental lithosphere was mobile on a large scale and accreted juvenile material at its margins, there is much evidence that it was easily susceptible to rifting and ductile stretching prior to orogeny (cf. Wynne-Edwards 1976; Watson 1978) perhaps not having gained everywhere the thickness, strength and rigidity of today.

The calc-alkaline granitoid rocks of Proterozoic orogenic zones closely resemble the gabbro-diorite-tonalite-granodiorite-granite suites of modern continental arcs rather than the tonalite-trondhjemite suites of Archaean times (see reviews by Barker *et al.* 1981 and Brown 1981). There is a growing consensus that magmatism resulted mainly from partial melting processes analogous to those of today at convergent oceanic-continental plate boundaries, a model which receives support from the space-time characteristics of these intrusive rocks and from isotopic data. Juvenile Nd and/or Sr initial ratios have been determined for the early Proterozoic calc-alkaline intrusives of SW Finland, S. Greenland and New Mexico (respectively, Arth *et al.* 1978; Van Breemen *et al.* 1974; Condie 1978) and for the late Proterozoic granitoids of Arabia (Duyverman *et al.* 1982). Significantly, the latter authors also showed that the Damara mobile belt of Nambia, an apparently ensialic zone of basement reactivation, does indeed contain granitic rock types derived from dominantly crustal sources. One might suppose, therefore, that the opposite extremes of horizontal displacements with ocean lithosphere subduction at continental plate margins and mainly vertical displacements with some crustal stretching, then thickening in ensialic orogenic zones are equally effective in producing mobile helt features.

As with the 5–10% of alkali-rich silicic magmas that occur at modern extensional plate margins, so the *c.* 1500–1900 Ma old anorogenic Proterozoic alkaline intrusive suites and associated anorthosites of Fennoscandia, Greenland and central N. America (reviewed by Emslie 1978) are thought to represent a major period of lithosphere stretching and generally abortive fragmentation. They are mentioned here because they were important new additions to the continental crust; the volumetric ratio of juvenile extensional alkaline magmatism to compressional calc-alkaline magmatism reached a peak of *c.* 1:3 in mid-Proterozoic times (refs. and discussion in Brown 1981).

The geological and tectonic patterns preserved in continental crust clearly testify to progressive unidirectional change, an evolution which, to a greater or lesser extent (depending on individual experience and prejudice), may be regarded as predictable on uniformitarian grounds. No matter what their individual stance in discussing the earlier Precambrian geological record, almost all the authorities cited above are agreed that plate tectonics, with the familiar process of ocean lithosphere subduction, had become established by 1000 Ma ago. The Proterozoic–Phanerozoic boundary is not marked, therefore, by profound tectonic change; the Mesozoic and Cenozoic style of activity along arcuate ocean-ocean boundaries and linear ocean-continent, arc-continent and continent-continent collision zones transgressed this boundary. Thus, for example, Palaeozoic mobile belts such as the Caledonide–Appalachian orogen and the Variscan fold belt are easily interpreted as collision zones whilst recent Tibetan intra-plate deformation provides a model for the late Proterozoic Grenville province (Dewey & Burke 1973; Baer 1981). Such correlations between the processes that formed ancient and modern orogenic belts may be extended further back in time if one accepts Windley's (1981) suggestion that the more deeply-eroded examples must lack high-level volcanics, ophiolites and nappe structures but expose, instead, a greater proportion of continental basement rocks.

Like their deeply-eroded ancient equivalent, Phanerozoic magmatic arcs contain abundant calc-alkaline rocks. They are dominated isotopically by a recent mantle imprint and it is widely accepted that, for the most part, they represent *new* additions to the continental crust (see discussion below). The abundance of alkaline rift-zone intrusions has diminished, however, since mid-Proterozoic times, perhaps reflecting the decreasingly ductile nature of the continental lithosphere. Of course, the continents still undergo rifting, splitting and fragmentation, but today, such processes are concentrated in regions where slow moving or stationary lithosphere is penetrated by hot rising plumes of anomalous mantle material (e.g. the NE African triple junction — Gass *et al.* 1978). This trend towards decreasingly active continental interiors, which is taken to indicate a progressive increase in lithosphere strength and thickness, is one of the more striking features of the geological record. Taken alone, however, it is clear that the record is able only to constrain rather than to argue a firm case for or against the different causative dynamic models that have been proposed. The following sections discuss different kinds of evidence which independently bear on this problem and which also help in understanding the long term growth and evolution of the continental lithosphere.

Continental accretion, recycling and growth

A combination of radiometric and geological evidence shows that orogenic activity has occurred repeatedly and episodically throughout the last 3800 Ma (e.g. Dickinson 1981, Fig. 1). Both the earliest dated activity and the subsequent orogenic record varies from continent to continent, though peaks of widespread activity occurred in the late Archaean (2600–2900 Ma), the late Proterozoic (800–1200 Ma) and, to a lesser extent, during early Phanerozoic times (400–600 Ma). In his extensive reviews of isotopic data from Precambrian rock units Moorbath (1977, see also Moorbath & Taylor 1981) interpreted these peaks in terms of 'accretion-differentiation' episodes, each of which contributed to the irreversible process of continental crustal growth. In other words, the isotopic evidence may indicate that there was no unique crust-forming event and, by implication, that the continental crust is still growing today.

Although few would argue that juvenile material has been added from the mantle to the continental crust throughout geological time (*crustal accretion*), attention is drawn below to arguments that a return flow of continental material back into the mantle (*crustal recycling*) may, since Archaean times, have balanced or exceeded the volumetric addition of new material. The latter view turns the statement above into an important and widely debated question: 'is the continental crust growing today?' The case for crustal recycling (Armstrong 1981) relies on (a) the subduction of continent-derived sedimentary material as layer 1 of oceanic lithosphere and the hydrothermal fixation of soluble elements from continental weathering into subducted oceanic crust (Fyfe 1982) and (b) constancy of continental freeboard coupled with a uniform crustal thickness since 2900 Ma ago (Wise 1974). Inferences concerning sediment subduction are based on observations of contemporary processes and are considered first.

Sediment subduction

Sediment subduction is supported by geochemical/isotopic data from some arc magmas which show spatial correlations, for example of K, Rb, Ba, ^{87}Sr and ^{207}Pb, with their concentrations in non-calcareous pelagic sedimentary material entering subduction zones (Karig & Kay 1981), and by volumetric data which show a deficiency of oceanic sediment accreted to some arc margins (data and references for Japan, the Marianas, Central America and Peru-Chile in Armstrong

1981). In a critical analysis of the fate of the 'missing' sediments, Karig & Kay (1981) concluded that the vast bulk of trench volcaniclastics and most of the underlying pelagic sediments not accreted laterally are 'subcreted' at shallow levels beneath the arc margin, a view supported by seismic reflection data from the Sunda and Cascade arcs. These authors envisaged that the 'magmatic production rate at arc systems far exceeds the limited (deep) sediment subduction rate': this deeply subducted material was considered to undergo ultrametamorphism and melting followed by only limited mixing between the fusion products and oceanic crust and mantle. The subducted volume returns immediately with interest. This view, which explains the geochemical correlations between arc magmas and subducted material for certain arcs (noted above) is in contrast to those of Armstrong (1981) and Fyfe (1982) who argued that, for most arcs, the geochemical signature of subducted sediments is swamped by a much more extensive crust-mantle fusion mixture, and also that much of the subducted volume passes through high-P refractory facies to be absorbed by the mantle.

Without labouring the point further, clearly, the volume of deeply subducted, recycled continental material is one of todays great uncertainties: whilst the volume might be negligible, an upper limit is set by the volume eroded from the continents each year (*c.*9 km^3 — Garrels & McKenzie 1971). If we accept that the majority clastic contribution is re-accreted, this leaves 0.75–1.0 km^3a^{-1} of pelagic material which may be accreted, subcreted, subducted and remelted, or absorbed into the mantle. Following this reasoning, Dewey & Windley (1981) estimated the latter at 0.44 km^3a^{-1} which they subtracted from their magmatic arc production rate (*accretion* in the terms of this paper) of 1.04 km^3a^{-1} to derive their preferred crustal *growth* rate of 0.6 Km^3a^{-1} during the Mesozoic and Cenozoic. Although a contemporary accretion rate exceeding 1 km^3a^{-1} may be rather high (e.g. Brown 1977 quoted 0.4 km^3a^{-1} which, conversely, may be an underestimate) the present author follows others cited above in finding no compelling case based on sediment subduction against limited growth ($\leq$ 0.6 km^3a^{-1}) of the continent volume today. Of course, this is much less than the time average growth rate of *c.*1.3 km^3a^{-1} over 3800 Ma — the causes and effects of more rapid growth in the past are considered later.

Continental freeboard

The second line of reasoning used to support a no-growth model for continental crust refers to the last 2900 Ma, during which time it appears (1) that the erosion level of Archaean cratons has remain close to the present level (Armstrong 1968; Wise 1974) and (2) that the average thickness of stable crust has not changed (Davies 1979) From these observations, and making the implicit assumption that the volume of water in the oceans has not varied, it is argued (see Armstrong 1981) simply that no long-term net gain or loss of continent or ocean volume can have occurred as this would have displaced sea-levels significantly. To explain the long continued reworking of continental crust through arc magmatism and orogeny since early Proterozoic times, this interpretation requires that recycling of continental materials balances accretion — new additions from the mantle to the crust can only be balanced by resorption of sediments into the mantle, a return to the case dismissed above.

Other processes which could help to explain the freeboard data are (1) an increasing importance of continental crustal sources for arc magmas with decreasing age and (2) increases in the volume of *both* the ocean basins and continents. Possible crustal melt sources for arc magmas are subducted continental detritus and the crust itself. The former can be dismissed as a major source since there is no volumetric correlation between magmatism and the amount of sediment available for subduction, whilst experimental data (considered separately below) show that melt contributions from within the crust itself are likely to have *decreased* with decreasing age.

The second process for maintaining freeboard — relative increases in the volumes of the ocean basins and the continental crust — is, in view of the declining vigour of internal Earth dynamics, a more logical proposal than that of zero growth with steady-state continental recycling. Reference to heat flow data compilations (e.g. Sclater & Crowe 1979) show that, today, between a third and a half of the Earth's heat flow is focused through the 15% of its surface represented by the youngest part of the ocean ridge zones less than 20 Ma old. The volume excess of all the ocean ridges over the level of abyssal plains, known from gravity data to be due to the relative thermal expansion of the young lithosphere (e.g. Talwani *et al.* 1965), can be estimated from the bathymetric — heat flow data of Parsons & Sclater (1977) as 3×10^8 km^3, about 6% of the present continental volume. 2900 Ma ago, when freeboard was 'established', terrestrial heat production and heat flow were at least twice their present values. On this basis, and in view of the geological evidence for stable Proterozoic continental masses, ocean ridge heat flow (either through more ridges or faster spreading, both giving more low density lithosphere) may have been at least four times greater: the 'excess' volume of oceanic lithosphere must surely have exceeded 20% of the present continental volume.

Rather than supporting a no-growth model, the evidence of a near-constant continental freeboard during the last 2900 Ma is interpreted in terms of at least 15% growth in the continental volume during that time, a similar conclusion to those of Dewey & Windley (1981) and Dickinson (1981). Moreover, there is evidence from detailed palaeogeographic and sedimentary facies map analysis (Hallam 1977) that progressive marine regression from the continents has occurred during the last 500 Ma, equivalent to a 1–2% increase in the continental volume *or* the equivalent increase in the volume of the ocean basins. Whilst these results alone are ambiguous in terms of continental growth models, they do show that the freeboard level has fluctuated and that relative changes in the rates of continental growth and ocean basin deepening have occurred during the last 500 Ma. Since it is the latter which is controlled more directly by the Earth's smoothly-declining heat production, one might speculate that continental growth was lagging behind ocean basin deepening at the end of Proterozoic times but has since 'caught up'.

Experimental data on arc magmatism

To complete this discussion of continental growth and recycling, a few brief comments on the sources of arc magmas are necessary. Relevant phase equilibria have been reviewed extensively in the literature (e.g. Wyllie 1977; Green 1982; Mysen 1982) from which the most plausible primary calc-alkaline melt sources are by partial melting of subducted, hydrated ocean crust or of volatile-fluxed mantle peridotite overlying the downgoing slab. Figure 1(a) shows that the contemporary sub-continental P-T conditions (geotherm 4)

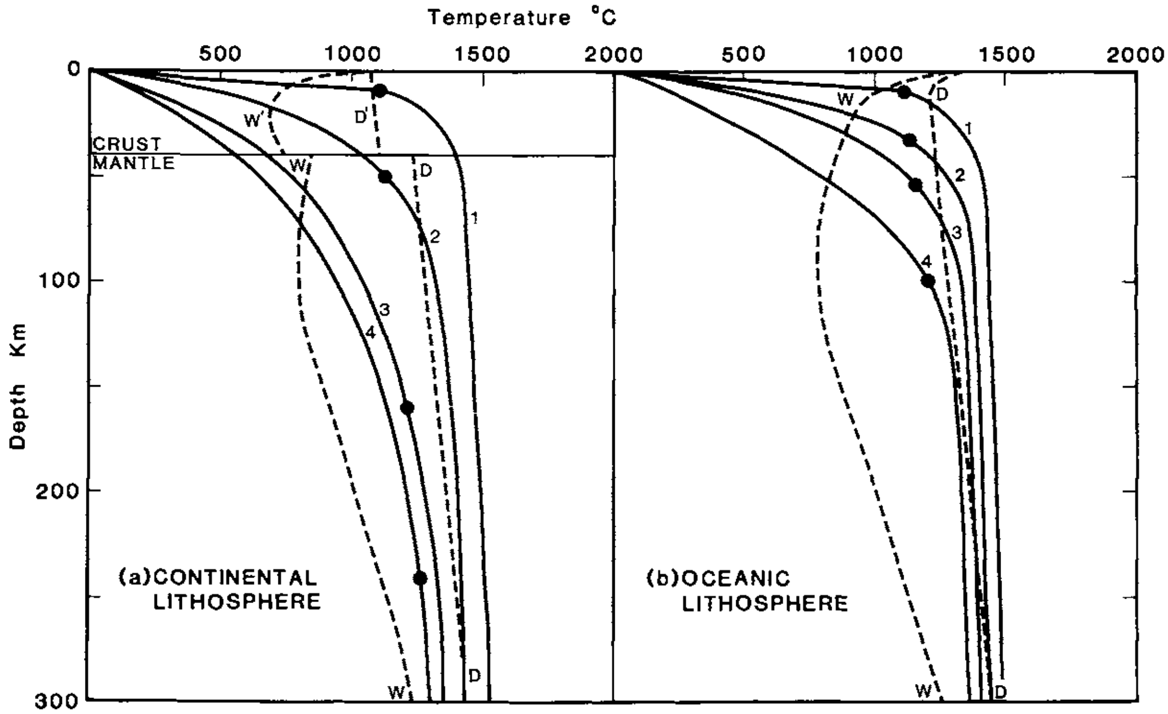

FIG. 1. Schematic illustration of possible crust and upper mantle geotherms consistent with the thermal and geological constraints discussed in the text for (a) continental and (b) oceanic lithosphere. Curves are labelled as follows: 1. Pre-Archaean, 2. Archaean, 3. Early Proterozoic, 4. Late Proterozoic and Phanerozoic. Also shown are the wet and dry solidus curves for mantle peridotite (W and D, from data in Green 1982 and Mysen 1982) and, in Fig. 1(a), for average continental crust (W' and D' from data in Brown 1981). Points indicated by solid circles on each geotherm indicate the approximate temperature and depth of the lithosphere — asthenosphere boundary (see text for discussion).

are, indeed, adequate for hydrous partial melting of the mantle to occur at depths of c.50–300 km (between curves W and D). Although, for many arcs, there may be a subducted sediment contribution (5–10% according to Hawkesworth 1982 — see also earlier discussion), predominantly, these sources are, respectively, of recent mantle extraction (ocean crust) or within the mantle itself. Subsequently, the rising melt may undergo extensive fractionation and/or be modified by the addition of further partial melt at crustal levels to yield the diverse acid-basic volcanic and intrusive suites observed in modern arcs (cf. Brown 1981). Effectively, this added contribution represents internally-recycled continental crust and it requires first that crustal melting conditions are encountered (i.e. between curves W' and D' in Fig. 1(a)) and, secondly, that the latent heat of fusion is provided either by fractional crystallization of a high-temperature residue or by cooling of super-heated rising magma. Only under these conditions will the modern continental geotherm (curve 4, Fig. 1(a)) be displaced into the field of partial melting: consequently, the contribution from crustal sources to magmatic processes in most arcs is thought to be minor, reaching significant proportions only in the more mature, thickened continental arcs and zones of crustal convergence.

For the most part, therefore, arc magmas represent additions to the crust, the accretion component of the equation for continental growth: growth = magmatically accreted volume (0.4–1.0 km^3a^{-1}) — volume recycled into the mantle (thought to be negligible, see earlier discussion). Earlier in the history of continental growth, not only was the accretion rate higher, but the crustal contribution to magmatism was almost certainly more important. Postulated continental lithosphere geotherms (curves 1–3 in Fig. 1(a) represent the change from pre-Archaean to early Proterozoic times) intersect the field of partial melting in crust of normal thickness which, according to the geological evidence, had developed by mid to late Archaean times. Crust–mantle interaction and internal re-cycling of the continental crust were both significantly more important processes than they are today.

Geothermal constraints on lithosphere thickness and dynamics

The inverse relationship between the Earth's heat production, thence surface heat flow, and the average thickness of conductive lithosphere was introduced earlier in this paper. Uncertainties in the present-day relative abundances of elements with long-lived radioisotopes having different half-lives, particularly in the K/U ratio of the Earth, limit the success with which estimates of past heat production can be made. A useful analysis of this problem by O'Nions *et al.* (1978) suggests an optimum K/U ratio of 10^4 which leads to a combined half-life of 2100 Ma for long-lived radioisotopes, and a pre-Archaean heat production c. four times the present value (this corresponds to the lower end of the range given by McKenzie & Weiss (1975) and is similar to Lambert's (1980) estimate). If 300 mWm^{-2} were lost uniformly over the Earth's surface, as required by models of uniform, thin globe-encircling pre-Archaean–early Archaean lithosphere, then with normal thermal conductivities of *c*.3 $Wm^{-1}K^{-1}$, the depth to the top of the asthenosphere (*c*.1000–1200°C) would be 10–12 km. This corresponds to the situation shown in Fig. 1 for pre-Archaean lithosphere (curves labelled 1) where no distinctions are drawn between geotherms and lithosphere thicknesses for oceans and continents. It is assumed that conduction of heat occurs uniformly across rigid lithosphere, that the top of the asthenosphere occurs at a temperature some 100–200°C below the mantle solidus (taken as the dry solidus in Fig. 1) where plastic deformation commences (cf. Pollack & Chapman 1977; Brown & Mussett 1981, Chapter 8) and that the geotherm then follows an adiabatic path in the convecting layer.

A variety of considerations, such as the probability of frequent surface impacts and the instability of thin lithosphere overlying vigorously active asthenosphere make it unlikely that the real Earth was ever covered by a widespread thin sheet of ocean-like lithosphere. Nevertheless, the early surface of the Earth probably comprised a mosaic of rapidly thickening and foundering 'platelets' with an *average* thickness close to 10–12 km.

Geological evidence for the production of permanent thick continental crust during the Archaean requires also the development of stable continental lithosphere: curve 2 represents a likely transitional geotherm where the lateral and vertical dimensions of Archaean continental lithosphere had reached *c*.50 km. However, the progressive development of the refractory mantle component of continental lithosphere (Jordan 1978) must have accentuated ocean-continent differences. A mid-Archaean heat flow of ≥ three times the present value requires an average lithosphere thickness of *c*.20 km so, at this stage, heat loss was already being focused into oceanic areas (compare curve 2, Fig. 1(b)). Since the oceanic accretion rate varies as the square of heat flux it has been argued (cf. Bickle 1978; Davies 1979) that either the rates of oceanic lithosphere generation and motion were an order of magnitude greater than today and/or that there were, similarly, many smaller oceanic plates with a much greater ridge length. Mantle convection models based on a slightly higher Rayleigh Number for the Archaean mantle are equivocal on this point since faster convection or a greater number of cells are equally viable solutions.

It is proposed that the rapid growth of continental crust in late Archaean times was coupled with an accentuated ocean-continent contrast leading to lithospheres approaching 200 km and 50 km thickness, respectively, by 2600 Ma ago. Rapid thickening of the continental lithosphere was consequent on (1) the development refractory root zones, as continental fragments coalesced and the sub-continental mantle was depleted of volatiles and lithophile elements (including long-lived radio-isotopes) and (2) the intersection of a cooler upper mantle geotherm with granulite facies stability conditions in the upper 200 km (Tarling 1980). A rapid increase in continental lithosphere thickness from *c*.50–200 km during the late Archaean peak of activity is consistent with the geological evidence of a rapid change in continental character at this time. Nevertheless, parts of the Proterozoic upper mantle were still sufficiently ductile to allow continental rifting and the development of inter-cratonic mobile belts; the more anhydrous refractory and rigid upper mantle zones probably lay beneath Archaean cratonic nuclei.

In early Proterozoic times heat flow would have decreased to an average of about 120 to 150 mW m^{-2}, corresponding to an average lithosphere thickness approaching 25 km. With a continental lithosphere of 200 km thickness, clearly, most of the Earth's heat loss had become concentrated into the oceanic areas (cf. contrast between curves labelled 3 in Fig. 1(a) and 1(b)). This situation must be qualified in two ways: (1) by this time, some 20–30% of the Earth's heat-producing radioisotopes are likely to have reached the continental crust (Pollock & Chapman 1977; Lambert 1980) such that, over the last 2600 Ma, continental heat flow has been at least twice the 20–25 mW m^{-2} which would arise from conduction alone through a 200 km thick lithosphere, (2) over the same period (as today, Sclater & Crowe 1979) ocean ridge heat flow has greatly exceeded the average for conduction through oceanic lithosphere. For these reasons, the thickness of typical early Proterozoic oceanic lithosphere is shown as *c*.50 km in Fig. 1(b). This may be a maximum estimate since other thermal models based on shorter heat production half-lives than 2100 Ma (e.g. McKenzie & Weiss 1975; Tarling 1980) would require an ever greater proportion of the larger Precambrian heat flux to be concentrated through thinner ocean lithosphere.

Similar qualifications apply to the last 2600 Ma which have been characterized by a progressive decrease in heat production coupled with further thickening of both oceanic and continental lithospheres due to small (<100°C) decreases in upper mantle temperatures (compare curves 3 and 4 in Fig. 1). Away from zones of active magmatism, the rheological properties and thickness of continental lithosphere may, during the last 1000 Ma, have reached a near-equilibrium state whereas old oceanic lithosphere has become more rigid. The possibility of decreasing Phanerozoic contrasts between oceanic and continental lithosphere rigidity led Tarling (1980) to suggest that continental splitting may have become more important since fractured continental lithosphere would be as weak as old oceanic lithosphere. This deduction is particularly sensitive to the choice of a thermal model since, arguing backwards from present day observations of 100 and 200–300 km thick oceanic and continental lithospheres, the more rapidly heat production increases, the greater the implied lithospheric contrasts during Proterozoic times. Certainly the geological and palaeomagnetic evidence (Piper 1976; Windley 1977, 1981) indicate that fragmentation and independent movement of continental masses have increased during the Phanerozoic; it is also suggested that Phanerozoic decreases in the rate of mantle convection and the possibility that continental plates may remain stationary for significant periods over zones of upwelling mantle convection should be regarded as important factors in determining recent crustal evolution.

Conclusions

Key geological features of the continental crust and their qualitative interpretation, as discussed in this paper, appear in Table 1 whilst semi-quantitative aspects are summarized in Fig. 2. Throughout the Earth's history, internal heat production has provided the dominant constraint on lithosphere evolution. During the first 800 Ma, internal activity was too vigorous for permanent continental crust to form and adiabatic asthenospheric mantle reached to within 10 km of the surface (on average). Once the lithosphere started to thicken

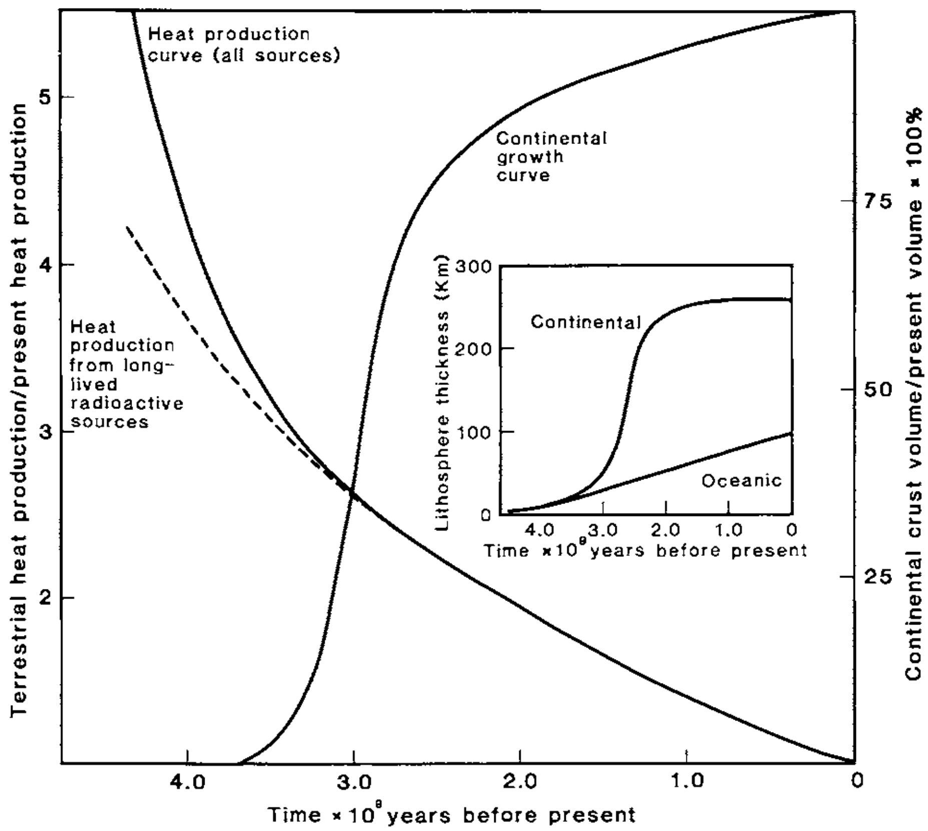

FIG. 2. Heat production (left hand axis) and continental growth (right hand axis) curves proposed in this paper: note that the latter need not be smooth but, in reality, is likely to fluctuate according to the disposition of continental masses, the length of magmatic arc systems and other variables affecting accretion rates. Inset shows changes in lithospheric thicknesses consistent with thermal and geological data discussed in the text.

TABLE 1 Summary of crustal characteristics and their interpretation in terms of lithosphere evolution

Period	*Geological features*	*Proposed interpretation*
Pre-Archaean	No crust > 3800 Ma preserved, period of of severe bombardment by planetisimals inferred from lunar and other studies	Thin, widespread ocean-like lithosphere in the form of rapidly-produced and foundering platelets. Frequent lithosphere disruption, little or no scope for salic magma production.
Archaean	Small stable crustal blocks with gneiss-greenstone characteristics. Local high pressure metamorphism, abundant tonalite-trondhjemite calc-alkaline series, with extensive intra-crustal reworking. Coarse clastic supracrustal sediments.	Dehydration and melting depths increasing, small stable continental rafts (< 50 km lithospheres) rapidly produced, achieving modern crustal thickness. Heat production and heat flow c.3x present value requiring many thin, hot and/or rapidly moving oceanic plates, possible back-arc openings and closings.
Proterozoic	Large stable continental platforms developed, crossed by linear dyke swarms, mobile belts with calc-alkaline and alkalic granite-anorthosite suites. Extensive thick sedimentary cover.	Rapid early thickening of continental lithosphere to c.200 km as Archaean fragments coalesced and formed rigid anhydrous, refractory roots. Ductility preserved between cratons allowing mobile belt development. Heat loss concentrated in oceans with lithosphere 50 km thick or less.
Phanerozoic	Major linear marginal and collisional orogenic belts with calc-alkaline batholiths, ophiolites etc. Increased continental fragmentation and independent movement; otherwise stable platforms.	Possibility that ocean/continent lithosphere rigidity and thickness contrasts have decreased. Sub-continental lithosphere less ductile but fragments over focused mantle heat sources.

as heat production decreased, melting depths became adequate to produce salic magmas and upper mantle rigidity became adequate to sustain coherent continental lithosphere, given that major heat loss was focused elsewhere. The bulk of the continental crust (c.75 – 80% of the present volume — see Fig. 2) formed in about 500 Ma during the late Archaean; subsequently, continental and oceanic lithospheres were destined to follow separate thermal evolutionary paths (cf. Fig. 1).

Continental lithosphere has been dominantly refractory, rigid and stable for c.2600 Ma, though a trend from large coherent masses showing internal ductility to more recent fragmentation may reflect continued cooling and greater sub-crustal uniformity. Interpretive trends in lithospheric thickness (see Fig. 2 inset) show a peak in ocean-continent contrast during early Proterozoic times, decreasing towards the present day. Progressive increases in oceanic lithosphere thickness, coupled with declining heat production, probable slower oceanic plate motions and greater average age, are taken to imply less thermally-expanded lithosphere and an increasing ocean volume over the last 2600 Ma. Thus, continental freeboard data, together with isotope, geochemical and experimental data on magma sources, indicate continued, though exponentially decreasing rates of continental growth during this time (also illustrated in Fig. 2).

As a postscript, it should be noted that this model for continental evolution does not argue strictly that modern plate tectonics has characterized the whole of the Earth's history. As Hargraves (1981) has indicated, there has been a long-term change from dominantly viscous drag at the base of the lithosphere to buoyancy-powered subduction as the vigour of mantle convection has declined. In the former case, the lithosphere is largely decoupled from convecting mantle beneath, though strong vertical drag forces operate over down-going convection limbs, whereas buoyancy-powered subduction necessitates greater coupling. Although Hargraves and others would limit subduction to late Proterozoic and Phanerozoic times, if important oceanic and continental lithospheric differences were established during the Archaean, as argued in this paper, an earlier shift in the balance from viscous drag to buoyancy forces might be appropriate. As with all processes and problems concerning lithospheric evolution, consensus will result only from a continued debate in which interpretive models are tested and re-evaluated.

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