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TOWARD A PHYSICAL TIME SCALE FOR THE NEOGENE - DATA FROM THE AUSTRALASIAN REGION

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ABSTRACT

Isotopic dating on rocks in close stratigraphic relationship with paleontologically well-dated sedimentary rocks provides the basis for constructing a physical time scale. For the Neogene the K-Ar isotopic dating method is of paramount importance; because of this an outline of some features of the method and its application is given. A compilation of physical ages from marine sequences shows that the number of reliable results for the Neogene is small, highlighting the need for much additional data.

Potassium-argon ages on rocks from the Australasian region provide useful data for the Neogene time scale. The boundary between the Janjukian and Longfordian stages in Victoria is slightly younger than the Oligocene-Miocene boundary and has been dated at about 21.4 m.y. The East Indies lower Tf (Tf_{1-2}) stage extends from about 12.5 to 15 m.y. ago from dating of volcanic rocks associated with sediments assigned to this stage in New Guinea. Two new K-Ar ages of 15.5 m.y. are reported for rocks from New Guinea that intrude upper Te to lower Tf sediments. Data from volcanic rocks associated with marine sequences containing definitive planktonic foraminifera in Fiji indicate that the age of the Miocene-Pliocene boundary is equal to or younger than 5.3 m.y., and an estimate of 4.9 ± 0.4 m.y. is given for this boundary. A combination of paleontological correlation, magnetic stratigraphy and K-Ar dating indicate that an age of about 1.7 to 1.8 m.y. is the best estimate for the age of the base of the Calabrian stage defined as basal Pleistocene. This level has been recognized in New Zealand but it is not known in Australia with any precision as yet.

INTRODUCTION

The development of an accurate physical or radiometric time scale for the Cenozoic has become of increasing importance in relation to many geological and geophysical studies. Time scales such as those of Holmes (1959), Kulp (1961), Funnell (1964) and Berggren (1969a,b; 1971; 1972) form an evolutionary sequence of progressively better correlations and age assignments as the amount of biostratigraphic and physical age data increase. The aim of the present paper is to discuss and briefly summarise the available data for the Neogene with more detailed discussion of results obtained from the Australasian region.

GEOCHRONOLOGY AND PHYSICAL TIME SCALES

The ideal situation is to have many reliable and internally consistent isotopic age determinations on suitable rocks, generally volcanic, whose positions within one or more of the stratigraphic or zonal schemes are determined accurately by the presence of diagnostic fossil assemblages. Only in the case of the North American Land Mammal Ages are adequate data available to construct an independent physical time scale (Evernden et al., 1964). However, results are becoming available on rocks from marine sequences which generally are more readily correlateable than the terrestrial successions. Nevertheless the aim should be to obtain sufficient physical age data for each zonal or stage scheme so that the correlations previously made can be checked independently. The importance of this approach already has been demonstrated - it seems firmly established from isotopic age data that earlier correlations of vertebrate bearing continental successions with the marine sequences were considerably in error (Berggren, 1969; Page and McDougall, 1970).

Most of the physical age data for the Cenozoic, and the Neogene in particular, have been obtained using the K-Ar isotopic dating method. When applied carefully to suitable rocks this method can yield reliable results, but when used on unsuitable materials conflicting results of little value will be obtained. In the context of the present topic it is appropriate to summarize briefly some of the important factors and constraints in applying the K-Ar method. Fuller discussions of these questions have been given by Evernden et al. (1964), McDougall (1966) and Dalrymple and Lanphere (1969).

As with all dating methods there are several assumptions that have to be met if a K-Ar measurement is to give a valid estimate of age. One of the most important of these assumptions is that at the time of formation of the rock or mineral chosen for dating, all pre-existing radiogenic argon was lost. A second very important assumption is that no loss or gain of potassium or radiogenic argon has occurred since the time of formation of the rock, apart from the changes produced by decay of K^{40} to Ar^{40}. Should the first assumption fail then the calculated K-Ar age will be too old, whereas failure of the second assumption generally leads to a calculated age that is too young because of loss of radiogenic argon by diffusion.

Care in the choice of samples for dating minimizes these problems; nevertheless an evaluation as to whether the assumptions are met must be made for each study undertaken. Such evidence can be obtained from the internal consistency or otherwise of results on stratigraphically closely related but different rocks or minerals. Generally the meaning or validity of a single, isolated K-Ar age is difficult to establish, and therefore should not be given the same weight as results from studies where internal checks on consistency indicate that the basic assumptions are met.

For time scale purposes volcanic rocks, particularly lavas but also tuffs, are the most useful materials for physical dating especially when they occur interbedded with sediments on which there is biostratigraphic control. Lavas often can be dated satisfactorily by the K-Ar method using whole rock samples provided that they are unaltered and holocrystalline. Glass or devitrified glass is prone to leak radiogenic argon even at ambient temperature so that rocks containing such material generally should not be used for dating. In the ideal case a series of lavas would be dated as whole rocks together with one or more minerals. Useful minerals for K-Ar dating include high temperature alkali feldspar, plagioclase, hornblende and biotite. Caution must be exercised when dating phenocrystic phases because they may have crystallized prior to eruption of the magma in an environment in which there was a high partial pressure of argon. In such cases the crystallizing phenocrysts may include some of this argon and result in a measured age that is greater than the true age of crystallization. Such an effect is most noticeable in minerals such as plagioclase and hornblende which have relatively low potassium contents and which therefore generate a relatively small amount of radiogenic argon per unit time.

Tuffs are more difficult to date than lavas because of the increased possibility of contamination by much older material (cf. Evernden and James, 1964; Evernden and Curtis, 1965). For tuffs one or more of the minerals mentioned above generally are separated for K-Ar dating. Volcanic glass has been used quite commonly for K-Ar age determination, but data on such material have to be treated with caution. Even when volcanic glass appears to be undevitrified it may leak radiogenic argon, causing measured ages to be too young. Devitrified, hydrated or otherwise altered glass is likely to yield ages that are much too young. On the other hand the rapid chilling of magma to glass may prevent the complete loss of radiogenic argon dissolved in the magma, and extraneous argon of this kind also may be carried in bubbles which are observed not uncommonly in volcanic glass.

Glauconite, a potassium-rich mineral that forms authigenically in the sedimentary environment, is about the only mineral that can be used for d ting sediments directly by the K-Ar method. Because of its composition, structure and small crystal size, glauconite is prone to leak radiogenic argon at temperatures perhaps as low as 150°C. Thus most K-Ar ages on glauconite must be regarded as minimum ages, although measured ages that are too old also are possible if detrital material is incorporated in the mineral as it forms.

Thus an isolated K-Ar age determination is of relatively little value and indeed could be very misleading. On the other hand concordant results on a number of related rocks and minerals provide confidence that the age is geologically meaningful.

During the last few years there has been a dramatic increase in precision of measurement of isotopic ratios, such that it is likely that in the future the Rb-Sr dating method also will be applied more to rocks of the age range being considered here.

DATA FOR THE NEOGENE TIME SCALE

In Table 1 are listed those age results that we consider to be useful and reliable for the construction of the physical time scale for the Neogene. For the present study we have chosen to include only those data related to marine sequences. Choice of data to be included in Table 1 was primarily on the basis that sufficient information was available to indicate that the physical ages are likely to be reliable. In most cases two or more closely related samples have been dated and the measured ages are internally consistent. Where inconsistencies of more than 10 percent of the age exist between closely related samples the results were excluded.

The few results on Neogene glauconites are omitted because of the lack of supporting results that demonstrate reliability. Similarly most results on volcanic glasses are excluded, with the notable exception of some data from the Experimental Mohole, Caudelupe Site, reported by Dymond (1966), who demonstrated that the results are internally consistent.

The magnetic stratigraphy of deep sea cores commonly allows correlation with the standard polarity time scale for the last 3 or 4 m.y. (cf. Opdyke, 1972). We have accepted a number of determinations of this kind for Table 1.

In the preceding discussion emphasis has been placed on reliability of the physical ages, but it is realized that there are uncertainties also in the stratigraphic assignments. Commonly it is difficult to evaluate independently the reliability of the stratigraphic and biostratigraphic control; in most cases we have accepted the assignment given in the papers referenced in Table 1.

The limited number of results in Table 1 emphasizes that we are far from being able to set up independent physical time scales for the various zonal or stage schemes. Clearly a much greater number of reliable results is required before this can be done. In the meantime reliance on paleontological correlation is necessary to construct an overall time scale, such as is given in Fig. 1. This time scale is only an approximation and undoubtedly will undergo considerable changes as more reliable physical ages become available and as the paleontological correlations become more precise.

NEOGENE DATA FROM THE AUSTRALASIAN REGION

Janjukian-Longfordian Boundary, Victoria

In southern Victoria near Geelong, the Maude Basalt lies stratigraphically between two limestone units which are dated as late Janjukian and early to mid Longfordian on the basis of foraminiferal faunas (Abele and Page, 1972). The field relations indicate that the Maude Basalt probably is very early Longfordian in age. The Janjukian-Longfordian stage boundary is regarded as virtually equivalent to the Oligocene-Miocene boundary from studies of planktonic foraminifera (Carter, 1964; Ludbrook and Lindsay, 1969; McGowran et al., 1971). Globoquadrina dehiscens first appears at the base of the

Longfordian and together with other indices shows that the Janjukian-Longfordian boundary of southern Australia occurs in zone N.4 of Blow (1969). The detailed stratigraphy, paleontology and geochronology are discussed by Abele and Page (1972) who report concordant ages for four separate whole rock samples of the Maude Basalt, with a mean of 21.4 $\pm$ 0.2 m.y. (This age is slightly greater than that referred to in Page and McDougall (1970), owing to a calibration adjustment). Thus the boundary between zones N.3 and N.4, generally accepted as the position of the Oligocene-Miocene boundary, is at or slightly older than 21.4 m.y. Similarly the boundary between the Zemorrian and Saucesian stages in California is correlated with the boundary between zones N.3 and N.4. Turner (1970) reports an age of 22.5 m.y. for the Zemorrian-Saucesian boundary. Thus the data from Victoria and California are mutually consistent and together indicate that the Oligocene-Miocene boundary has an age of about 22.5 m.y.

East Indies Letter Stages

The East Indies letter classification or stage scheme is based upon assemblages of larger foraminifera and their evolutionary trends; no stratotypes have been defined. This scheme was introduced by Van der Vlerk and Umbgrove (1927) because of the difficulties encountered in correlating the rocks in the East Indies with the European stages. The East Indies latter stages are designated Ta (oldest) through to Th (youngest), covering virtually the whole of the Tertiary; recent summaries of the scheme include those by Clarke and Blow (1969) and Adams (1970).

There is relatively little physical age data available on rocks associated with the East Indies letter stages. Ladd et al. (1967) recorded ages of 16 to 17 m.y. on basalts from holes drilled at Midway atoll, but regarded the ages as too young. McDougall (1971) preferred to accept the ages at face value. The basalts at Midway are overlain by a Te, probably upper Te, assemblage of large foraminifera. Page and McDougall (1970) provided age data on lavas interbedded with sediments assigned to the lower Tf (Tf_{1-2}) stage in Papua New Guinea, and suggested that the lower Tf extended from about 15 to 12.5 m.y. ago. Hornblende separates from two porphyritic bodies intrusive into sediments (Movi Beds) immediately below the Daulo Volcanics yield K-Ar ages of 15.5 m.y. (Table 2). The Movi Beds are a newly defined formation which contains an upper Te to lower Tf fauna (Bain et al., in prep.; Binnekamp and Belford, 1970); the new results reported in Table 2 are consistent with the earlier data as they show that the Te stage is 15.5 m.y. or older.

Clarke and Blow (1969) correlated the lower Tf stage with the middle Miocene planktonic foraminiferal zones N.9 (part) to the N.12-N.13 boundary (Blow, 1969). Page and McDougall (1970) reported that the dated Daulo Volcanics in Papua New Guinea had a planktonic foraminiferal assemblage in sediments below, consistant with zones N.11 and N.12. Berggren (1972) in attempting to reconcile all the physical age data and the paleontological correlations, placed considerable emphasis on our reference to zones N.11 and N.12 below the dated lavas. He drew attention to the possibility that the base of the Tf stage could be somewhat older than the 15 m.y. age given by Page and McDougall (1970) based upon the oldest measured age from the Daulo Volcanics. Re-examination of the fossil assemblage by Belford (pers. comm.) confirms the N.11-N.12 assignment, but subsequent detailed mapping of the sediments (Bain et al., in prep.) shows that they occur in a partly fault-bounded block. This means that the exact relationship of the N.11-N.12 sediments to the Daulo Volcanics in fact cannot be determined. Thus, it is no longer possible to argue for an older age for the lower Tf using these data. Indeed, as mentioned earlier, the Movi Beds underlying the Daulo Volcanics, contain an upper Te to lower Tf larger foraminiferal assemblage, which together with the age data makes it unlikely that the base of the lower Tf is very much older than 15 to 15.5 m.y.

Page and McDougall (1970) suggested an age of about 12.5 m.y. for the lower Tf (Tf_{1-2}) - upper Tf (Tf_3) boundary. This was based largely on the report by Rodda et al. (1967) that upper Tf to early Tg sediments contain boulders from granites which have been dated at 9 to 11 m.y. Recently Gill and McDougall (1972) re-emphasized that similar larger foraminiferal faunas also occur in close association with planktonic foraminiferal zones N.17 to N.19 in Fiji, posing serious problems. Thus the arguments used by Page and McDougall (1970) and further elaborated on by Berggren (1972) are not as definitive as previously thought. In fact we have no clear physical age control on the younger limit of the lower Tf stage. Berggren (1972) suggested several alternative correlations of the East Indies letter stages with the planktonic foraminiferal zones and the classic European marine stages. At present biostratigraphic arguments seem unable to distinguish between the possibilities and little physical age data are available. In this context it should be noted that Van Couvering and Miller (1971) gave an estimate of 13.5 m.y. for the age of the N.13-N.14 boundary. This estimate is based in part upon some K-Ar ages of Konecky et al. (1969) on whole rock samples that overlie Badenian sediments in Czechoslavakia. These ages were calculated using the K^{40} decay constants employed in east European countries; an age of 13.5 m.y. reduces to 12.9 m.y. if calculated with the constants normally used in western countries (cf. Table 2).

It is not our intention here to become involved in more detailed discussions of correlations, but from the examples given the need for additional physical age det rminations to help resolve the uncertainties should be obvious. At present we only have data from Midway and New Guinea which are directly related to the East Indies letter classification.

Miocene-Pliocene Boundary - Data from Fiji

Any discussion of the Miocene-Pliocene boundary must commence by reference to the classic Neogene stratigraphy of Italy. Selli (1964) defined in Sicily a neostratotype for the upper Miocene Messinian stage which is overlain by the 'trubi' marl of the Zanclian stage, regarded as lower Pliocene. This definition for the Miocene-Pliocene boundary has been widely accepted. Studies by Blow (1969) indicate that this boundary lies within planktonic foraminiferal zone N.18, just below the base of zone N.19, which is defined by the first evolutionary appearance of Sphaeroidinella dehiscens. This species is first observed just above the base of the "trubi" marl sequence, overlying the Messinian strata.

Potassium-argon dating of volcanic rocks in Fiji provides some control on the age of the Miocene-Pliocene boundary as planktonic foraminiferal zones bracketting this boundary are recognized in associated sediments. Gill and McDougall (1972) reported concordant hornblende and whole-rock ages of 5.7 ± 0.1 m.y., for the Namosi Andesite in southeast Viti Levu, Fiji. The Namosi Andesite is a formation laterally equivalent to the Suva Marl, samples from which have yielded faunas indicative of zones N.17, N.18 and N.19 (Blow, 1969; Parker, 1967). Correlation of the Namosi Andesite relative to the sediments containing the definitive faunas is not known in detail, but at one locality sediments containing microfossils indicating zone N.18 overlie Namosi Andesite. Thus the measured age of 5.7 m.y., is regarded as a reliable maximum age for the Miocene-Pliocene boundary.

In western Viti Levu, McDougall (1963) recorded concordant ages of 5.35 ± 0.1 m.y., for two samples of biotite separated from lavas of the Koroimavua Volcanic Group in the Sambeto Range. A tuff sample (PEK/F4) from beds conformably overlying these volcanics yielded a fauna indicative of zone N.18 (Blow, 1969; Bartholomew, 1959), and sediments below the lavas are considered to be no older than upper Miocene (Bartholomew, 1959; 1960). As the Miocene-Pliocene boundary lies near the top of zone N.18 we conclude that the 5.35 m.y. age of the Koroimavua lavas is equivalent to or slightly older than that of the boundary.

A younger limit for the age of the Miocene-Pliocene boundary was provided by Hays et al. (1969) from their study of deep-sea cores of the equatorial Pacific. They showed that Sphaeroidinella dehiscens is present at the bottom of core V24-59 which level has an age of about 4.5 m.y., by magnetic stratigraphy. Because the first appearance of S. dehiscens is at the base of zone N.19 and because the Miocene-Pliocene boundary is in the upper part of zone N.18, then this boundary must be somewhat older than 4.5 m.y. Combining this information with the Fijian data the age of the Miocene-Pliocene boundary may be estimated to be 4.9 ± 0.4 m.y. (Gill and McDougall, 1972). This estimate is consistent with the rather limited amount of reliable physical age data available from elsewhere (Table 1).

The Fijian study provides an object lesson in correlation. When the ages of 5.35 m.y. for the Koroimavua lavas were reported by McDougall (1963), the assignment of the associated sediments to the upper Miocene to lower Pliocene was questioned, because it was believed at that time that the age of the Miocene-Pliocene boundary was firmly established at 12 to 13 m.y. This estimate for the age of the boundary was based upon correlations from the well-dated (K-Ar) North American mammalian faunas to the European stages, as there were few or no physical age data for this part of the time scale on samples from marine sequences. As geochronological data became available from marine sequences it was increasingly obvious that the earlier correlations were in error (Berggren, 1969; Page and McDougall, 1970). Indeed if McDougall (1963) had accepted the paleontological assessment available at that time for the Fijian sequences, the young age for the Miocene-Pliocene boundary may have been recognized much earlier than was the case.

Another point raised by the Fijian data relates to the use of large foraminifera as indices of age, and the East Indies letter classification in general. As noted above the Suva Marl contains planktonic foraminifera indicative of zones N.17, N.18 and N.19. However, in sediments interbedded with those yielding the planktonics are larger foraminifera regarded as of Tf (middle Miocene) age (Ladd, 1934). There seems to be little doubt that the larger foraminiferal fauna in these sediments is characteristic of the Tf stage.

To explain this anomaly, which has been recognized previously by Ibbotson (1960), there either must have been reworking or the fauna lived much longer in Fiji than elsewhere. Gill and McDougall (1972) have shown that there is insufficient information to be able to distinguish between these two possibilities.

A final more general point is that in discussion of the Miocene-Pliocene boundary reference is commonly made to the Sphaeroidinella dehiscens datum plane. Some authors define this datum plane as being at the level of the first appearance of the taxon, while others use the term to mean the level at which the taxon greatly increases in abundance. Certainly to the non-paleontologist this can be very confusing, and we make a plea to authors to state unequivocally the sense in which a particular term is being applied.

Pliocene - Pleistocene Boundary

A great deal has been written on the Pliocene-Pleistocene boundary in recent years and an excellent summary of this topic is given by Hays and Berggren (1971). As we have little new information only a brief discussion will be given here.

Definition of the base of the Pleistocene in terms of climatic cooling has been abandoned by most workers, at least partly because it has become increasingly evident that this approach is quite unsatisfactory (cf. McDougall and Stipp, 1964). There now seems to be nearly universal acceptance of the stratigraphic definition that the base of the marine Calabrian stage in Italy is basal Pleistocene. With advances in the study of microfossils, correlation of strata from elsewhere with the Italian stratotype has become practicable.

Blow (1969) defined Neogene planktonic foraminiferal zone N.22 as the level of the initial evolutionary appearance of Globorotalia truncatulinoides from its immediate ancestor G. tosaensis. Bayliss (1969) showed that G. truncatulinoides first occurs in the Calabrian type section about 30 m above its base, and noted that G. tosaensis is extremely rare in the underlying sediments. Jenkins (1971) therefore questioned whether the evolutionary transition from G. tosaensis to G. truncatulinoides in fact occurs in the type Calabrian as has been accepted by Blow (1969) and most other workers. This same evolutionary sequence has been recognized in many deep-sea cores and is used as the primary means of correlation to the type Calabrian.

Magnetic stratigraphy tells us that the evolutionary transition at the base of zone N.22 occurs within the Olduvai (or possibly the Gilsa) normal polarity event, corresponding to an age of about 1.7 to 1.8 m.y. Undoubtedly the use of other paleontological indices and magnetic stratigraphy (cf. Nakagawa et al., 1971) in the type section of the Calabrian will result in firmer and more precise correlations becoming possible in the future.

In sequences in Australia and New Zealand the boundary between the Pliocene and the Pleistocene traditionally has been drawn at the level of the first indication in the late Tertiary of marked climatic cooling. Using this criterion Gill (1957; 1961) suggested that for Australia the Pliocene - Pleistocene boundary should be placed at the base of the marine Werrikooian stage in Victoria. Biostratigraphic and other studies presently in progress should enable this suggestion to be evaluated in terms of the stratigraphic definition of Pleistocene.

In New Zealand the onset of cooling is noted at about the boundary between the Waitotaran and Hautawan stages. Potassium-argon ages on volcanic rocks erupted at about this time indicate an age of around 2.5 m.y., for this cooling. Oxygen isotope measurements and magnetic stratigraphy (Devereaux et al., 1970; Kennett et al., 1971) strongly support this view. Kennett et al., (1971) have shown that the first definite appearance of Globorotalia truncatulinoides is near the middle of the Hautawan stage at an age of about 1.8 m.y., from the magnetic stratigraphy correlations. Thus, the first major cooling in the late Tertiary is recognized in New Zealand sequences some 0.7 m.y., earlier than the Pliocene-Pleistocene boundary, using the presently accepted stratigraphic definitions that the Calabrian stage is basal Pleistocene.

CONCLUSIONS

Much progress has been made during the last decade in the establishment of physical time scales for the Neogene. Nevertheless the amount of reliable isotopic age data available on stratigraphically well-controlled samples remains small (Table 1). It can be seen in Fig. 1 that there are still large uncertainties. For example, there are very few physical age determinations on stratigraphically controlled rocks in the age range 8 to 13 m.y.; thus there is considerable latitude in the assignment of age to most of the schemes illustrated in Fig. 1. Adjustment of the age of zonal or stage boundaries by more than 2 m.y. in this part of the time scale seems possible (Van Couvering and Miller, 1971; Berggren, 1972), and highlights the need for a great deal more physical age data. Such results are required to provide the necessary control for the time scales and to facilitate independent evaluation of the correlations between the various schemes. It is an area in which collaboration between stratigraphers, paleontologists, paleomagneticians and isotope geochronologists can be very fruitful.

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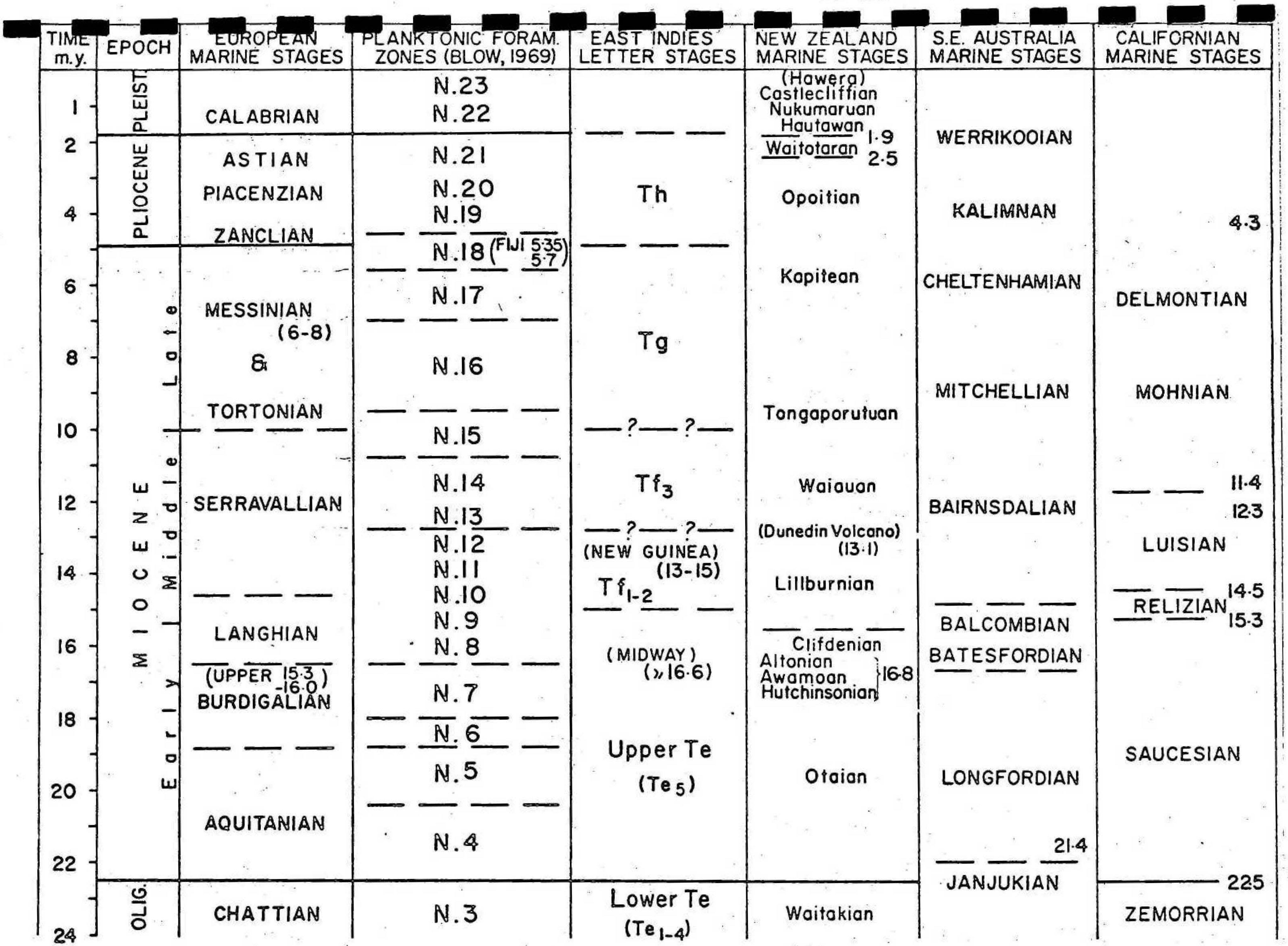
Figure 1 : A Neogene physical time scale.

Table 1 : List of acceptable physical ages on rocks from marine sequences relevant to the Neogene time scale.

Table 2 : Potassium-argon ages on rocks intrusive into sediments below the Daulo Volcanics, Eastern Highlands, New Guinea.



TIME m.y.
EPOCH
EUROPEAN MARINE STAGES
PLANKTONIC FORAM. ZONES (BLOW, 1969)
EAST INDIES LETTER STAGES
NEW ZEALAND MARINE STAGES
S.E. AUSTRALIA MARINE STAGES
CALIFORNIAN MARINE STAGES
1
2
4
6
8
10
12
14
16
18
20
22
24
PLEIST.
PLIOCENE
MIOCENE
Early
Middle
Late
OLIG.
CALABRIAN
ASTIAN
PIACENZIAN
ZANCLIAN
MESSINIAN (6-8)
&
TORTONIAN
SERRAVALLIAN
LANGHIAN
(UPPER BURDIGALIAN 15.3 -16.0)
AQUITANIAN
CHATTIAN
N.23
N.22
N.21
N.20
N.19
N.18 (FIJI 5.35 5.7)
N.17
N.16
N.15
N.14
N.13
N.12
N.11
N.10
N.9
N.8
N.7
N.6
N.5
N.4
N.3
Th
Tg
?
?
Tf3
(NEW GUINEA) (13-15)
Tf1-2
(MIDWAY) (≥16.6)
Upper Te (Te5)
Lower Te (Te1-4)
(Hawera)
Castlecliffian
Nukumaruan
Hautawan
1.9
Waitotaran
2.5
Opoitian
Kapitean
Tongaporutuan
Waiauan
(Dunedin Volcano) (13.1)
Lillburnian
Clifdenian
Altonian
Awamoan
Hutchinsonian
16.8
Otaian
Waitakian
WERRIKOOIAN
KALIMNAN
CHELTENHAMIAN
MITCHELLIAN
BAIRNSDALIAN
BALCOMBIAN
BATESFORDIAN
LONGFORDIAN
21.4
JANJUKIAN
4.3
DELMONTIAN
MOHNIAN
11.4
12.3
LUISIAN
14.5
RELIZIAN
15.3
SAUCESIAN
22.5
ZEMORRIAN

Fig 1

TABLE 1. LIST OF ACCEPTABLE PHYSICAL AGES ON ROCKS FROM MARINE SEQUENCES RELEVANT TO THE NEOGENE TIME SCALE

Age	Stratigraphic Position	Material Dated	Dating Method	Reference	Locality
1.8	Pliocene-Pleistocene boundary	Sediment cores	Magnetic Stratigraphy	Berggren *et al* (1967); Glass *et al* (1967); Phillips *et al* (1968); Hays *et al* (1969).	Ocean basins
1.9	Waitotoran-Hautawan	Sediments	Magnetic Stratigraphy	Kennett *et al* (1971)	New Zealand
2.5	Opoitian-Waitotoran boundary	Sediments	Magnetic Stratigraphy	Kennett *et al* (1971)	New Zealand
4.3 ± 0.3	Miocene-Pliocene, or Pliocene	Glass shards	K - Ar	Dymond (1966)	Experimental Mohole, Gaudelupe site.
> 4.5	Miocene-Pliocene boundary	Sediment core	Magnetic Stratigraphy	Hays *et al* (1969)	Equatorial Pacific Core V24-59
5.35 ± 0.1	Zone N.18	Biotite from lavas	K - Ar	McDougall (1963); Blow (1969); Gill and McDougall (1972)	Koroimavua Group, Fiji
5.7 ± 0.1	Zones N.17 to N.19	Whole rock lavas, hornblende	K - Ar	Gill and McDougall (1972)	Namosi Andesite, Fiji
6 - 7.5	Pre-Tabianian	Micas from granites	Rb - Sr, K - Ar	Eberhardt and Ferrara (1962) Tongiorgi and Tongiorgi (1964)	Elba, Italy
6.5 - 8.0	Messinian	Biotite from tuffs	K - Ar	Charlot *et al* (1967) Choubert *et al* (1968)	Melilla tuffs, Morocco
11.4 ± 0.6	Mohnian	Glass shards	K - Ar	Dymond (1966))	Experimental Mohole, Geudelupe sito
12.3 ± 0.4	Luisian	Glass shards	K - Ar	Dymond (1966))	
10.6 - 12.9*	Sarmatian, overlying Badenian (younger than N.13)	Whole rock lavas	K - Ar	Konecny *et al* (1969) Van Couvering and Miller (1971)	Czechoslovakia
13.7 ± 0.2	Late Lillburnian to Early Tongaporutuan	Anorthoclase, whole rock lavas	K - Ar	McDougall and Coombs (1972)	Dunedin Volcanics, New Zealand
12.5 - 15	Lower Tf (Tf_{1-2})	Whole rock lavas, minerals	K - Ar	Page and McDougall (1970)	New Guinea
13.7 - 14.5	Relizian - Luisian	Plagioclase from lavas	K - Ar	Turner (1970)	Western U.S.A.
15.3 ±	Saucesian-Relizian	Plagioclase from lavas	K - Ar	Turner (1970)	Western U.S.A.
15.3 - 16.0	Upper Burdyalian	Whole rock lavas	K - Ar	Bout, Frechin and Lippolt (1966) McDougall and Roche (unpub.)	Upper flow, Gergovie, France
16.6 ± 0.9	Upper Te	Whole rock lavas	K - Ar	Lanphere *in* Ladd *et al* (1967)	Midway, Pacific Ocean
16.8 ± 0.4	Hutchinsonian-Altonian	Whole rock lavas, plagioclase	K - Ar	Obradovich *in* Bandy *et al* (1970)	Manukau Breccia, New Zealand
21.4 ± 0.2	Janjukian - Longfordian	Whole rock lavas	K - Ar	Abele and Page (1972)	Maude, Victoria, Australia
22.5 ±	Zemorrian - Saucesian	Plagioclase from lavas	K - Ar	Turner (1970)	California

* Ages recalculated to conform with decay constants used by western workers.

Table 2: Potassium-argon ages on rocks intrusive into sediments below the Daulo Volcanics, Eastern Highlands, New Guinea.

Sample No.	Mineral	K (wt. %)	Radiogenic Ar^{40} (10^{-11} mol/g)	$\frac{100.\text{Rad. } Ar^{40}}{\text{Total } Ar^{40}}$	Age (m.y.) $\pm$ 2s.d.	Locality
69-5895	Hornblende	0.402,0.402	1.105	46.9	15.4 $\pm$ 0.5	4.8 km northeast of Chuave Patrol Post.
69-5896	Hornblende	0.437,0.444	1.222	20.4	15.5 $\pm$ 0.6	6.3 km northeast of Chuave Patrol Post.

$\lambda_\beta = 4.72 \times 10^{-10}\ yr^{-1}$;

$\lambda_e = 0.585 \times 10^{-10}\ yr^{-1}$;

$K^{40}/K = 1.19 \times 10^{-2}$ atom percent.