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Australasian Cenozoic sedimentary cycles, global sea level changes and the deep sea sedimentary record

Sedimentary cycles
Sea level
Stratigraphy
Correlation
Deep-sea hiatuses

Cycles sédimentaires
Niveau de la mer
Stratigraphie
Corrélation
Hiatus profonds

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ABSTRACT

A global sea level history determined by Vail *et al.* (1977) consists of 23 sea level cycles between the base of the Cenozoic and the Pliocene-Pleistocene boundary (65 Ma to 1.8 Ma). We have examined the relationship between this global sea level history and the New Zealand and Australian continental margin shallow marine sedimentary record and the record of deposition, nondeposition and erosion in the deep sea. These in turn are examined in relation to the history of polar glaciation.

A series of sedimentary cycles in the Australian Cenozoic marginal marine sequences are bounded by unconformities of varying duration. The tectonic stability and aridity of the western and southwestern margin of Australia during the Cenozoic produced a sedimentary record dominated by hiatuses. This contrasts with the relatively completes sequences of the tectonically more active southeastern marginal basins. The four major sedimentary cycles in the Australian Cenozoic (Paleocene to Early Eocene ; Middle to Late Eocene ; latest Oligocene to late Middle Miocene and latest Miocene to Quaternary) are correlated with the supercycles Ta, Tb, Tc, Td and Te of Vail *et al.* (1977).

In New Zealand, the majority of the Cenozoic Stages represent classic sedimentary cycles bounded by unconformities or correlative conformities formed as a result of large rapid eustatic sea level changes. The marine Tertiary sequence in New Zealand consists of 23 stages. Of the 18 stage boundaries between the end of the Paleocene (53 Ma) and the end of the Miocene (5 Ma), 16 appear to correlate with the boundaries of eustatic sea level cycles.

Eustatic sea level lowstands are well recorded (as unconformities) in New Zealand as a result of its unique tectonic setting during the Cenozoic. For most of the early Cenozoic until the Middle Oligocene this region began to be uplifted as the Pacific-Australian plate boundary migrated onto New Zealand, climaxing in the latest Cenozoic (Late Pliocene-Quaternary), thus exposing a nearly complete sequence of marine Cenozoic strata well suited for the study of sedimentary cycles.

Unconformities, which form as a result of eustatic sea level changes, represent a very useful correlation tool, especially on continental margins with a sufficiently high terrigenous sediment supply and can be used to supplement paleontological correlations. Their ultimate usefulness for correlation depends on the rapidity of the sea level changes or the speed at which sedimentary facies change in response to sea level changes. If sea level changes rapidly (10 m/1,000 yrs) as suggested by Vail *et al.* (1977) then stratigraphic resolution will be high ($\pm 10^5$ yrs) but resolution decreases as the speed of sea level change decreases. The use of unconformities for stratigraphy may well prove to be a vital tool linking the classic land-based sections around the world with deep sea sections.

Eustatic sea-level change also significantly effects the supply of terrigenous and dissolved material to the deep sea and is inferred to partly control biogenic sediment production. Marine transgressions trap terrigenous material and organic carbon and restrict the supply of dissolved material to the open ocean thus increasing carbonate dissolution in the deep sea. The history of abundance of deep-sea hiatuses thus appears to be directly related to sea-level change. High sea levels result in maxima in hiatus abundance ; low sea levels in minima. Bottom water advection and the corrosiveness of oceanic waters to biogenic skeletal material also appear to directly affect the abundance of deep sea hiatuses.

Fluctuations in continental ice volume do not appear to control global sea level prior to the Miocene or perhaps the Oligocene. Thus another mechanism is required to explain the large, rapid falls of sea level reported by Vail *et al.* (1977) through the Early Cenozoic and Cretaceous. Despite the inferred Middle Miocene (~ 13 Ma) buildup of Antarctic ice, the character of individual sea-level cycles through the Late Cenozoic remains similar to that of the Middle and Early Cenozoic. However supercycle (Te) and the latest Miocene-Pliocene supercycle (Tf) exhibit mid-supercycle sea-level peaks distinguishing them from the earlier supercycles which terminate at times of highest sea level. This change in character may represent glacio-eustatic effects superimposed upon the longer term eustatic driving mechanisms.

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RÉSUMÉ

Cycles sédimentaires du Cénozoïque australien ; variations globales du niveau de la mer et histoire des sédiments en mer profonde.

L'histoire des variations globales du niveau de la mer au Cénozoïque (Vail *et al.*, 1977) est comparée à l'histoire des cycles sédimentaires sur les marges continentales de Nouvelle-Zélande et d'Australie.

En Australie, les séquences sédimentaires cénozoïques sont limitées par des hiatus de durée variable. Les hiatus sont importants sur les marges stables et maigres du Sud-Ouest, les séquences étant plus complètes sur les marges sud-orientales, tectoniquement plus actives. Les quatre cycles sédimentaires principaux se corrèlent avec les supercycles Ta à Te de Vail *et al.* (1977).

En Nouvelle-Zélande, les cycles sédimentaires sont limités par des discordances dues aux rapides variations eustatiques du niveau de la mer. 16 limites de cycles sur 18 sont corrélables aux limites des cycles de variation eustatique du niveau de la mer.

Les baisses eustatiques du niveau de la mer au Cénozoïque sont bien visibles dans les séries cénozoïques à terre en Nouvelle-Zélande, exposées à la faveur d'un soulèvement à la fin du Tertiaire.

Les discordances, dues aux variations eustatiques du niveau de la mer, permettent d'affiner les corrélations paléontologiques, surtout dans le cas des variations eustatiques rapides ($10\ m/10^3$ ans).

Les variations eustatiques contrôlent partiellement les apports terrigènes et les dissolutions en mer profonde. Ainsi, une transgression marine piégera le matériel terrigène et organique et réduira l'apport de matériel dissous vers l'océan, ce qui augmentera la dissolution des carbonates, et par voie de conséquence, fera apparaître un hiatus de sédimentation en mer profonde. La circulation de l'eau au fond de la mer et l'action corrosive des eaux profondes sur le matériel biogénique contrôlent directement l'abondance des hiatus profonds.

Les fluctuations du volume de glace sur les continents n'affectent les variations du niveau de la mer qu'à partir du Miocène. Le caractère des super-cycles individuels au Cénozoïque supérieur reste comparable aux cycles du Cénozoïque inférieur et moyen. Toutefois on reconnaît des cycles intermédiaires dans les super-cycles Mio-Pliocène. Il semble donc que les effets glacio-eustatiques se superposent aux mécanismes gouvernant l'eustasie à long terme.

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INTRODUCTION

The history of global-sea level change during the Cenozoic, produced by Vail *et al.* (1977) and revised by Vail and Hardenbol (1979), shows an overall decrease in sea level since the Cretaceous, punctuated by numerous rapid falls of sea level (Fig. 1). The documentation of eustatic sea-level fluctuations has been fundamental to our increased understanding of sedimentary cycles (depositional sequences) recorded on the continental margins. Eustatic sea-level changes in combination with regional tectonic activity and climate have played a critical role in the sculpturing of continental margins of the world. Also, sea-level fluctuations large enough to cause marine transgressions and regressions of the shoreline across the continental shelf, control the amount of terrigenous debris supplied to the

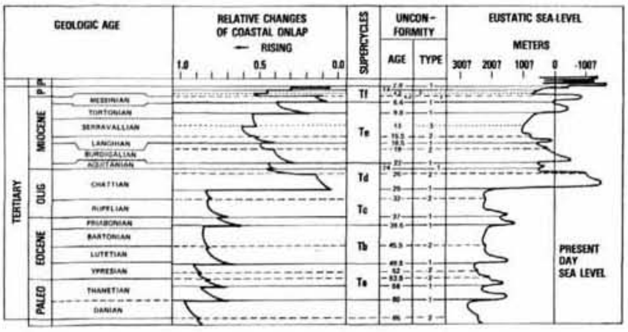


Figure 1

Global cycles of relative sea-level change during the Cenozoic and correlation with planktonic microfossil zonations, European stages and the geomagnetic time scale (after Vail, Hardenbol, 1979 ; Vail, 1980). Unconformity types (right column) are as follows : Type 1 indicates a rapid fall of sea level ; Type 2 indicates a rapid rise of sea level following a stillstand or slow fall ; and Type 3 indicates an increase in the rate of sea level fall, but less than the rate of subsidence of the continental margin.

deep sea and the supply of nutrients and elements necessary for biological productivity in the open ocean. Sea-level change therefore plays a major role in controlling deep-sea sedimentation.

Within the framework of sea-level history documented by Vail *et al.* (1977), we have examined the Cenozoic continental margin sedimentary record of New Zealand and Australia, the Cenozoic deep-sea sedimentary record and the Cenozoic polar glacial record. Our aim is to draw attention to relationships that exist between the various parameters. We also hope to bring into focus certain anomalous events that have become apparent from comparison of the records.

The classic, internationally recognized chronostratigraphic subdivisions (stages) of the Cenozoic of Europe were derived from shallow marine sedimentary sequences, at least partially controlled by eustatic sea level changes (Vail *et al.*, 1977). In addition there are a number of regional chronostratigraphic schemes throughout the world such as in California (Kleinpell, 1938), New Zealand (Thomson, 1916 ; Allan, 1933 ; Finlay, Marwick, 1940 ; 1947), Australia (Carter, 1959) and the Andaman-Car Nicobar Islands (Srinivasan, 1978) and others. Such stratigraphic subdivisions were largely established for two reasons : 1) to set up a standard of reference by which rocks from different parts of a country may be correlated with one another ; and 2) for interregional correlation (Thomson, 1916). These local chronostratigraphic schemes have formed a valuable base for regional correlation but interregional correlation has proven particularly difficult, especially with New Zealand due to the provincial nature of its flora and fauna. Most of the important zone defining microfossil species found in waters around New Zealand during the Cenozoic were absent from the equatorial regions, where the most widely used microfossil zonation schemes have been established (Bolli, 1966 ; Banner, Blow, 1965 ; Blow, 1969 ; Berggren, 1969 ; Martini, Worsley, 1970).

The Deep Sea Drilling Project has helped considerably to reduce New Zealand's stratigraphic isolation by supplying long relatively complete Cenozoic sequences within different water masses between New Zealand and the equatorial regions. Continued work with such sequences involving zonation schemes derived from all microfossil groups, stable isotope studies and paleomagnetic dating should result, eventually, in a truly interregional correlation scheme.

An approach that has perhaps been under utilized for interregional correlation is the use of unconformities as a stratigraphic tool. Nearly one hundred years ago Eduard Suess recognized the importance of sedimentary cycles in stratigraphy and attributed them to changes in sea level (Vella, 1965). Since Suess' time several workers have suggested that sedimentary cycles or depositional sequences may be recognized worldwide and thus may represent an important correlation tool (Stille, 1924 ; Wells, 1960 ; Hallam, 1963 ; Vella, 1965 ; 1967 ; and others). The mechanism that produces the unconformities which bound each sedimentary cycle has remained elusive. Recently Vail *et al.* (1977) rekindled interest in eustatic sea level changes as a mechanism controlling continental margin sedimentation and sedimentary cycles. However, the factors that control global sea level changes remain unclear. Vail *et al.* (1977) have shown that the boundaries (unconformities) between certain depositional sequences are global in nature and have suggested that many continental margin unconformities are formed during eustatic lowstands of sea level. As a result, petroleum exploration, using multichannel seismic reflection profiling techniques, now makes use of depositional sequences as basic subsurface mapping units over widely separated areas of the world. As early as 1960, Wells had noted that paleontological correlation can be supplemented or reinforced by matching successions of sedimentary cyles or the unconformities at the boundaries of sedimentary cycles. Outside of the petroleum industry they have not been widely used and are only briefly mentioned in the International Code of Stratigraphic Nomenclature (Hedberg, 1976).

In 1976, Vella recognized eight sedimentary cycles in the New Zealand Paleogene, four of which he correlated throughout New Zealand and possibly with southeastern Australia. Since Vella's classic study the concept of using sedimentary cycles as a stratigraphic tool has received no further attention in New Zealand.

In this paper we suggest that the majority of the New Zealand Cenozoic stages are classic sedimentary cycles bounded by unconformities (or their correlative conformities) formed as a result of eustatic sea level changes. The evidence is presented as a model which we hope will supplement and improve stratigraphic correlations within New Zealand and between New Zealand and other areas. This is particularly important in New Zealand at this time because of considerable efforts in intrabasinal correlation and sedimentary basinal analysis by the New Zealand Geological Survey.

The sedimentary sequences of the Australian and New Zealand continental margins have been extensively studied by many investigators and have provided an important basis for the study of temperate Cenozoic faunas and paleoclimates.

Our approach has been to examine the shallow marine sedimentary sequences on the Australian and New Zealand margins, looking primarily for sedimentary cycles or depositional sequences bounded by unconformities. To a first approximation unconformities in shallow marine sequences represent lowstands of sea level and result from non-deposition or subaerial erosion of the continental shelf. In contrast, shallow marine to nonmarine depositional sequences form during high-stands of sea level above the shelf edge, as for example during the Holocene transgression. We have compared the shallow marine record in the Australian region with the sea level history of Vail *et al.* (1977) and Vail and Hardenbol (1979) in order to differentiate those unconformities created by global sea-level change from those due to local tectonic factors. Most of our stratigraphic control is based on the first and last appearances of stratigraphically important planktonic foraminiferal species.

By comparing the shallow marine record with the curve of sea-level history, it is possible to differentiate those unconformities created by sea-level change from those controlled by local tectonic factors. Having made comparisons between the Australasian record of shallow sedimentary cycles and the global sea-level curve, we then compare these with the history of polar ice-volume change. This comparison has helped us to determine whether the sedimentary cycles and global sea-level curve have been controlled by ice-volume change at any time during the Cenozoic.

AUSTRALIAN SEDIMENTARY CYCLES

The Cenozoic segment of the Australian continental margin (Fig. 2) was formed essentially during four depositional sequences separated by unconformities (McGowran, 1979). Briefly (Fig. 3 *a*), they consist of 1) a Paleocene to Early Eocene period of deposition, bracketed by unconformities at the Cretaceous-Tertiary and Early-Middle Eocene boundaries, with maximum transgression onto the margin during the Late Paleocene (Foraminiferal Zone P4) ; 2) a Middle to Late Eocene deposition period which is most widespread in the Late Eocene (Foraminiferal Zone P15-P17) ; 3) a latest Oligocene to late Middle Miocene depositional period with the most prominent episode to transgression beginning in Foraminiferal Zones N3-N4 and reaching maximum extent near the Early-Middle Miocene boundary. However, at the time of maximum transgression a widespread hiatus occurs

Figure 2

Australian Cenozoic sedimentary basins (from McGowran, 1979). Diagonal hatching — Tertiary basins. Broken hatching — basins of non-marine sedimentation in the cratonic interior. Cross hatching — Tertiary sediments and volcanics within the New Guinea mobile belt.

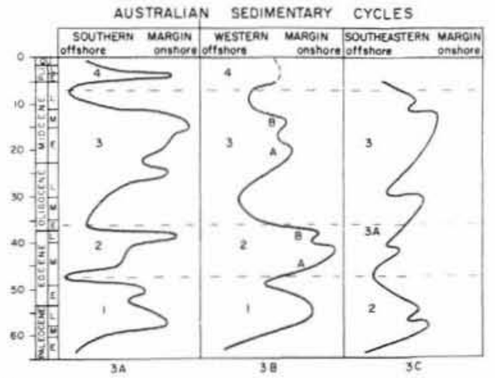


Figure 3

Australian Cenozoic sedimentary cycles. 3A : sedimentary cycles southern Australian divergent margin (after McGowran, 1979) ; 3B : sedimentary cycles in the western Australian margin (Quilty, 1980) ; 3C : sedimentary cycles in the southeastern Australian margin-Otway Basin (Glenie et al., *1968). Numbers applied to the succession of marine transgressions are those of the respective authors. Dashed lines represent boundaries between sedimentary cycles of McGowran (1979).*

during N8 to N9 (late Early Miocene) ; and 4) a latest Miocene to Quaternary depositional sequence is also recognized. Quilty (1977), in a study of Cenozoic sequences of the Western Australian margin, also recognized four major cycles of sedimentation but subdivided them further (Fig. 3 *b*). The duration of the sedimentary cycles in both the southern and western passive margins of Australia is remarkably similar.

The southeastern margin of Australia consists of a number of Tertiary marine basins, including the Gippsland, Otway and Bass Basins, which have been extensively explored for petroleum. The Gippsland Basin lies mostly offshore beneath the eastern edge of Bass Strait (Fig. 2). Steele (1976), using seismic methods, recognized a series of depositional sequences, primarily in the Latrobe Group and the Lakes Entrance Formation. By applying the methods of Vail *et al.* (1977) to these sequences, Steele (1976) obtained a record of the relative change of sea level during the Tertiary (Fig. 4). Since the Gippsland Basin has been subsiding during the Cenozoic the record shows an overall rise in sea level from the lowermost Tertiary until the Middle Miocene, punctuated by numerous rapid falls of sea level of short duration but varying magnitude.

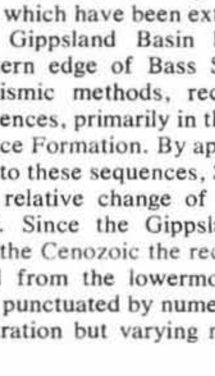


Figure 4

Relative sea-level history in the Gippsland Basin based on seismic stratigraphic techniques by Steele (1976). Note general rise in sea-level during Cenozoic, indicative of subsidence of the Gippsland Basin.

The Otway Basin lies to the west of the Gippsland Basin separated from it by the Port Phillip Basin (Fig. 2). The Tertiary sediments of the Otway Basin were deposited during two main depositional cycles : the first cycle extends from the Paleocene to the Early-Middle Eocene boundary. The second cycle extends from the Early-Middle Eocene to the Pliocene (Fig. 3 *c*) and is divided into two shorter cycles separated by a major regression during the Middle Oligocene (Glenie *et al.*, 1968). In contrast to the western and southwestern margin of Australia, the southeastern margin exhibits a relatively complete Cenozoic sedimentary sequence, as for example in the Gippsland (Steele, 1976 ; Partridge, 1976) and Otway Basins (Bock, Glenie, 1965 ; Glenie *et al.*, 1968). The most notable differences are first, a relatively complete Early to Middle Oligocene record is found in the southeast in contrast to the virtual absence of the Oligocene elsewhere in the Australian margin. Second, many smaller scale depositional sequences separated by unconformities can be recognized in the southeast in contrast to the western margin where broad depositional cycles span millions of years. Such differences result from different tectonic and climatic conditions around the Australian margin which, as in the southeastern margin, have undergone substantial change during the Cenozoic while remaining relatively constant in other areas such as the western margin. By comparing depositional sequences in such widely separated areas and by looking for isochronous sequence boundaries, it is possible to sort out local tectonic or climatic events from more extensive continent-wide events as shown by McGowran (1979).

McGowran (1979) demonstrated that major unconformities between depositional sequences are broadly isochronous over all areas of the Australian margin and that these boundaries transcend local tectonic and sedimentary environments (Fig. 3).

NEW ZEALAND SEDIMENTARY CYCLES

The New Zealand Cenozoic appears to represent one extremely long sedimentary cycle, beginning with subsidence of the Cretaceous land mass during the early Cenozoic (Fig. 5). Deposition of progressively deeper marine sequences and the associated transgression continued until the Middle Oligocene. After this, uplift and associated regression began as compressive stresses increased in response to migration of the boundary between the Pacific and Indian plates onto the New Zealand land mass (Weissel *et al.*, 1977 ; Carter, Norris, 1976 ; Nelson, Hume, 1977). As a result, a nearly complete classical transgressive-regressive sequence was preserved beginning and ending with shallow-marine to nonmarine sediments. Peak transgression during the Oligocene is marked by widespread limestones and fine grained upper bathyal deeper water deposits rich in pelagic microfossils (Brown *et al.*, 1968). Superimposed onto the Cenozoic cycle of sedimentation are a larger number of shorter sedimentary cycles which will be discussed later.

Again, we stress the fact that interregional unconformities appear to exist within New Zealand that appear to be related to unconformities in Australia (Vella, 1965). Thus any mechanism which produces sedimentary cycles bounded by such unconformities must act in unison in both regions.

AUSTRALIAN SEDIMENTARY CYCLES AND GLOBAL SEA LEVELS

Vail *et al.* (1977) recognized six supercycles of eustatic sea-level change during the Cenozoic (Fig. 1). The boundaries between supercycles occur at 60, 49.5, 39.5, 29, 22.5, 6.6 and 2.8 Ma. Each supercycle consists of a relatively slow overall rise of sea-level followed by a large (100-200 m) rapid (within a million years) drop of sea level. In the Early to Middle Cenozoic the supercycles are remarkably regular averaging approximately 10 Ma in duration. In addition to the supercycles, there are twenty-six cycles, each representing a single slow rise followed by a rapid fall of sea level.

The composite Cenozoic sedimentation history produced by McGowran (1979) for the Australian margin and the cycle histories of Quilty (1980) and Glenie *et al.* (1968) correlate well with supercycles Ta, Tb, Tc, Td and Te of Vail *et al.*

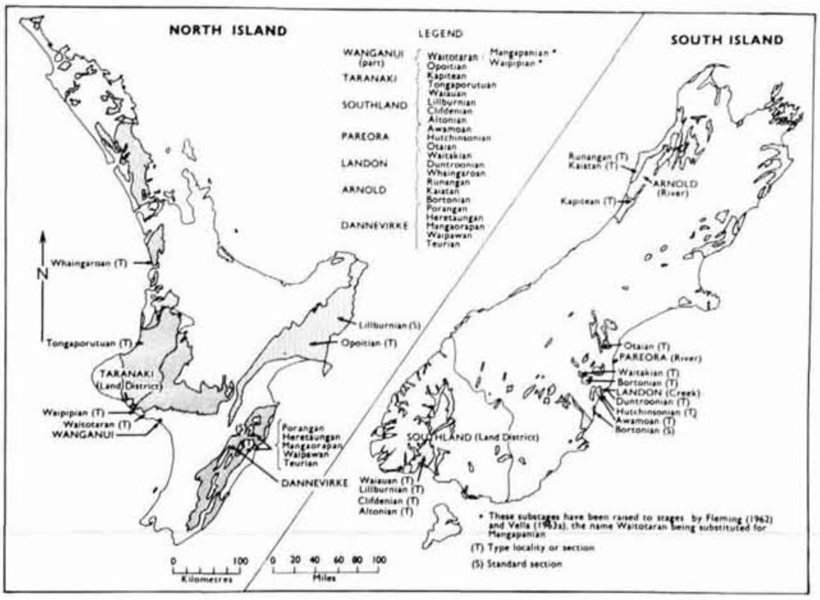


Figure 5
Location map of important New Zealand marine stratigraphic Cenozoic sections (Hornibrook, 1978).

(1977) (Fig. 6). However, there are a number of differences : the top of supercycle Td (23.5 Ma) has no counterpart in the Australian sequence ; the duration of the hiatuses between the Australian sedimentary sequences, especially the western margin, are considerably longer (a few million years) than the boundaries between supercycles (a million years) ; and the hiatus between cycles 3A and 3B (Quilty's terminology, 16-19 Ma) does not appear to be related to rapid, conspicuous sea level lowerings which mark the beginning of supercycle boundaries.

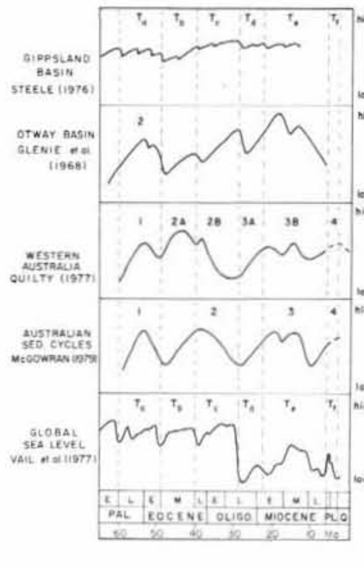


Figure 6

Correlation of several Australian sedimentary cycle histories with the global sea-level record of Vail et al. *(1977). Tertiary supercycles shown for global sea-level change.*

The absence of the upper part of supercycle Td throughout the passive margins of Australia requires explanation since supercycle Td is a major event in global sea-level history (Vail *et al.*, 1977). The very character of supercycle Td probably contains the explanation since it differs from most other supercycles in that it represented a time of particularly low global sea level. As a result it is associated with characteristic deep marine onlap patterns of marine facies (Fig. 7 *b*) formed as sediment bypassed the continental shelf and was deposited directly beyond the shelf edge.

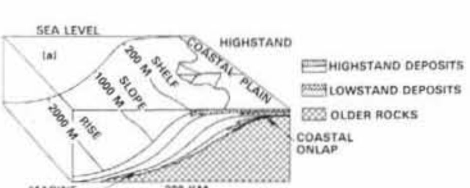


Figure 7

Depositional patterns (facies relationships) during a high-stand (a) and a lowstand (b) of sea level (after Vail et al., *1977).*

Since sea level was below the continental shelf edge for most of supercycle Td no deposition took place on the shelf and subaerial erosion probably occurred on previously deposited shelf sediments. Thus on the Australian margin we would not expect to find shallow shelf sediments during supercycle Td until sea level again rose above the continental shelf. This event is probably recorded by the onset of sedimentation at the beginning of cycle 3A during the latest Oligocene to earliest Miocene (N3-N4). Both Carter (1978) and McGowran (1979) report that sedimentation recommences contemporaneously around Australia with the deposition of calcareous clay and biogenic limestone facies in the Late Oligocene.

Hiatuses between sedimentary cycles in Australia, although wide-spread, vary considerably in their duration (McGowran, 1979 ; Quilty, 1977 ; Glenie *et al.*, 1968). In most areas, especially the western and southern passive margins, they are considerably longer than the duration of unconformities dividing the supercycles of Vail *et al.* (1977) (Fig. 8). We attribute the variability in duration to one major factor, that is, to varying amounts of subaerial erosion of older shelf sediments deposited during preceding highstands of sea level. We consider this to be largely due to the depth at which sea level dropped at the beginning of each supercycle. The lower the lowstand of sea level, the longer it takes for the sea to return to its previous position above the shelf edge and thus reinstate sedimentation. The duration of hiatuses will depend on the amount of time the sea stays below the shelf edge.

A major, possible complicating factor in the interpretation of sedimentary cycles in relation to the Vail *et al.* (1977) sea-level curve, is that hiatuses can also form at times of particularly high stands of sea level. Whenever a rapid rise of sea level occurs resulting in a highstand, terrigenous sediments are trapped in estuaries and other nearshore environments thus starving the continental shelf and ocean basins. Periods of non-deposition and even erosion thus result, creating unconformities in shallow marginal sequences. The Holocene transgression is an excellent example of

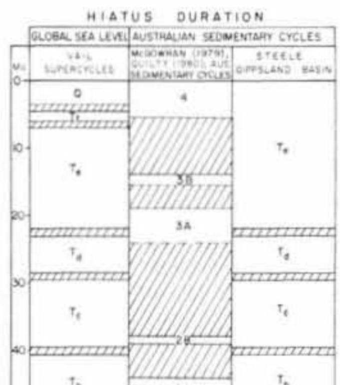


Figure 8

Comparison of the duration of hiatuses in the Australian continental margin sedimentary sequences with the times of unconformities separating the global sea-level supercycles of Vail et al. (1977).

this process (Emery, 1968). The hiatus between sediment cycles 3A and 3B in the late Early Miocene in Australia appears to have formed during such a highstand of sea level (Fig. 6). Highstands of sea level, especially after a rapid rise of sea level, often result in starvation of the middle and outer shelf areas. Vail (1980) reports a rapid rise of sea level beginning at 15.5 Ma which follows a rapid fall beginning at 16.5 Ma (Fig. 1). Such a rapid rise could have caused sediment starvation and hiatus formation on the Australian continental margin and may well correspond to the hiatus reported by Quilty (1980) and McGowran (1979) in planktonic foraminiferal zones N8 and N9. Glenie *et al.* (1968) record minor sea-level falls, superimposed upon times of peak transgression, in the Middle Paleocene and Middle Miocene (Fig. 3) which may reflect short-term starvation of the shelf regions as a result of sea-level highstands.

The only studies (Steele, 1976 ; Partridge, 1976) in the Australian region which have employed the seismic stratigraphic techniques of Vail *et al.* (1977) have clearly shown the existence of sea-level cycles. These sea-level histories exhibit patterns similar to those elsewhere in the world. In fact, Vail *et al.* (1977) were able to incorporate the Australian record into their global sea-level history. Steele (1976) and Partridge (1976), using seismic techniques and well data (palynology and electric-logs), defined depositional sequences in the Gippsland Basin of southeastern Australia and produced a relative sea-level record for a major portion of the Tertiary (Fig. 4). Sea-level fluctuations of various magnitudes are recorded in the Gippsland Basin sequences studied by Steele (1976). Vail *et al.* (1977) designate three orders of sea-level fluctuation (first, second and third order) based on the amplitude of sea-level change on continental margins around the world. The first order cycles (supercycles) of Vail *et al.* (1977) (lowstands at 60, 49.5, 39.5 and 29 Ma) are easily correlated with depositional cycle records around Australia (Quilty, 1977 ; Glenie *et al.*, 1968). The Gippsland Basin also contains second and third order cycles (Steele, 1976), which Vail *et al.* (1977) succeeded in correlating with sequences from the North Sea, northwest Africa and California (Fig. 9). The amplitude of the relative sea-level change is, however, much reduced in the Gippsland sequence because of tectonically controlled subsidence in the basin during the Cenozoic, as noted by Carter (1978). Depositional sequences around the Australian continental margin have been controlled primarily by eustatic sea-level changes and to a much lesser extent by local tectonic and climatic controlled changes in sediment supply.

The majority of first order sea-level fluctuations of Vail *et al.* (1977) are recorded in the Australian margin. Expansion of multichannel seismic profiling of subsurface sequence studies such as that of Steele (1976) around the Australian margin and more detailed stratigraphic investigations of uplifted sequences should further confirm the global nature of the sea-level history produced by Vail *et al.* (1977).

Glenie *et al.* (1968) reported that depositional cycle boundaries in the Otway Basin correlate with certain Australian Tertiary stage boundaries. If so, then the Australian Tertiary stages may reflect the effect of global sea-level history on the Australian margin and hence may prove useful for global correlation.

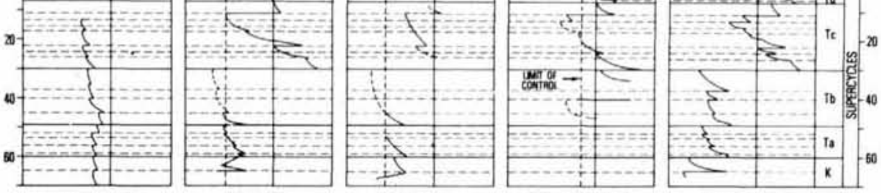
NEW ZEALAND SEDIMENTARY CYCLES AND GLOBAL SEA LEVEL CHANGE

In 1967 Vella suggested that essentially all of the New Zealand Stages represent single sedimentary cycles and that their boundaries correspond to cycle boundaries. This far-reaching conclusion was made by Vella after he had examined in some detail the Late Paleogene New Zealand Stages. He attempted correlations using planktonic foraminifera, with sedimentary sequences in Australia, North America and Europe. As a result, Vella (1967) believed that sedimentary cycles in New Zealand matched similar cycles in sedimentary sequences else-where in the world. Hornibrook (1967) questioned a number of Vella's (1967) correlations within the New Zealand Paleogene sequences which were accordingly taken into account by Vella in his placement of a number of the cycle boundaries.

The New Zealand stages are defined by paleontological criteria. Initial definition of the stages mainly utilized Mollusca and benthonic foraminifera. Very few planktonic foraminifera were employed in these early studies (Thomson, 1961 ; Allan, 1933 ; Finlay, Marwick, 1940, 1947). More recently, planktonic microfossils, in particular, foraminifera and calcareous nannofossils have been used for interregional correlation (Jenkins, 1966 ; Hornibrook, Edwards, 1971). However, correlations with Cenozoic tropical planktonic foraminiferal zonation schemes has proved to be very difficult.

Correlation problems primarily result from the fact that the present divisions of the Cenozoic in New Zealand are based on fossils that are found to be stratigraphically useful locally and that boundaries have been placed at the points in the sequence which are most easily identified paleontologically without regard for overseas correlation. Benthic fossils are used extensively and Mollusca became ingreasingly important in the Upper Cenozoic (Hornibrook, 1978).

Figure 9
Correlation of the sea-level histories between the Gippsland Basin and those of three continents. These were used to form the composite global sea-level curve (at right) by Vail et al. *(1977).*



Before examining the correlations between the New Zealand boundaries and global sea-level changes it is appropriate to look at the evolution of the New Zealand Tertiary classification system to understand why its subdivisions are related to sea-level changes. Initially, the stage boundaries were not arbitrarily selected. R. S. Allan (1933) proposed to select well defined type localities as a standard for each stage. The stages were defined using distinct stratigraphic breaks and macrofaunas as correlation tools. The stages named by Thomson (1916) and Allan (1933) were confined to sediments deposited in molluscan and brachiopod-rich shallow waters. Both Thomson and Allan were quite clear that they were establishing a set of time units corresponding to all of Cenozoic time. However, by restricting their work to shallow marine facies they were not able to obtain a complete Cenozoic sequence. Foraminiferal studies by H. J. Finlay in the 1930*s* culminated in Finlay and Marwick (1940 ; 1947) utilizing deeper water sedimentary facies to produce a much more complete New Zealand Cenozoic stage classification. However the stages defined by Allan (1933) represent well defined sedimentary cycles bounded by distinct stratigraphic breaks. Furthermore, the stages of Finlay and Marwick (1947) were defined largely by benthonic foraminifera, which are highly facies dependent. Changes in sea level can affect benthonic foraminiferal populations dramatically, particularly at depths shallower than 500M. Even at depths greater than 500M, benthonic foraminifera may respond indirectly to sea-level changes as distinct water masses migrate over the continental slope. Thus, although in many areas no stratigraphic breaks are apparent there are changes in benthonic foraminiferal populations related to water depth changes which have been used to define stage boundaries (Kennett, 1967). Such faunal events related to sea-level change were utilized by Vail *et al.* (1977) to date unconformities formed at shelf depths. By tracing each unconformity to its "correlative conformity" (Vail terminology) in deeper water facies the age of an unconformity may be obtained from microfossil evidence. We suggest that Finlay and Marwick (1947) inadvertently often chose stage boundaries based on benthonic foraminiferal assemblage changes that reflect the boundaries of sedimentary cycles in shallow marine sequences uplifted around New Zealand.

Also, studies of planktonic foraminifera increased dramatically in the late 1950s and 1960s. Planktonic foraminifera, because of their rapid evolution, wide distribution and abundance are an extremely useful correlation tool. As a result, many of the New Zealand stages are now recognized using planktonic foraminifera (Jenkins, 1971). For instance, Hornibrook (1969) suggested that the original boundary between the Clifdenian and Lillburnian stages is more *conveniently* defined using planktonic foraminifera evolution (the boundary between *Praeorbulina glomerosa circularis* and *Orbulina suturalis*) rather than stratigraphic breaks or benthonic foraminiferal assemblages as originally used. Hence we suggest that many of the New Zealand stage boundaries, as currently identified, only represent approximations to Allan's (1933) and Finlay and Marwick's (1940 : 1947) rigorously defined stage boundary definitions.

By using a number of datums considered to be reliable (Fig. 10) we have assigned ages to several of the New Zealand stage boundaries by correlating from the dated zonation schemes of Bolli (1966) and Berggren (1969) using age assignments of Berggren (1971) ; Saito (1977) ; Srinivasan and Kennett (1980) ; (Fig. 10). Figure 11 represents a review of those correlations that have been made by a number of workers between the temperate and tropical zonation schemes. A few modifications have been made in the light of a recent review by Srinivasan and Kennett (1980) of the Neogene planktonic foraminiferal zone boundaries (Fig. 11). The New Zealand Paleogene stage boundaries have been correlated by Berggren (1971) to the P zones of Berggren (1969).

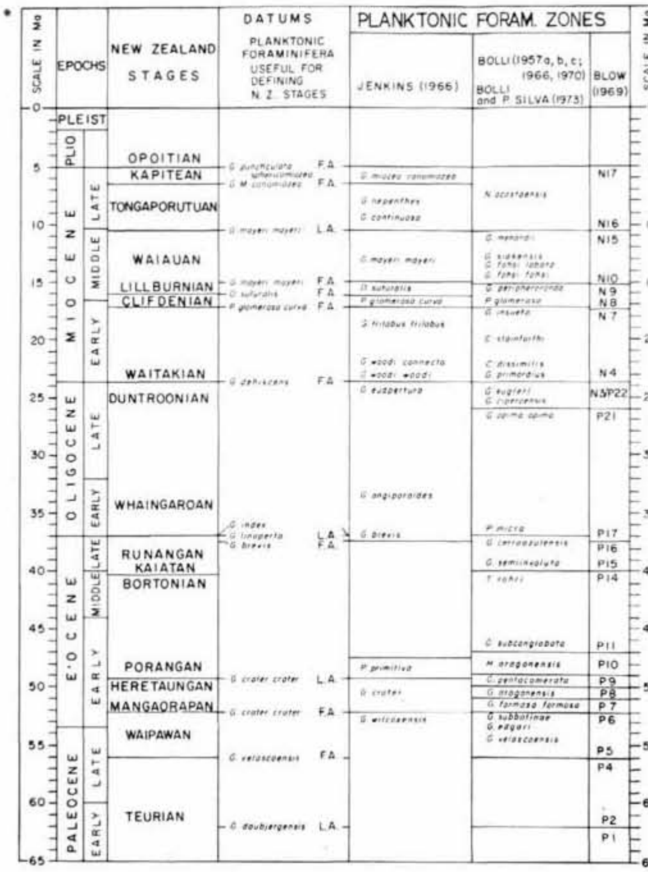


Figure 10

Chronology and correlation of New Zealand Tertiary stages with the tropical zonation schemes of Bolli (1957a, b, c ; 1966 ; 1970), Bolli and Premoli-Silva (1973) and Blow (1969), Berggren (1969) using selected planktonic foraminiferal datums.

Since sedimentary cycles are controlled by eustatic sea-level changes, they have proven to be extremely useful for worldwide stratigraphic correlation. Vella (1967) has shown their value in correlating the New Zealand Eocene stages with Northern Hemisphere sequences. The New Zealand stage boundaries appear to correlate with the sea-level cycles of Vail *et al.* (1977). Between the Cretaceous-Tertiary and Miocene-Pliocene boundaries Vail *et al.* (1977) record 20 sea-level cycles, incorporated within six supercycles. Within the same period there are 18 New Zealand stages.

CORRELATION OF THE NEW ZEALAND STAGE BOUNDARIES WITH GLOBAL SEA LEVEL CHANGES

Figure 10 shows the New Zealand stages, the New Zealand planktonic foraminiferal zones and datums used by Jenkins (1966 ; 1971) to delineate the zones. We first discuss the New Zealand stage boundaries, between the Cretaceous-

Figure 11

Correlation between New Zealand Tertiary stages, the New Zealand planktonic foraminiferal zonal scheme of Jenkins (1971) and the tropical planktonic foraminiferal zonal schemes of Bolli (1957 a, b, c ; 1966), Blow (1969), and Berggren (1969).

Tertiary and Miocene-Pliocene boundaries, that are most easily correlated with the tropical planktonic foraminiferal zonation schemes and hence with sea-level changes (Fig. 12). Planktonic foraminifera are the most well known and most reliable correlation tool available. Calcareous nannofossils are also useful but have not been as extensively studied as planktonic foraminifera in New Zealand.

The Cretaceous-Tertiary boundary in New Zealand is placed at the base of the Teurian. The Teurian-Waipawan boundary is correlated by Edwards and Hornibrook (1978) with the base of Bolli's (1966) *Globorotalia velascoenis* zone (P4/P5 ; Berggren, 1969) based on the first appearance of *G. velascoensis* in New Zealand and equatorial regions. The base of the *G. velascoensis* zone marks the boundary between sea-level cycles T.P. 2.1 and T.P. 2.2 at 56 Ma (Vail, Hardenbol, 1979).

The last appearance of *Globorotalia crater* defines the Heretaungan-Porangan boundary and is correlated with the base of Bolli's (1966) *Hantkenina aragonensis* zone (Berggren, 1971 ; Bolli, Krasheninnikov, 1977). Vail and Hardenbol (1979) dated a large rapid fall of sea-level at the Early-Middle Eocene boundary (49 Ma). Vail (1980) has since redated this event at 49.5 Ma within the *Globorotalia pentacamerata* zone (within P9 of Berggren, 1969) which is very close to the estimated age of the Heretaungan-Porangan boundary in New Zealand (49 Ma ; Edwards, Hornibrook, 1978).

The Bortonian-Kaiatan boundary occurs in the lower part of the *Globorotalia inconspicua* zone (Jenkins, 1971, 1974 and marks the boundary between the Middle and Late Eocene in New Zealand. Vail *et al.* (1977) record a rapid fall of sea level at this boundary between P14 and P15 (~ 39.5 Ma).

The Eocene-Oligocene boundary in New Zealand occurs at the Runangan-Whaingaroan boundary and is now well correlated with the tropical Pacific by the use of oxygen isotopic evidence (Keigwin, 1980). Keigwin suggested that the last appearance of *Globigerapsis index* and *Globigerina linaperta* in DSDP Site 277 near New Zealand is synchronous with the last appearance of *Hantkenina primitiva* and *Globorotalia cerro-azulensis* in the tropics during zone P17 after he recorded similar dramatic shifts in the oxygen isotopic records in both areas. The last appearance of the above species is recorded at the rapid isotopic shift. Vail *et al.* (1977) recorded a sea-level drop in the middle of zone P17. More recent work by Vail (pers. comm. to T. S. L.) has shown that this sea-level fall was rapid and more significant than originally suggested.

A large sea-level fall recorded by Vail *et al.* (1977) in the Middle Oligocene dated at 29 Ma (Vail and Hardenbol, 1978) appears to be synchronous with a relatively prominent unconformity (the Marshall Paraconformity) at the base of the Duntroonian, particularly in the South Island of New Zealand (Carter, Landis, 1972). The Duntroonian stage is not easily distinguished on the basis of planktonic foraminifera of calcareous nannofossils and interregional correlations is therefore difficult.

The base of the Miocene in New Zealand has been placed at the base of the Waitakian coincident with the first appearance of *Globoquadrina dehiscens* (Jenkins, 1966 ; 1971). The base of the Miocene is usually placed at the first appearance of the *Globigerinoides* lineage (Hardenbol, Berggren, 1978). Srinivasan and Kennett (1980) in a study of Neogene planktonic foraminiferal datums in a number of DSDP sites between New Zealand and tropical regions consider that *G. dehiscens* is a reliable datum to mark the base of the Miocene in higher latitudes thus lending support for Jenkins (1966 ; 1971) placement of the boundary at the base of the Waitakian. However, coccolith evidence (the last appearance of *Reticulofenestra bisecta* and *Zygrhbalithus bijugatus*) indicates that the lower part of the Waitakian is Chattian (Oligocene) in age (Hornibrook, pers. comm. to J. P. K.). Hornibrook also reports that there is a stratigraphic break near the base of the type section of the Waitakian Stage. A rapid fall in sea level is recorded by Vail *et al.* (1977) at the Oligocene-Miocene boundary which apparently correlates with the lower Waitakian stage.

The Altonian-Clifdenian boundary is correlated with the N7-N8 zone boundary using the first appearance of *Praeorbulina glomerosa curva* and is coincident with the boundary between sea-level cycles T.M. 1.3 and T.M. 2.1.

The extinction of *Globorotalia mayeri mayeri* occurs at the base of the Tongaporutuan and is correlated with the base of N15 (Srinivasan, Kennett, 1980). The base of N15 is at approximately 10.5 Ma and occurs within 0.7 Ma of a major, rapid sea-level fall recorded by Vail and Hardenbol (1979) at. 9.8 Ma.

Within the latest Miocene, the Kapitean stage coincides with the appearance of *Globorotalia conomiozea* and correlates with the middle of zone N17 (Kennett, 1966). Vail *et al.* (1977) record a rapid sea-level fall at 6.6 Ma in the middle of

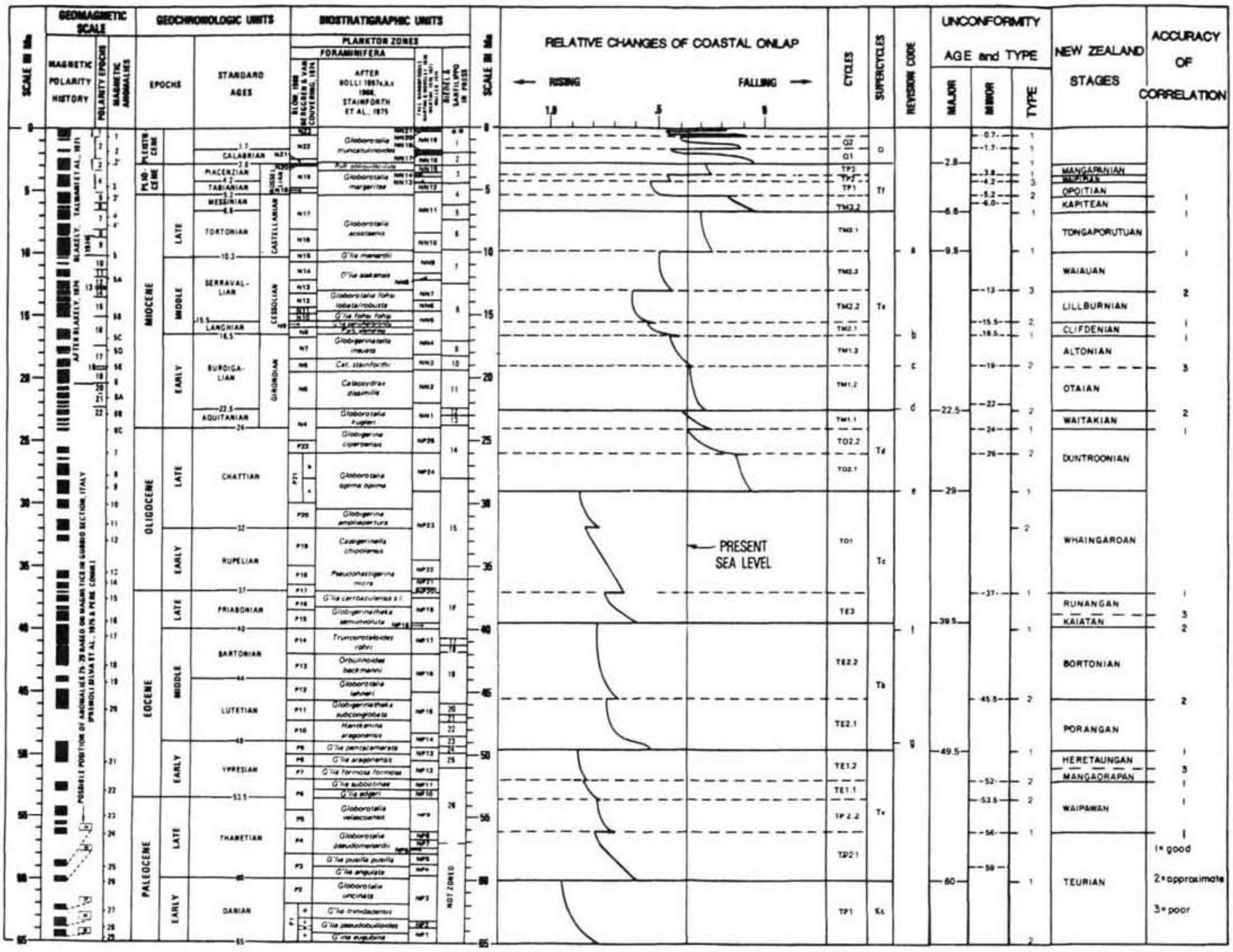


Figure 12
Chronology and correlation of New Zealand Tertiary stages with the global sea-level cycles of Vail et al. *(1977). Reliability of correlations (right column) : 1 = good ; 2 = approximate ; 3 = poor. Character of unconformities and other information as in caption for Figure 1.*

zone N17. Loutit and Kennett (1979) dated the first appearance of *G. conomiozea* at ~ 6.1 Ma in the Blind River section of New Zealand. Within the error of the dating technique it is quite likely that the base of the Kapitean and the sea-level fall at 6.6 Ma are coincident.

All of the preceding correlations link New Zealand stage boundaries and rapid falls of sea level which are marked by Type 1 unconformities as defined by Vail (1980) (Fig. 13). Vail (1980) records one other Type 1 unconformity at 60 Ma that does not appear to be recorded in New Zealand. Stratigraphic control is poor in the Teurian but Hornibrook (Fleming, Hornibrook, 1959 ; and pers. comm. to J. P. K.) reports that there are breaks and changes in sedimentary character in the Teurian type section at Te Uri Stream which may have resulted from eustatic sea-level changes.

Unconformities have also been produced by rapid rises in sea level which effectively starve the mid to outer shelf and deeper water regions. These unconformities have been called Type 2 unconformities (Fig. 13) by Vail (1980). Several sea-level cycle boundaries bounded by Type 2 unconformities correlate with New Zealand stage boundaries.

The initial appearance of *Globorotalia crater* defines the Waipawan-Mangaorapan boundary and is correlated, via its close relative *Globorotalia caucasica*, with the base of P7

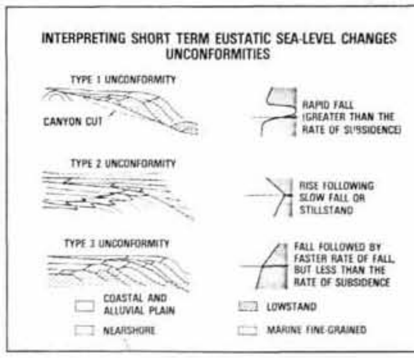


Figure 13
Types of unconformities produced by sea-level changes (Vail, 1980).

and the base of the *Globorotalia formosa formosa* zone of Bolli (1966). A rapid rise of sea-level occurs at the base of P7 (Vail, Hardenbol, 1979).

The Porangan-Bortonian boundary occurs within the *Globigerinatheka index index* zone of Jenkins (1971) which is correlated with the *Globigerinatheka subconglobata subconglobata* zone of Bolli (1966) by Hornibrook (1978) and Bolli and Krasheninnikov (1977). Vail and Hardenbol (1979) record a rapid rise of sea level at the top of P11 (45.5 Ma) close to the Porangan-Bortonian boundary.

A rapid rise of sea level is recorded by Vail and Hardenbol (1979) at the zone N4-N5 boundary (22,5 Ma). The Waitakian-Otaian boundary occurs in the middle of the *Globigerina woodi connecta* zone which approximately correlates with the N4-N5 zone boundary (Hornibrook, 1978). Hornibrook (1959) reported that an erosion interval occurring between the Waitakian and Altonian stages on the east coast of the North Island of New Zealand has removed most of the Otaian sediments. This erosion event may have occurred as a result of a rapid rise of sea level reducing sediment supplies to the mid outer shelf and deeper regions of the margin.

The Clifdenian-Lillburnian boundary is correlated with the base of N9 using the first appearance of *Orbulina suturalis* and thus may be related to a rapid rise of sea-level at zone N9-N10 boundary (15.5 Ma ; Vail, 1980).

The top of the Kapitean is equated with the Miocene-Pliocene boundary in New Zealand and has been dated at 5.3 Ma by Loutit and Kennett (1979). Vail and Hardenbol (1979) record a rapid rise of sea level near the N17 and N18 boundary at the Miocene-Pliocene boundary (5.2 Ma).

A third type of unconformity (Type 3) results from a slow fall of sea level immediately followed by an accelerated fall (Fig. 13) (Vail, 1980). Such a fall occurs near the base zone N13 (13 Ma). The Lillburnian-Waiauan boundary which occurs within the *Globorotalia mayeri mayeri* zone (Srinivasan, Kennett, 1980) is close to the base of N13.

To summarize : sixteen out of eighteen New Zealand Cenozoic stage boundaries are correlated either exactly or within a few hundred thousand years of global sea-level changes dated by Vail and Hardenbol (1979) and Vail (1980). Thus we suggest that the majority of New Zealand stages (16 of 18) between the Cretaceous-Tertiary and Miocene-Pliocene boundaries represent sedimentary cycles controlled by global sea-level changes.

The Mangaorapan-Heretaungan and Kaiatan-Runangan boundaries clearly do not correlate with any sea-level changes. Also the Otaian-Altonian boundary does not appear to correlate with any change in sea level. However a small rapid rise of sea level (producing Type 2 unconformities) within zone N6 has not been well defined biostratigraphically (P. R. Vail, pers. comm. to T. S. L.). Further work may show that the Otaian-Altonian boundary correlates with the rapid sea-level rise.

Also, rapid rises of sea level in P6 (53.5 Ma) and at the base of P22 (26 Ma) are apparently not related to New Zealand stage boundaries. The rapid rise of sea level at the base of P22 in the Late Oligocene, following the dramatic sea-level fall in P21, may be represented by the resumption of sedimentation on the continental shelf at the base of the Duntroonian stage. The Duntroonian stage (Middle Oligocene) is generally represented as a widespread unconformity in the Australasian region, and possibly elsewhere, named the Marshall Paraconformity by Carter and Landis (1972). This was postulated to have formed as a result of submarine erosion at the initiation of the circum-Antarctic current in the middle Oligocene (Carter and Landis, 1972). We suggest instead, that this event occurred as a result of the dramatic sea-level fall recorded by Vail *et al.* (1977) in zone P21 (29 Ma) in the middle Oligocene. The lack of sedimentation and apparent widespread slow sedimentation in the shallow water facies of the Duntroonian stage (the type section for the Duntroonian stage is only 1.3 M thick) may have been caused by a cessation of sedimentation when sea level dropped rapidly over the shelf-slope break. Sedimentation did not resume until sea level again reached inner shelf depths. A more complete Duntroonian stage (sedimentary cycle) should thus be found in deeper sections, either fully pelagic facies or in marine onlap sections formed as sediment bypassed the continental shelf and was deposited directly beyond the shelf edge. In such sections the Duntroonian stage could be dated micropaleontologically and thus more easily correlated within New Zealand and elsewhere.

Finally a fall in sea level in the middle Oligocene (P19-P20 boundary) may correspond to the boundary between two depositional sequences that Vella (1967) recorded in the Whaingaroan.

DISCUSSION

The amount of time represented by each hiatus between sedimentary cycles is dependent upon the position of the section on the continental margin (i.e., water depth) at the time of deposition, the speed of sea-level change, the length of time that sedimentation ceased and the location and amount of erosion.

The position of the type and standard section for each stage is critical. Sections which always remain submerged (upper to mid-slope depths) will be more complete than sections which are periodically exposed during sea-level lowstands (continental shelf). It is probable that many of the New Zealand Cenozoic and perhaps Mesozoic stages, which are defined on the basis of shallow water Mollusca or benthonic foraminifera, do not represent complete sedimentary cycles, i.e., do not represent all of geologic time. For instance, the base of the Kapitean stage (latest Miocene) which is defined in its type location at Kapitea Creek, is represented by an unconformity and its basal part by a greensand of probable slow deposition. However the Kapitean is currently defined by the evolutionary first appearance of the planktonic foraminifera *Globorotalia conomiozea* which occurs below the base of the Kapitean type section. The Kapitean stage is thus now defined from other more complete (and deeper water) sections other than the type section.

The Cenozoic tectonic development of New Zealand has resulted in the uplift of marine basins with widely differing depth histories. By tracing sedimentary cycle boundaries (unconformities) in shallow basins to their correlative conformities in deeper basins it should be possible to designate new, more complete sections which represent all of the time within each sedimentary cycle. To some extent this has already been done by Finlay and Marwick (1940, 1947) who often drew knowledge from other temporally longer sections (standard sections) to build up a composite picture of the faunas found within a particular stage (see Scott's discussion of the Otaian stage, 1960). Finlay and Marwick (1940 ; 1947) did not adhere to strict type locality definitions.

A reappraisal of the New Zealand type and standard sections is required to test our model that the majority of the stages represent sedimentary cycles produced by eustatic sea-level changes. Each important section must be reexamined, unconformities defined, traced into conformable sequences and dated. Microfaunas must be reevaluated in terms of water depth information. We recommend that type localities and section should be retained but new standard sections should be designated which full represent the time within each sedimentary cycle. It is important to make full use of multichannel seismic reflection profiles and offshore drill-hole data to obtain maximum control of the ages of the important unconformities found on the continental margins.

Sedimentary cycles produced by global sea-level changes are by definition bounded by unconformities. If global sea-level changes occur as fast as 10 M/1,000 yrs, as suggested by Vail *et al.* (1977), the ages of the sedimentary cycle boundaries are essentially isochronous (Excepting cases of erosion resulting from marine transgressions and regressions). If, however, the marine transgressions and regressions take place over several hundred thousand years or more, the unconformities are diachronous even where no erosion takes place. If we are to fully utilize unconformities and their correlative conformities as time planes, then it is important that more understanding is developed concerning the rapidity and magnitude of sea-level changes during the Cenozoic. Isochronous markers are central to accurate global correlation and an understanding of the speed and magnitude of the sea-level changes will ultimately determine the accuracy of this method in global correlation.

In conclusion we suggest that :

1) Sixteen out of eighteen New Zealand stage boundaries between the Cretaceous-Tertiary boundary (65 Ma) and the Miocene-Pliocene boundary (5 Ma) correlate (± a few hundred thousand years) with global sea-level changes reported by Vail *et al.* (1977) and Vail and Hardenbol (1979).

2) The majority of New Zealand stages within this time period represent natural sedimentary cycles bounded by unconformities or correlative conformities.

3) The New Zealand stages were established largely on the basis of lithofacies changes, Mollusca, and shallow water benthonic foraminifera all of which were susceptible to global sea-level oscillations. The currently defined New Zealand stages are approximations of the original stage definitions and are based predominantly on deeper water benthonic foraminifera or open ocean planktonic foraminifera.

4) A one-to-one correlation does not exist between the New Zealand stage boundaries and global sea-level changes during the Cenozoic. Some stage boundaries do not correlate with the global sea-level changes reported by Vail *et al.* (1977). Also a number of sea-level changes are apparently not recorded in the New Zealand sedimentary record.

5) Sedimentary cycles of the Tertiary of Australia can likewise be explained by global sea-level changes identified by Vail *et al.* (1977). However, the extent of the unconformities bounding sedimentary cycles is also controlled by the tectonic and climatic character of the Australian continental margin. Unconformities in the divergent, tectonically quiet southern and western margins, which have received little terrigenous sediment during the Tertiary, represent more time than unconformities on the tectonically more active southeastern margin where sediment supply was high during the Tertiary.

6) Many of the sedimentary cycle boundaries in Australia and New Zealand were produced by rapid falls of sea level. However, a number of the cycle boundaries were produced by rapid rises of sea level.

7) Global sea-level changes manifested as unconformities contain much potential for global correlation assuming that sea-level changes are rapid and that sedimentation processes respond quickly to such changes.

8) The beautifully exposed Cenozoic sequences in New Zealand represent uplifted sedimentary basins of widely differing water depths which provide an ideal setting in which to test these ideas. Also multichannel seismic reflection profiles and drill hole data which are now available in certain areas around New Zealand will increase the stratigraphic control (both areally and temporally) needed to trace unconformities to their correlative conformities in deeper water sections offshore.

DEEP-SEA HIATUSES AND GLOBAL SEA LEVEL CHANGE

Erosion, non-deposition, or sediment accumulation on the sea floor is determined by the dynamic balance between the rate at which sediments are supplied to the bottom and the rate at which they are removed (Moore *et al.*, 1978). Sediment supply to the deep sea is predominantly of two types : 1) terrigenous detritus and 2) biogenic material.

In a recent review of Cenozoic hiatuses in the deep sea (Moore *et al.*, 1978) it was suggested that the primary cause for widely distributed hiatuses (in time and space) has been sediment erosion and corrosion by oceanic bottom waters. Maxima in hiatus abundance appear to be related to changes in the boundaries of ocean basins particularly those changes that affected the high latitude regions of bottom water formation. The effects of eustacy on the deep-sea record have been grossly neglected. Eustatic changes in sea level can have a dramatic effect on the distribution and timing of deep-sea sedimentation (Rona, 1973 ; Davies *et al.*, 1977 ; Hay, Southam, 1977 ; Worsley, Davies, 1979 ; Arthur, 1979 ; Davies, Worsley, 1980).

To a first approximation, the mass of sediment supplied to the deep sea is directly proportional to the area of the continental land mass exposed and increases exponentially with the elevation of the land. The mass of material supplied in solution to the deep sea is proportional to the area of the continent exposed (Hay, Southam, 1977). However, local tectonic activity and climate can significantly affect such patterns. Generally when sea level is low, larger areas are available for erosion but elevation does not necessarily change. For example, if sea level dropped to below the shelf edge today, the area of the continents would increase by ~ 19 percent. Also, when sea level is below the shelf edge, rivers tend to supply material directly to the deep sea. Canyon cutting may be initiated which serves to further channel terrigenous material into the deep basins. At times of sea-level highstands, especially following a rapid rise of sea level, as in the Holocene, sediment supplied by rivers tends to be trapped in drowned river valleys (estuaries) close to the shoreline. The middle to outer shelf region becomes starved of sediment and very little sediment is supplied to the deep sea. In some regions, such as the western margin of Australia, sediment supply from the adjacent low-lying, arid landmass is small and sea-level fluctuations exert very little control on deep-sea terrigenous sedimentation. Also parts of the continental margin of western Australia are constructed of pelagic carbonates which often become diagenetically altered and thus more resistant, as a result of subaerial exposure during sea-level lowstands (Hay, Southam, 1977). As a result, probably very little carbonate debris was supplied to the deep sea from these regions during the Cenozoic.

Berger (1970 ; 1976) and Broecker (1980) have also suggested that eustatic sea-level fluctuations may exert a significant effect on biogenic productivity in the pelagic realm. During times of sea-level lowstands, large quantities of organic carbon deposited during a previous highstand may be eroded or dissolved from the shelf and returned to the oceanic reservoir. Broecker (1980) postulated that during times of low sea level, biogenic productivity should increase, thus supplying more biogenic material to the sea floor. Conversely, during times of high sea level, biolimiting nutrients are trapped on the continental shelf and primary productivity in the open ocean should thus decrease. In summary, lowstands of sea level result in increased terrigenous supply, increased supply of dissolved material and an increase in the supply of nutrients to the deep sea. We have examined the deep-sea hiatus record of Moore *et al.* (1978) in relation to the above models.

THE CENOZOIC DEEP-SEA HIATUS RECORD

Maxima in hiatus abundance in the world ocean (a composite of all oceans) occur at the Cretaceous-Tertiary boundary (65 Ma) ; in the Late Paleocene (53-55 Ma) ; and Late Eocene (42-43 Ma) ; the latest Eocene (37-39 Ma) ; the late Early Miocene (17-18 Ma) ; the late Middle Miocene (10-12 Ma) ; and in the Pliocene (3-4 Ma). We have examined hiatuses in the southwest Pacific because this oceanic region is adjacent to a tectonically active landmass which has supplied sediment throughout the Cenozoic. We contrast this with the eastern Indian Ocean which is adjacent to a tectonically quiet passive margin with restricted sediment supply during the Cenozoic.

In the southwest Pacific (Fig. 14), at the same time that sea level rose between 65 and 50 Ma, a hiatus abundance maximum is recorded. Between 58 and 60 Ma, a hiatus minimum occurs immediately following a rapid drop of sea level to a lowstand at 60 Ma. The next major hiatus peak occurs at the Paleocene-Eocene boundary which is also a time of high sea level. In fact, two rapid rises of sea level occur at 53.5 and 52 Ma. At approximately 48 to 49 Ma, a minimum in hiatus abundance is recorded which corresponds to a time of sea-level lowstand at 49.5 Ma. The Eocene-Oligocene boundary is marked by a major peak in hiatus abundance which continues into the Early Oligocene. Sea level was relatively high through this period but exhibits two small but rapid falls at 39.5 and 37 Ma which correspond to minima in hiatus abundance. Hiatus abundance decreases dramatically around 30 Ma and fluctuates rapidly thereafter. A large decrease in hiatus abundance occurs at approximately 15 Ma following a pronounced maxima between 19 and 16 Ma which is probably related to rapid rises of sea level at 19 and 15.5 Ma separated by a rapid fall at 16.5 Ma. A further large decrease occurs during the Pliocene to Quaternary. The correspondence between global sea level and deep-sea hiatus abundance is remarkably consistent in the southwest Pacific.

An almost identical, but more subdued hiatus abundance record is recorded in the East Indian Ocean (Fig. 15) where

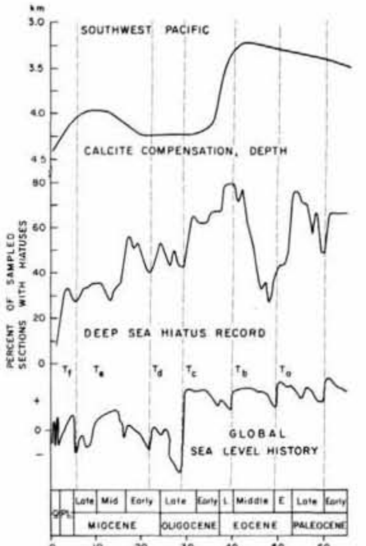


Figure 14

Comparison of the southwest Pacific Cenozoic deep-sea hiatus record (Moore et al.*, 1978) with the global sea-level record of Vail* et al. *(1977) and the history of the calcite compensation depth (CCD) in the Pacific (van Andel, 1975). Dashed lines represent boundaries between sea-level supercycles of Vail and Hardenbol (1979).*

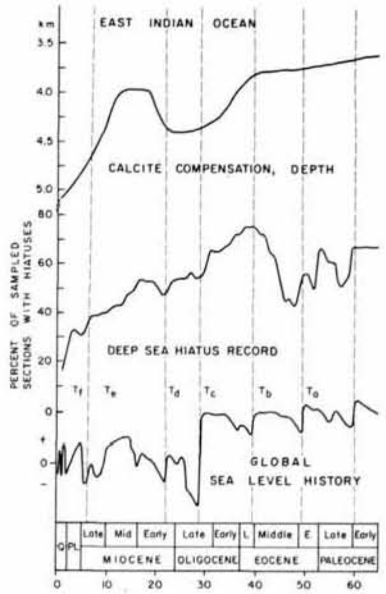


Figure 15

Comparison of the eastern Indian Ocean deep-sea hiatus record (Moore et al.*, 1978) with the global sea-level record of Vail* et al. *(1977) and the history of the calcite compensation depth (CCD) in the Indian Ocean (van Andel, 1975). Dashed lines represent boundaries between sea-level supercycles of Vail and Hardenbol (1979).*

terrigenous sediment supply has been restricted during most of the Cenozoic. The remarkable correspondence between the two records and their consistent relationship to sea-level changes suggest that global sea level may play an important role in controlling sedimentation in both regions.

A specific and clear demonstration of the role that global sea level plays in controlling the supply of sediment to the deep sea was given by Partridge (1976). He compared the sedimentary histories in the Gippsland Basin of southeastern Australia and DSDP Site 283 in the Central Tasman Sea (Fig. 16). Times of sediment deposition at Site 283 were considered to represent periods when fine-grained sediments were bypassing the continental shelf, reflecting the availability of sediment to the deep basins. The timing of commencement and termination of sedimentation at site 283 correlates with depositional sequence boundaries in the Gippsland Basin which correspond to eustatic cycle boundaries. Thus transport of sediment to the Tasman Sea from southeast Australia occurred during times of lowstands of sea level (Partridge, 1976).

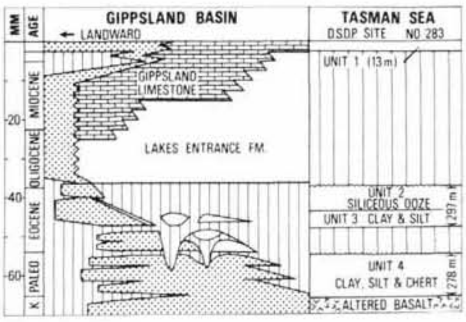


Figure 16

Correlations between the Gippsland Basin and DSDP Site 283 showing the timing of og commencement and termination of sedimentation during the Cenozoic (Partridge, 1976). Vertical lines indicate intervals of non-deposition and/or erosion.

We have also examined the relationship between the corrosiveness of oceanic bottom waters to calcium carbonate, as reflected by the calcite compensation depth (CCD) during the Cenozoic, and its effect on hiatus abundance. In the Pacific, the CCD shoaled during the Early Cenozoic, reaching its highest point at the Eocene-Oligocene boundary, reflecting increasing corrosiveness of bottom waters. The major peak in hiatus abundance in the southwest Pacific occurred at the Eocene-Oligocene boundary, corresponding to the shallowest level of the CCD in the Cenozoic and relatively high sea level. At the Eocene-Oligocene boundary the CCD fell dramatically and, with the exception of a brief shallowing in the Miocene, continued to fall to its present day level. In the East Indian Ocean a similar relationship is apparent between hiatus abundance and the CCD history during the Cenozoic (Fig. 15).

Comparison of the CCD curves with the hiatus abundance curve shows the important relationships discussed by Moore *et al.* (1978). Specifically, that shallow CCD times correspond to general periods of hiatus abundance and conversely times of deep CCD correspond with times of minima of hiatus abundance.

The Cenozoic deep-sea hiatus record is more easily understood when examined in relation to both the CCD record and global sea-level history since both play a major role in controlling hiatus abundance in the deep sea. Two periods of time in the Cenozoic demonstrate this.

In the southwest Pacific and East Indian Oceans, a minimum in hiatus abundance occurs at the Early-Middle Eocene boundary (48 Ma) which corresponds to a global lowstand of sea level. Also the CCD was nearly at its shallowest depth in the Cenozoic (Fig. 14 and 15). Carbonate preservation was probably poor although the supply of both carbonate and organic carbon to the deep sea should have increased during the lowstand of sea level. Thus the minima in hiatus abundance in this region probably resulted from increased supply of terrigenous material to the deep sea during the global low-stand of sea level.

In the Middle Oligocene (30 Ma), following a dramatic increase in carbonate preservation at the Eocene-Oligocene boundary (deepening CCD), a marked drop in hiatus abundance is recorded in both the southwest Pacific and eastern Indian Ocean (Moore *et al.*, 1978). At the same time, sea level dropped dramatically, thus resulting in increased terrigenous sediment supply and increased biogenic productivity. These effects, when combined with a moderately deep CCD (4.3-4.4 km), produced a minimum in hiatus abundance at approximately 30 Ma.

In summary, it appears that global sea level in combination with local tectonic and climatic conditions plays a critical role in controlling deep-sea sedimentation. Sea level controls terrigenous sediment supply to the deep sea and probably controls pelagic biogenic productivity. However, sea level is only one of a number of factors including deep-sea bottom water advection and corrosiveness which control sedimentation in the deep sea. It is significant that the global sea-level record of Vail *et al.* (1977) provides a fundamental framework to explain the deep-sea hiatus record.

POSSIBLE CAUSES OF RAPID SEA-LEVEL FLUCTUATIONS

Eustatic sea-level changes may be caused by changes in the volume of the ocean waters or change in the volume of the ocean basins. Pitman (1979) summarized a number of specific causes which are briefly discussed here. Fluctuations in the volume of continental ice may have produced rapid sea-level changes of up to 1,000 cm/1,000 years (Pitman, 1978). No other known mechanism can create such rapid change. Differentiation of mantle materials during plate tectonic processes, sediment deposition and removal in the ocean, crustal shortening (continental collision) and production and destruction of mid-plate volcanic chains all produce relatively small changes in the rates of eustatic sea-level change (0.02 to 0.2 cm/1,000 yrs, Pitman, 1978).

Changes in the volume of the mid-oceanic ridge system may produce major fluctuations in sea level (Hallam, 1963 ; Russell, 1968 ; Menard, 1969 ; and Valentine, Moores, 1972). Hays and Pitman (1973) showed quantitatively that the Late Cretaceous transgression and the Cenozoic regression could be explained by volume changes of the mid-oceanic ridge system. Volume changes may result from increasing or decreasing spreading rates, creation of a new ridge system by rifting or destruction of an old ridge by cessation of spreading or by subduction. Ridge volume, and hence sea level, is a function of spreading rate and ridge length at any given time (Pitman, 1978). Pitman estimated that changes in the geometry of the mid-oceanic ridge system could produce, at most, a 1 cm/1,000 years change in sea level. He concluded that, in the absence of continental ice, fluctuations of eustatic sea level are primarily due to changes in the volume of the mid-oceanic ridge system (1 cm/1,000 yrs). Also Pitman (1979) assumed that, based on the work of Steckler and Watts (1978), divergent margin subsidence is primarily driven by thermal processes, and proceeds at a rate equal to or greater than 2 cm/1,000 years. Therefore, in the absence of continental ice fluctuations, the shoreline on divergent margins should always remain above the continental shelf edge.

Hays and Pitman (1973) have shown that the large overall fall in sea level since the Late Cretaceous is the result of a decrease in the volume of the mid-oceanic ridge system. Pitman (1978) estimated sea level due to this change in ridge volume for the period from 85 Ma to 15 Ma (Fig. 17). He estimated that the rate of change of sea level has varied throughout the Cenozoic, resulting in transgressions in the Eocene and Miocene when the rate of sea level lowering decreased. Regressions occurred in the intervening intervals when sea level fell at a faster rate (Fig. 17). Superimposed upon this general overall decrease in sea level, and the associated transgressions and regressions, are the dramatic fluctuations in absolute sea level recorded by Vail *et al.* (1977). figure 17 records the position of the shoreline during the Cenozoic resulting from changes in mid-oceanic ridge volume. If sea level changes as a result of some other mechanism (producing a rapid change) then the position of this shoreline will vary accordingly. For example, in the Middle Oligocene (29 Ma) Vail *et al.* (1977) recorded the largest lowering of sea level in the Cenozoic (– 250 m). The apparent magnitude of the fall may be exaggerated by the fact that sea level had already regressed a considerable distance towards the shelf edge by 29 Ma (Fig. 17). Hence, a rapid fall of sea level occurring at 29 Ma would tend to be amplified in the geologic record. Conversely, during transgressive periods in the Middle Eocene and Middle Miocene, when the shoreline was further from the shelf edge, the apparent magnitude of a rapid sea-level drop would be decreased.

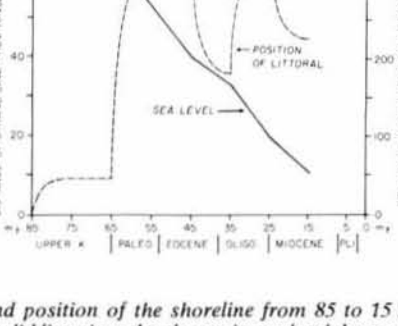


Figure 17

Sea level and position of the shoreline from 85 to 15 Ma (Pitman, 1979). The solid line gives the change in sea level due to the change in mid-oceanic ridge volume. The dashed line gives the position of the shoreline with respect to the hingeline of the continental margin as a function of the rate of sea-level change (Pitman, 1979).

Therefore, we suggest that the magnitude of a sea-level drop may be exaggerated by the position of the shoreline (distance from margin hingeline) at the time of initiation of rapid sea-level falls. The apparent magnitude of sea-level falls will be reduced when the shoreline is farthest from the shelf edge (transgression). When the shoreline is closest to the shelf edge (regression) the apparent magnitude of a sea-level fall will be exaggerated.

Also, during times of lowstand of sea level (below the shelf edge) the shelf coastal plain platform may become an erosional surface. Some isostatic rebound of the continental margin may occur as sediments and water are removed. Any such rebound will also tend to exaggerate the rate of sea-level fall and the extent of sea-level drop.

In summary, to gain a better understanding of the possible causes of sea-level changes, there are two problems which must be addressed. The first is to obtain better estimates of the overall fall of sea level during the Cenozoic. This will require refinements in the magnetic anomaly time scale and hence spreading rate estimates and a better understanding of past plate geometry. Secondly, the existence of large, rapid falls of sea level must be further documented. This may be accomplished by the use of geohistory curves (Van Hinte, 1978) and correction of these curves for sediment compaction and loading (Steckler, Watts, 1978) in combination with seismic reflection techniques.

Global ice fluctuations are the only known cause of such rapid sea-level fluctuations. However, large volumes of continental ice appear to be restricted to the Late Cenozoic, since the Middle Miocene (14 Ma to present, Shackleton, Kennett, 1975 ; Savin *et al.*, 1975). Since rapid sea-level fluctuations are recorded by Vail *et al.* (1977) through the Cenozoic it is appropriate to examine the Cenozoic polar glacial ice history and its possible relation to the global sea-level curve of Vail *et al.* (1977).

POLAR GLACIAL EVOLUTION AND GLOBAL SEA LEVEL CHANGES

Numerous rapid falls of sea level are a prominent feature of the global sea-level record during the Cenozoic and Mesozoic (Vail *et al.*, 1977). The existence of these rapid falls of more than 100 m in a period of less than a million years is controversial. The rapidity of the falls suggests a glacial origin since the formation and destruction of large volumes of continental ice is the only known mechanism which can change sea level at the required rate (up to 1,000 cm/1,000 yrs). However, present thinking favors the hypothesis that large continental ice masses are a Late Cenozoic feature (post Middle Miocene).

Kennett (1977) comprehensively reviewed the evolution of the Antarctic icecap and the circum-Antarctic Ocean. The present scenario for the formation of the Antarctic icecap is based on a number of factors including the changing position of the continents around Antarctica during the Cenozoic, changes in deep-sea sediment facies, the presence or absence of deep-sea unconformities, oxygen isotopic evidence, the distribution of ice-rafted debris, terrestrial glacial evidence and biogeographic information.

Almost all interpretations of the glacial history of Antarctica are based on analyses of the deep-sea sedimentary record since terrestrial glacial evidence for the Early and Middle Cenozoic is fragmentary. However, good exposures have been reported in the latest Cenozoic and have provided important information on more recent (latest Miocene to Recent) Antarctic glacial evolution (Mayewski, 1975).

The most important factor in controlling the development of the Antarctic icecap was the northward movement of the Gondwanaland continents away from their polar position. As a result, the circulation pattern of the Southern Ocean progressively changed to one of unimpeded circum-Antarctic flow. At the Eocene-Oligocene boundary, when a shallow water passage south of Australia opened, surface water temperatures dropped sufficiently to allow sea-ice development close to Antarctica. The Antarctic continent became even more thermally isolated sometime between 30 Ma and 22 Ma as the Drake Passage opened sufficiently to allow completion of the circum-Antarctic deep water circulation system. At approximately 15 Ma a large continental icecap formed on Antarctica during a time of apparent global warming.

Oxygen isotopic analysis of biogenic calcium carbonate provides the most direct method of estimating paleotemperatures. However, an increase in the $^{18}O/^{16}O$ ratio of biogenic carbonate during a glacial period reflects both an increase in the $^{18}O/^{16}O$ ratio of seawater and enrichment of $\delta^{18}O$ in the carbonate due to a temperature decrease. The isotopic signal obtained from biogenic carbonate is therefore not easily interpreted since the isotopic composition of continental ice masses is not precisely known (Savin, 1977).

Shackleton and Kennett (1975) interpreted the Cenozoic oxygen isotopic data from Subantarctic deep-sea sequences to demonstrate that the Antarctic icecap began to form rapidly in the early Middle Miocene. However, it is very difficult to differentiate the magnitude of ice volume or temperature effect during the dramatic enrichment in $\delta^{18}O$ at approximately 14 Ma (Savin *et al.*, 1975). It is generally assumed that a large proportion of the oxygen isotopic signal after the Middle Miocene is due to fluctuations in the volume of ice sheets in polar regions. Pre-Middle Miocene oxygen isotopic data are presumed to provide a direct measure of oceanic paleotemperatures.

Using these assumptions, Shackleton and Kennett (1975) suggested that waters around Antarctica approached the freezing point at the base of the Oligocene. Keigwin (1980) has shown that the dramatic isotopic enrichment recorded by Shackleton and Kennett (1975) at the Eocene-Oligocene boundary in DSDP Site 277 is due to a rapid decrease in temperature of 3-4 °C as originally suggested by Shackleton and Kennett (1975). Seasonal sea-ice formation may have begun at this time, but it was not until about 26 Ma that significant confirmed ice-rafted debris appeared (Hays, Frakes *et al.*, 1975), signaling the first appearance of calved glaciers at sea level. Sometime between these two dates, Antarctic coastal waters probably reached freezing point and valley glaciation expanded and reached the coast. Oxygen isotopic evidence, according to present thinking does not indicate significant ice volume on Antarctica before the Miocene but some evidence, such as ice-rafted debris suggests that inland glaciation did begin during the Oligocene or perhaps earlier (Hays, Frakes, 1975). Other independent evidence, including the distribution of biogenic siliceous and calcareous sediment, biogeographic evidence on land and in the sea indicates that the Oligocene was a time of expanding glaciation on Antarctica and of cooling coastal waters, especially during the Middle to Late Oligocene (Kennett, 1977). *Nothofagus*, (Southern Beech) was present around Antarctica during the Oligocene but gradually began to die out during the Late Oligocene to Early Miocene (Kemp, Barrett, 1975) probably in response to increased glaciation. A key factor in polar glacial history was the development of circum-Antarctic circulation which resulted from the movement of Australia northward at about 38 Ma and the opening of the Drake Passage sometime during the Late Oligocene.

Similar types of evidence from the Early Cenozoic suggest that Antarctic coastal water temperatures were relatively warm (temperate) and that if any glaciers were present they must have only been present as valley glaciers at high elevations. One of the major hindrances to determining the time of initiation of Antarctic glaciation and its subsequent history is the critical lack of deep-drilled marine sequences close to the continent.

In summary, it appears unlikely that large volumes of continental ice existed prior to the Middle Miocene or the Oligocene at the earliest. Even if the isotopic composition of continental ice has been underestimated, independent evidence, although fragmentary, suggests that oceanic and terrestrial conditions of the Antarctic region preclude the existence of an ice sheet on Antarctica. We note that despite the inferred Middle Miocene buildup of Antarctic ice, the character of individual sea-level cycles through the Late Cenozoic remains similar to that of the Middle and Early Cenozoic. However, there is a definite change in the character of the supercycles in the Late Cenozoic. Supercycle (Te) and the latest Miocene-Pliocene supercycle (Tf) exhibit mid-supercycle peaks of sea level distinguishing them from the earlier supercycles which terminate at times of highest sea level. This change in character may represent glacio-eustatic effects superimposed upon a longer term eustatic driving mechanism. Also, this longer term eustatic driving mechanism has probably remained constant during the Mesozoic and Cenozoic, since the character of the individual sea-level cycles has not changed during this period (Vail *et al.*, 1977). If continental ice sheets are a Late Cenozoic feature only, then an alternate mechanism must be found to produce the rapid falls of sea level reported by Vail *et al.* (1977).

CONCLUSIONS

We have examined relationships between the Cenozoic global sea-level curve of Vail *et al.* (1977), the Australian and New Zealand continental margin shallow-marine sedimentary record, the deep-sea sedimentary record and the history of polar glaciation. We have found that:

1) Cenozoic depositional sequences in the Australian Cenozoic marginal marine sequences appear to be primarily controlled by eustatic sea-level changes and to a lesser extent by local tectonic and climatic conditions.

2) The majority of the New Zealand stage boundaries between the end of the Paleocene (53 Ma) and the end of the Miocene (5 Ma) appear to be isochronous with eustatic sea-level cycle boundaries. The majority of the New Zealand stages are natural stratigraphic units each corresponding to one sedimentary cycle. Thus, we are able to correlate the New Zealand stages with the tropical planktonic foraminiferal zonation schemes of Bolli (1966), Blow (1969), and Berggren (1969).

3) Global sea-level changes during the Cenozoic played a critical role in controlling deep-sea sedimentation. Sea-level high-stands (transgressions) trap terrigenous sediment and organic carbon on continental shelves and restrict the supply of dissolved material to the deep sea resulting in a hiatus maxima in the deep sea. Conversely, sea-level lowstands are associated with hiatus minima in the deep sea. Other controls of deep-sea sedimentation include local tectonic and climatic conditions on continental margins, bottom water advection, local upwelling and corrosiveness of oceanic waters to biogenic skeletal material.

4) It is unlikely that fluctuations in continental ice volume control global sea level through the entire Cenozoic. Large volumes of polar ice appear restricted to the Late Cenozoic

(post Middle Miocene, 14 Ma). Thus, some other mechanism must be found to account for the apparent rapidity and magnitude of the sea-level falls recorded by Vail *et al.* (1977) throughout the Mesozoic and Cenozoic.

5) As Suess (1888) suggested, sedimentary cycles or depositional sequences bounded by unconformities may be used "to distinguish sedimentary formations in all parts of the world". The use of unconformities as a stratigraphic tool, in the absence of multi-channel seismic profiles and well data, is restricted to uplifted shallow-marine sequences but may well prove to be a vital tool in linking the classical land-based sections with those in the deep sea.

6) The detailed Cenozoic sea-level history of Vail *et al.* (1977) has provided a fundamental framework with which to correlate sedimentary sequences around the world and to demonstrate the temporal relationships between continental margin sedimentation and deep-sea sedimentation. However, no satisfactory mechanism has been found to explain the rapidity of the sea-level changes recorded by Vail *et al.* (1977).

7) Depositional sequences are extremely useful for global correlation and thus should be formally recognized in the International Code of Stratigraphic Nomenclature.

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