

ATOLL RESEARCH BULLETIN

NO. 322

**TECTONIC AND ENVIRONMENTAL HISTORIES IN THE PITCAIRN GROUP,
PALAEOGENE TO PRESENT: RECONSTRUCTIONS AND SPECULATIONS**

BY

T. SPENCER

**ISSUED BY
NATIONAL MUSEUM OF NATURAL HISTORY
SMITHSONIAN INSTITUTION
WASHINGTON, D.C., U.S.A.
October 1989**

TECTONIC AND ENVIRONMENTAL HISTORIES IN THE PITCAIRN GROUP, PALAEOGENE TO PRESENT: RECONSTRUCTIONS AND SPECULATIONS

BY

T. SPENCER¹

ABSTRACT

Interpretation of SEASAT geoid anomaly data and improved seafloor mapping of the south-central Pacific suggest a complex tectonic history for the islands of the Pitcairn group. While Oeno atoll formed at ~ 16m.y.BP at a 'hotspot' now south of the Easter micro-plate, subsequent progressive island development at Henderson (13m.y.), Ducie (8m.y.) and Crough seamount (4m.y.) resulted from the lateral leakage of magma from the Oeno lineation along an old fracture zone, itself originating during the Tertiary reorientation of the Pacific plate. At all four islands cessation of volcanism was followed by subsidence and the development of a carbonate cap. By comparison, Pitcairn has been the product of recent (<1m.y.) volcanic activity along an independent, subparallel hotspot lineation. Nevertheless, this activity has interacted with the older island chain by transforming Henderson Island, through the process of lithospheric flexure, into an uplifted atoll with ~ 30m of relief.

These tectonic processes have been accompanied by changes in sea level and oceanographic conditions. As the Holocene record shows, the deciphering of the sea level record at these islands is difficult; sea level change has been a response not only to glacio-eustatic processes but also to a range of isostatic, and possibly geoidal, effects. Although the Pitcairn group at ~24°S occupies a marginal position for reef growth and development, reconstructions of palaeoceanographic conditions for the Tertiary and Quaternary suggest that the tropical water masses were largely unaffected by either changes in ocean circulation systems or climatic cooling and that water temperatures in the past have been very similar to those experienced at the present time.

INTRODUCTION

Until recently the tectonic and environmental history of the south-central Pacific has been poorly understood. Previous reconstructions of ocean basin and island histories have had to rely upon the relative paucity of information supplied from infrequent and low density bathymetric traverses of research vessels. However, within the last decade the application of new remote-sensing technologies, improved mapping of the sea floor and the transfer of deep-drilling techniques refined in more accessible oceans has vastly expanded the volume of information available from even the remotest parts of the south Pacific.

¹ Department of Geography, University of Manchester, Manchester M13 9PL, UK; present address: Department of Geography, University of Cambridge, Cambridge CB2 3EN.

Thus, for example, critical phases in the geological history of the oceanic lithosphere in this region took place in the early Miocene, a time period for which sea-floor magnetic anomalies are few and broadly separated in age, yielding more conjectural estimates for sea-floor spreading rates than at subsequent time periods. However, following the operation of the altimeter satellites GEOS-3 and more particularly SEASAT, variation in the height of the sea surface - the marine geoid - is now known throughout the south Pacific (e.g. Sandwell 1984) to a height of accuracy of 10-30 cm and a horizontal resolution of 10-50 km. Geoidal signals have a high level of correlation with sea-floor topography and have thus been used to discover previously undetected bathymetric features (e.g. Sabers et al. 1988), precise seamount geometry then being determined by multi-beam sonar mapping (e.g. Pontaise et al. 1986).

As the following syntheses demonstrates, degree of detail of this kind has revolutionised the level of explanation of regional geodynamics and environmental change in the south-central Pacific. As a result, some preliminary reconstructions of tectonic and environmental histories for the Pitcairn group can be attempted.

PLATE-TECTONIC EVOLUTION OF THE SOUTH-CENTRAL PACIFIC

TERTIARY RE-ORIENTATION OF THE PACIFIC PLATE.

At the beginning of the Palaeogene (65 million years B.P.) the prototypes for all the major oceans were already in existence. The Pacific was a multi-plate ocean, separated by subduction zone margins from the Asian and, in all probability, the Australian plate and bounded to the east by a complex series of mid-ocean ridges and triple junctions from the Farallon, Kula and Phoenix plates (Williams 1986; Figure 1). Although the Pacific Ocean was gradually reduced in size during the Palaeogene, with rates of subduction exceeding those of seafloor spreading, the Pacific plate itself increased in size at the expense of the plates on its western margin. The re-orientation of the plate, from a NNW to WNW spreading direction at ~42 million years B.P., preserved in the hot-spot traces of intra-plate islands and seamounts, most notably by the 'bend' in the Hawaiian Islands - Emperor Seamounts chain, is well known. Of equal significance in the South Pacific, however, was the collision of the Pacific-Kula-Farallon boundary with the Farallon-Americas trench at ~26 million years B.P. This event stopped all seafloor spreading and subduction in this region and initiated direct coupling between the Pacific and Americas plate. This fusion created a major, progressive re-orientation of seafloor spreading patterns and plate geometries as follows: i) ~20 million years B.P.: clockwise rotation of the southern portion of the Pacific-Farallon ridge and the development of the Galapagos rift; ii) ~10 million years B.P.: break-up of ridge south of Baja California and ridge 'jumps' and iii) ~5 million years B.P.: development, from the north, of the Gorda-Juan de Fuca-San Andreas-Gulf of California spreading system with eastward ridge jump below Baja California and westward shift of the Galapagos triple junction (Herron 1972, Herron and Tucholke 1976, Handschumacher 1975).

These adjustments were echoed in progressive plate re-orientations between 15-30° which have been fully documented by Okal and Cazenave (1985) using a full SEASAT dataset. At 40 million years B.P., spreading was taking place along the Mendoza and Roggeveen ridges (terminology : Mammerickx et al. 1980), offset 500 km by the transform fault of the Austral Fracture Zone (Figure 2a). Between 36 and 25 million years B.P. the Mendoza ridge propagated south, causing ridge re-orientation and the development of two fracture zones : FZ1, a re-orientation and re-located version of the Austral Fracture Zone and FZ2, dividing the northern and southern sections of the Roggeveen ridge (Figure 2b); Okal and

Cazenave 1985, Sailor and Okal 1983, Okal and Bergeal 1983). Both these fracture zones are orientated N95°E, intermediate between the strike of the ancient Farallon ridge (N70°E) and the orientation of spreading of the present East Pacific rise (N110°E), are 400-500 km in length and may be associated with contemporary seismic activity (Figure 3; Okal and Cazenave 1985, Okal et al. 1980).

The southern Roggeveen and the Mendoza ridges subsequently 'jumped' westwards at 20 and 18 million years B.P. respectively, completing the re-alignment of the Pacific plate boundary with a straightening of the spreading system, and perhaps providing the western boundaries of the present Easter micro-plate (Figure 2c - f); Hey et al. 1985).

STRUCTURAL TRENDS AND ISLAND AGES : SOUTHERN TUAMOTU ARCHIPELAGO AND PITCAIRN ISLAND.

The islands of the Pitcairn Group- Ducie and Oeno atolls, Henderson Island and Pitcairn Island itself - lie between 23.9°S - 24.7°S and 124.7°W - 130.7°W. The islands are both widely spaced and isolated from their nearest neighbouring groups : Ducie atoll is 1000 km west of the Easter micro-plate; Oeno atoll is 450 km east of the Minerve reefs and the Gambier Islands (Figure 3). Henderson Island rises from water depths of ~3,500 m; similarly, Pitcairn Island has been constructed from a sea floor at least 2,000 m and perhaps 3,500 m below the ocean surface. By comparison, Oeno atoll appears to rise from the southern side of a broad plateau at 1,600 m and Ducie atoll is probably not a simple feature (Mammerickx et al. 1975) but the surface expression of a complex collection of seamounts (Canadian Hydrographic Service 1982). Further to the east, clearly present in the SEASAT geoid but omitted from bathymetric charts, is a major structure topped by two seamounts reaching 1,000 m below present sea level at 25°S, 122.2°W and 24.8°S, 121.7°W; Okal (1984) has proposed that this feature be known as Crough Seamount. Geoidal signatures have revealed further seamounts around 25.6°S, 121.2°W and 26.2°S, 121.8°W (Okal and Cazenave 1985). Finally, the presence of several small seamounts near the East Pacific Rise is indicated by a cluster of geoidal anomalies. The regional synthesis of these recently-discovered submerged features indicates a much more complex and extensive island group lineation than has been apparent up to now from the disposition of islands above sea level. The alignment and spacing of the islands is strongly suggestive of an origin related to a relatively-fixed melting anomaly, or 'hotspot', to the east of the islands and seamounts.

Island chain hotspot traces should show a progressive increase in island age away from the hotspot itself in the direction of plate motion. However, given island morphologies in the Pitcairn Group, the only established dates for island genesis are from Pitcairn Island where potassium-argon dating of exposed volcanics has identified two phases of volcanism, at 0.46-0.63 and 0.76-0.93 million years B.P. The latter period probably represents the main phase of island construction (Duncan et al. 1974). Basaltic lavas which form the islands of Mangareva, Aukena and Makapu, Gambier Islands, cooled between 5.2 and 7.2 million years B.P. (Brousse et al. 1972) and recovered basalts from Fangataufa atoll and Mururoa atoll have been dated to 7.1-9.1 million years B.P. and 6.5 and 8.4 million years respectively (Brousse 1985), Duncan and Claque 1985). Volcanic migration rates have been calculated at $12.7 \pm 15.5 \text{ cm yr}^{-1}$ (Duncan and Claque 1985) for the Pitcairn-Gambier sequence and at $10.7\text{-}11.0 \text{ cm yr}^{-1}$ (Brousse 1985) when extended to Mururoa atoll. These propagation rates are comparable to the values calculated for other S.

Pacific island chains (e.g. Austral-Cook Islands: $10.7 \pm 1.6 \text{ cm yr}^{-1}$; Marquesas: 10.4 ± 1.8 ; Society Islands: 10.9 ± 1.0 ; Duncan and Claque 1985).

Unfortunately, however, Pitcairn Island's evolution throws little light on the origins of the other islands of the Pitcairn Group. Although almost certainly underlain by volcanics, none of the carbonate caps of Ducie, Oeno or Henderson have been deep-drilled to reach basement basalts. In the absence of such direct data, current reconstruction must be based upon the interpretation of the marine geoid.

SEASAT DATA AND ISLAND ORIGINS : OENO, HENDERSON, DUCIE AND CROUGH

Methodological comment

Fundamental to the concept of plate tectonics is the notion of a strong, rigid lithosphere overlying a weak, fluid asthenosphere. Lithospheric rigidity can be determined by studying the response of oceanic lithosphere to surface loads. By comparing calculated profiles of flexure, computed from known or assumed load shapes, with observed bathymetry, patterns of island uplift and subsidence, seismic refraction results and gravity and/or geoid anomalies, it appears that the most useful first-order model of oceanic plate behaviour is that in which the lithosphere is modelled as a thin elastic plate overlying weak substratum (e.g. McNutt and Menard 1978, Watts 1978). As oceanic lithosphere cools with age (e.g. Parsons and Sclater 1977) it thickens and becomes less responsive to surface loads. Not surprisingly, therefore, summaries of flexure studies (Figure 4) show that the elastic thickness of the lithosphere, T_e , increases with the age of the plate at the time of loading, estimated by subtracting the age of the load from the age of the seafloor (Watts et al. 1980, Watts and Ribe 1984). Thus, for example, seamounts formed on young lithospheres are associated with low values of T_e (e.g. 5 km elastic thickness of age = 5 million years) whereas islands produced on old lithosphere are characterised by high values of T_e (e.g. 25 km elastic thickness of age = 80 million years). These differences are preserved through time and thus indicate whether islands were formed either at mid-ocean spreading ridges on thin oceanic plate or at mid-plate locations on much thicker lithosphere. This distinction is reflected in geoidal signature: on-ridge genesis is indicated by small amplitude (0.4-0.5m per km seamount height), short-wavelength geoidal anomalies whereas off-ridge origin is shown by large amplitude (1.4-1.5 m/km), large-wavelength perturbations (Watts and Ribe 1984; Figure 5).

Geodynamics and island ages in the Pitcairn Group

SEASAT geoid anomaly data in the south-central Pacific shows that the broad plateau of the Tuamotu Archipelago has no strong geoidal signature; furthermore, thick (27 km) oceanic lithosphere has been reported beneath the island of Rangiroa (Talandier, in Okal and Cazenave 1985). Both these lines of evidence suggest an on-ridge origin for the plateau and its islands. In the southern Tuamotus, the atolls of Tatakato, Pukaruha and probable Reao conform to this explanation and Okal and Cazenave (1985) have suggested that these islands formed at a hotspot near to, and then on, the junction of the Austral Fracture Zone and the Mendoza mid-ocean ridge (Figure 2a). It seems likely that the meeting of the hotspot and the Austral Fracture Zone triggered the southerly propagation of the Mendoza ridge (see above) and the subsequent deactivation of the Austral Fracture Zone and concomitant initiation of the FZ1 fracture zone (Figure 2b). When the spreading system jumped westwards, the hotspot became an off-ridge melting anomaly on the Farallon plate,

generating the seamount sequence on the Sal-y-Gomez ridge (Figure 2c) -f) and see Duncan and Hargraves 1984, Schilling et al. 1985). By comparison, the more southerly Tuamotu atolls, from Hao south-east to the Acteon group and Marutea, show strong geoidal signals, indicating that they must have formed off-ridge. Importantly, SEASAT data for Oeno, Ducie and Crough seamount also suggest an off-ridge origin. Unfortunately, no comparable data is available for Henderson Island but given its location it seems reasonable to assume an off-ridge origin for Henderson as well. The extension of the Hao-Marutea trace passes through the location of Oeno and predicts the present position of the hotspot to be ~300 km south of the south-western boundary of the Easter micro-plate (Figure 3). Henderson Island, Ducie and Crough seamount, however, do not fall on this alignment but form a linear chain at a 15° angle to it (Figure 3). Okal and Cazenave (1985) have suggested that this deviation has resulted from the interaction of the hotspot with the line of lithospheric weakness represented by the old fracture zone FZ2, with, following the theory of Morgan (1978), lateral leakage from the hotspot leading to the successive construction of Henderson, Ducie and Crough.

If such speculations on island origins are correct, then what ages can be assigned to the islands within the Pitcairn group, aside from Pitcairn Island itself? Okal and Cazenave (1985) give ocean floor age estimates of 10 million years for Crough Seamount, 14 million years for Ducie, 19 million years for Henderson and 27 million years for Oeno. In the vicinity of Pitcairn Island, Duncan et al. (1974) have suggested that the plate has an age of 30 million years. Using the methodology of Cazenave and Dominh (1984), where the comparison of observed and theoretical geoid heights is used to define a best fit of lithospheric flexural rigidity (D) and effective elastic thickness (T_e), Okal and Cazenave (1985) have calculated the age of the plate at the time of loading, or flexural age, as 5-7 million years. This would suggest the following island ages (± 1 million years):- Crough Seamount : 4 million years; Ducie : 8 million years; Henderson : 13 million years; and Oeno : 16 million years. These dates, however, can only be seen as first approximations as the flexural age is not a true measure of the age of the plate at time of loading because of the likelihood of plate re-heating at the time of island emplacement (for theory : Detrick and Crough 1978, McNutt 1984). Interestingly, in the Central Pacific a range of geophysical anomalies indicate low elastic thickness over the region between -10°S to -30°S and from 120° to -160°W (Cochran 1986, McNutt and Fischer 1986, Calmant and Cazenave 1987). McNutt and Menard (1982) and Calmant and Cazenave (1986) have interpreted these low thicknesses as indicative of thermal rejuvenation.

Cessation of volcanic activity and island subsidence subsequently led to the development of carbonate caps, of unknown thickness, on the islands of Oeno, Henderson and Ducie (and Crough?). With the later, and independent, development of the Duke of Gloucester Islands - Pitcairn Island hotspot lineation, Henderson was affected by lithospheric flexure processes. On the relatively thin and deformable lithosphere of the South Pacific, the emplacement of relatively young (<2 million years) volcanoes, as a result of 'hotspot' activity, has produced a near-volcano moat and a peripheral bulge, or arch, at some distance from each load. The coincidence of sea-level coral reefs with this radius has resulted in the formation of raised reef islands (McNutt and Menard 1978). Flexural moats have been defined around some mid-plate volcanoes (e.g. Hawaii: Ten Brink and Watts 1985, Watts et al. 1985; Marquesas: Fischer et al. 1986) and reef limestone uplift, apparently associated with flexure, demonstrated from the Society Islands (Pirazzoli 1983), the N.W. Tuamotu Archipelago (Lambeck 1981a, Pirazzoli and Montaggioni 1985) and the southern Cook Islands (Lambeck 1981b, Stoddart et al. 1985, Spencer et al. 1987, Calmant and Cazenave 1986). In the Pitcairn group lithospheric flexure under the weight of the Pitcairn Island has resulted in - 30 m of uplift at Henderson. The raised reef topography of

Henderson, and estimates of the rate of flexure-controlled uplift, are considered in more detail by Spencer and Paulay (this volume).

In summary, therefore, the final patterning of the islands of the Pitcairn group appear to have been derived from three different tectonic settings: old hotspot (Oeno); leakage from old hotspot (Henderson-Crough); and young hotspot (Pitcairn Island), with further tectonic activity at Henderson. Thereafter, the development of island morphologies within the group has been a reaction to mid-latitude environmental change; this is considered in more detail below.

Geodynamics and island age: complications and speculations

Although the scenario outlined in some detail above is an attractive one, recent SEASAT studies in the area to the south-west of the Cook-Austral Islands archipelago have revealed, importantly, the presence of further fracture/fault zones with the same alignment (N95°E) as FZ1 and FZ2 but extending linearly over much greater distances than Okal and Cazenave's (1985) fractures, in excess of 1000 km (Diament and Baudry 1987). Rather than regarding such lineations as the short-lived (~8million years; Okal and Cazenave 1985) product of mid-ocean ridge re-orientation, it has been suggested that these features indicate far more fundamental intra-plate deformations, being either the result of differential movement between the northern and southern sections of the Pacific plate or the consequence of changes in the absolute motion of the plate as a whole (Diament and Baudry 1987).

Such ideas have resurrected the notion of a south Pacific 'hotline', linking melting anomalies from the Tonga trench to the Nazca plate (Bonatti and Harrison 1976, Bonatti et al. 1976, Turner and Jarrard 1982). This hypothesis has found support from the detection of gravity undulations, of 150-200 km wavelength and trending WNW in the direction of plate motion, between the East Pacific Rise and French Polynesia by Haxby and Weissel (1986). They suggest that such features result from small-scale thermal instabilities beneath young lithosphere being organised into longitudinal rolls by the shear from fast-moving plates. Interestingly, such patterns are predicted by some convection models (e.g. Richter 1973). SEASAT altimeter data further suggests that such small-scale convection develops within 5 to 10 million years of the initiation of plate cooling (Haxby and Weissel 1986); thereafter this pattern is 'frozen in' to the lithosphere (Buck 1985, Buck and Parmentier 1986).

Finally, however, it should be noted that two further models have been proposed to explain these undulations: compressive buckling (McAdoo and Sandwell 1985) and tensional cracking (Winterer and Sandwell 1987) of the lithosphere. Clearly, careful bathymetric work will be required to discriminate between these different explanations before a more comprehensive 'hot-line' hypothesis can be formulated.

ENVIRONMENTAL HISTORY OF THE SOUTH-CENTRAL PACIFIC

The growing inventory of high-quality deep-sea cores from the Deep Sea Drilling Project (DSDP) and the determination of fine resolution down-core environmental records from core lithology, stable isotopes and microfloral and microfaunal assemblages now permits the reconstruction of palaeoclimatic and palaeoceanographic conditions, for precise time slices, as far back as the beginning of the Palaeogene.

Clearly, however, such reconstructions are dependent upon the spatial coverage of deep-sea core sites. Unfortunately, the density of sites in the Pacific Ocean is poor by comparison with the Atlantic Ocean and biased towards the northern hemisphere and the eastern equatorial region. Thus, only broad inferences can be made, as yet, as to former circulation patterns and sea surface palaeotemperatures in the south-central Pacific.

Methodological comment

A key tool in environmental reconstruction of ancient oceans is oxygen isotope stratigraphy. $^{18}\text{O}/^{16}\text{O}$ ratios in foraminifera from deep-sea sediments have been used as indicators of past climates since it was demonstrated in the 1950s that ^{18}O enrichment ($\delta^{18}\text{O}$) in foraminifera varies with ^{18}O in the water from which their carbonate skeletons have been precipitated but differs from the water value by an amount determined by temperature (e.g. Epstein 1953, Emiliani 1955; but see also Shackleton 1984 for problems of temperature calibration and potential effects of within-sediment diagenesis of skeletal calcite). However, the deep-sea palaeotemperature record is complicated by historical fluctuations in ocean isotopic composition. It is generally agreed that there have been changes in isotopic composition of the order of 1.0-1.6 (180 per mil to PDB standard) resulting from the repeated growth and decay of isotopically light ice sheets on the Northern Hemisphere continents (Shackleton 1967, 1984) but the interpretation of the Tertiary record has proved more problematical. One hypothesis (Matthews and Poore 1980) suggests that there have been no significant changes in low latitude sea surface temperatures in the Cenozoic and, therefore, that the fluctuations in the oxygen isotope record only reflect changes in global ice volumes. This hypothesis, however, challenges the widely-held view (e.g. Shackleton and Kennett 1975, Savin 1977, Woodruff et al. 1981) that Antarctica was essentially ice-free until the middle Miocene and, therefore, that the Palaeogene isotopic record is one of temperature change in the deep ocean. Recently more accurate oxygen isotope measurements (Shackleton 1986), whilst providing general support for this second model, have shown the interpretation to be too simplistic, and it now seems likely that ice was present in Antarctica in the early (Shackleton et al. 1984b), middle (Keigwin and Keller 1984) and latest (Miller and Fairbanks 1985) Oligocene. These arguments need to be borne in mind when interpreting the oxygen isotope record of the past 70 million years.

TERTIARY PALAEOGEOGRAPHY AND PALAEOCEANOGRAPHY

Shackleton's (1984) compilation of oxygen isotope data (Figure 6) shows that the early Cenozoic was characterised by high temperatures at both low- and mid-latitudes and thus relatively small equator-to-pole and surface-to-bottom temperature gradients. Dramatic cooling of both mid-latitude and deep-ocean waters occurred at the Eocene-Oligocene boundary (Figure 6; Keigwin 1980) and there was a further divergence, of mid-latitude and deep-ocean water temperatures, in the middle Miocene (Savin et al. 1981, Shackleton and Kennett 1975, Woodruff et al. 1981). Evidence for increasingly vigorous atmospheric-oceanic circulation, at both 15-16 myr B.P. and 9-5 myr B.P., associated with climatic deterioration, is provided by the appearance of diatomites in Pacific rim sedimentary sequences (Ingle 1981), increasing biogenic silica accumulation (from 16 myr B.P., peak at 8 myr B.P.: Leinen 1997) and rising calcium carbonate supply rates (peaking 14-15 myr B.P.: Van Andel et al. 1975) in the equatorial Pacific. Changes towards large grain sizes in the particle-size distribution of aeolian dust from 11.8 myr B.P. imply a significant intensification of the southern hemisphere tradewinds (Rea and Bloomstine 1986); this

mirrors the better-known North Pacific record (e.g. Rea and Janecek 1982, Rea et al. 1985).

It has been suggested that these events in the Pacific Ocean were indirectly driven by palaeogeographic changes, themselves the result of the re-arrangement of plate boundaries; thus an attractive scenario argues that the Oligocene development of the Circum-Antarctic Current, as Australia and, later, South America, became detached from Antarctica, progressively isolated the Antarctic from lower latitude influences and resulted in cooler, polar temperatures, increased presence of sea-ice, cooler bottom water temperatures and, ultimately, the development of a major continental ice sheet on East Antarctica (Kennett 1977, 1982). Allied to the establishment of this high-latitude, circum-global circulation was the modification and ultimate loss of the circum-equatorial Tethys seaway, firstly by the mid- to early late-Miocene closure of the Indo-Pacific passage in the Indonesian region as a result of the continued northward migration of Australia/New Guinea (Edwards 1975, Hamilton 1979), and secondly by the late Miocene constriction - Pliocene closure of the Atlantic-Pacific connection through the isthmus of Panama (Keigwin 1978).

These palaeogeographic changes both steepened pole-to-equator temperature gradients and altered the pattern of atmospheric and oceanic circulation towards their present arrangement. In the equatorial region these changes enhanced surface productivity of ocean waters by diverting large volumes of water into westerly boundary currents which were then returned as an intensified surface circulation and by the development of an equatorial undercurrent (Figure 7). As a result, east v. west Pacific biogeographic differences were reduced and replaced by a latitudinal provincialism (Kennett et al. 1985). However, and importantly, these changes took place around the largely unaffected subtropical gyres.

QUATERNARY ENVIRONMENTS: SEA LEVEL CHANGE IN THE PLEISTOCENE

The oxygen isotope record from deep-sea cores (Figure 8) indicates strong climatic fluctuations from ~ 3.2 myr B.P. (Shackleton and Opdyke 1977, Shackleton et al. 1984a). There have been 10 completed glacial-interglacial cycles in the last one million years, with 8 cycles, at approximately 100,000 yr intervals, since 0.73 myr B.P. (Shackleton and Opdyke 1973). Each cycle shows a characteristically 'saw-toothed' pattern of slow, progressive ice build-up and rapid de-glaciation (Broecker and Van Donk 1970; Figure 8).

The mass and melting history of continental ice sheets can also be derived from radiometrically-dated coral reef sequences preserved by uplift on tectonically active coasts. The difference between the altitude of a reef of known age and the present sea level provides a precise measure of the past sea level if the tectonic component can be estimated and subtracted (e.g. Haiti: Dodge et al. 1983; Barbados: Matthews 1973). Using these principles, a particular fine record of sea level change has been constructed from the Huon Peninsula, New Guinea (Chappell 1974), Bloom et al. 1974, Chappell 1983) where coral reef terraces have been preserved along a rapidly rising ($0.9-3.5 \text{ m kyr}^{-1}$) coastline. The sea-level curve from this locality has been recently refined, following comparison with the deep-sea core oxygen isotope record (Chappell and Shackleton 1986; Figure 9). The curve shows, for the last glacial-interglacial cycle fast rising sea levels (up to 8 m kyr^{-1}) during the major post-glacial transgressions, culminating in interglacial high sea level stands between 118-138,000 yr B.P. (reef complex VII) and from 8,200 yr B.P. (reef I). The Last Interglacial sealevel on the Huon Peninsula (124,000 yr B.P.; reef VIIa) is assumed to

have reached ~ 6.0 m above present sea level, as elsewhere around the globe (Moore 1982). Estimates for low sea levels prior to these transgressions, determined by maximum ice volumes, have been set at -130 m, less than the levels of -135 to -165 m suggested by evidence from northern Australia but in excess of the values of -90 to 110 m reported from the eastern seaboard and Gulf of Mexico coasts of the USA (Chappell and Shackleton 1986). Such differences are to be expected as they relate not only to the record of sea level change but also to the behaviour of the continental margins. Slower rising sea levels (up to 2.5 m kyr^{-1}) were recorded by Huon fringing reef development at 100,000 (reef VIa), 81,000 (Va), 59,000 (IVa), 45-40,000 (IIIa-IIIb) and 28,000 (II) yr B.P. (Figure 9). These reefs represent a series of high sea level interstadials falling progressively further below present sea level at $-9 \pm 3 \text{ m}$, $-19 \pm 5 \text{ m}$, $-28 \pm 3 \text{ m}$, $-41 \pm 4 \text{ m}$ and $-44 \pm 2 \text{ m}$ respectively (Figures 9, 10). Low sea levels between interstadials have proved difficult to define in the absence of dateable reef deposits but inference suggests a range of 37 to 55 m below present sea level (Aharon and Chappell 1986).

In the South Pacific, presumed Last Interglacial (i.e. reef complex VII) reef limestones reach 3.48 m above present sea level at Ngatangia and 2.2 m at Nikao, Rarotonga, Southern Cook Islands, and have been attributed to glacio-eustatic fluctuations in sea level (Stoddart et al. 1985). At Makatea Island, NW Tuamotu Archipelago, cliff-veneering apron reefs, bounded at their upper margin by notch lines and caves 5-8 m above present sea level (Montaggioni et al. 1985) have been assigned to the time interval 100-140,000 yr B.P. on the basis of limited uranium-series age determinations (Veeh 1966). In the Southern Cook Islands, Last Interglacial reef limestones on Mangaia (101-135,000 yr B.P.: Veeh 1966, Spencer et al. 1987) reach 14.5 m above present sea level; raised reefs presumed to be contemporaneous with the Mangaian deposits attain heights of ~10 m or more on Atiu, Mauke and Mitiaro. Finally, at Henderson Island, presumed Last Interglacial reef units also reach ~10 m above present sea level (Spencer and Paulay, this volume). Apart from Rarotonga, all these reef limestones exhibit upper altitudinal limits considerably above characteristic Last Interglacial elevations, they must, therefore, indicate an additional component of tectonic uplift superimposed upon the sea levels associated with glacio-eustatic sea level changes. For all these localities it has been argued that uplift has resulted from the up-arching of oceanic lithosphere under loading from neighbouring volcanoes of Pleistocene age (McNutt and Menard 1978, Lambeck 1981a, 1981b, Spencer and Paulay this volume). However, it is difficult to apply this explanation to Rurutu, Austral Islands, where Last Interglacial (188-126,000 yr B.P.) limestones reach 8-10 m above present sea level and mid-plate thermal rejuvenation, in association with mid-plate hotspots, may also have been involved (Pirazzoli and Veeh 1987).

In spite of these tectonic processes, uplift rates have not been sufficient at these mid-ocean settings to raise interstadial reefs above present sea level (see Figure 9). Thus at Henderson Island where the uplift rate over the last 125,000 years has been estimated at between 0.04 and 0.09 m/1000 yr (Spencer and Paulay, this volume), deposits equivalent to New Guinea reef complex VIa should be found at ~0 to -5 m; thus they may floor the contemporary reef flat and form the shallow terrace offshore. By extension, reef complexes Va and IVa should be found in water depths of ~ -11 to 16 m and -20 to -25 m respectively; they may, at least in part, form the second, deeper coral ledge at Henderson and the broad offshore shelves, with their shelf breaks at 25-30 m, seen at Ducie atoll.

QUATERNARY ENVIRONMENTS: PALAEOCLIMATOLOGY AND PALAEOCEANOGRAPHY

As well as adapted to relatively rapid fluctuations in sea level, Pleistocene coral reefs in the Pitcairn Group must also have been subjected to a varying climatic and oceanographic environment.

It is possible to successfully reconstruct palaeoceanographic conditions by the application of biological transfer function techniques, originally devised by Imbrie and Kipp (1971), to micropalaeontological data. When mapped over the ocean floor, mathematically-defined micro-faunal assemblages from surface sediments show a close fit to water mass and current distributions. Thus changes in assemblages within cores indicates different oceanographic conditions. It is further possible to express assemblage information in terms of physical or ecological variables; in particular, equations which express sea surface temperature as a function of various assemblage have been derived using regression analysis. The down-core application of such transfer functions permits estimates of past temperatures to be made (e.g. CLIMAP Project Members 1976, 1984).

Reconstruction of sea surface temperatures for the Last Interglacial ocean suggests little change from present conditions: 60 per cent of the estimates from deep-sea cores differ from today's values by amounts less than the typical ± 1.0 to 1.5°C standard error of estimate (CLIMAP Project Members 1984). However, within the southern hemisphere subtropical gyre palaeotemperature estimates from core V19-53 (17°S ; 113°W) suggest a southern summer (February) mean sea surface temperature of 29.3°C and a souther winter (August) temperature of 27.0°C , in each case 3.9°C warmer than the equivalent present sea surface temperature at that latitude (Table 1). Were such changes to have extended south they would have had important implications for the maintenance of reef growth at Henderson, Oeno and Ducie. However, not all the deep-sea core evidence suggests such changes in temperature (Table 1).

For the last glacial maximum, at 18,000 yr B.P., deep-sea core evidence suggests that the Southern Ocean as a whole was cooler by 2°C (CLIMAP Project Members 1976). Similar reconstruction from $^{18}\text{O}/^{16}\text{O}$ ratios in reef molluscs (Konishi et al. 1974, Fairbanks and Matthews 1978, Shackleton and Matthews 1977), allowing for global ice volume effects, have given comparable results. In particular, careful studies using specimens of the giant clam *Tridacna gigas* from the Huon terrace sequences by Aharon and Chappell (1986) have suggested a cooling of $2\text{-}3^{\circ}\text{C}$ in surface ocean temperature from early (reef complexes VI-V) to late (IV-V) interstadials in the last glacial-interglacial cycle. However, as with the pre-Pleistocene and Last Interglacial record, little temperature change appears to have affected the mid-latitude gyre of the south-central Pacific (CLIMAP Project Members 1976), a pattern repeated for the Indian and S. Atlantic Oceans (Moore et al. 1981). In the north Pacific, ocean surface temperatures have been reconstructed as similar to those at present (Thompson 1981) or, for the vicinity of Hawaii, perhaps up to 2°C warmer (Rind and Peteet 1985). Rather than temperature change, the northern hemisphere glaciations led to further intensification of the atmospheric-oceanic circulation.

The south-east Trades are the driving force of surface currents in the eastern equatorial Pacific. Studies of quartz abundance distribution in core V19-29 (3°S , $83^{\circ}56'\text{W}$) have suggested that the tradewinds were more intense during the glacial phases, in particular between $73\text{-}61 \times 10^3$ yr B.P. and $43\text{-}16 \times 10^3$ yr B.P. (Figure 11; Molina-Cruz 1977). Changing distributions of radiolarian assemblages in deep-sea core (Moore et al. 1981) indicate that the strengthening of the glacial trades was accompanied by a more meridional pattern of wind stress and an intensification of the equatorial surface circulation, with

increased equatorial countercurrent and decreased equatorial undercurrent (Romine 1982). The eastern boundary flows showed greater westward penetration with Peru Current upwelling reaching its most westerly at the glacial maximum (Romine and Moore 1981).

Stoddart's (1973) calculation of the effect of glacial sea surface temperatures on the extent of the reef seas suggests a simple, latitudinal shift towards the equator of the area of the south Pacific able to support coral growth (Figure 12). However, from the more comprehensive deep-sea core coverage now available it is clear that the major glacial to inter-glacial changes in climatic and oceanographic conditions produced variations in the intensity of circulation not latitudinal migrations of the climatic belts (Figure 12). The expansion of the polar seas at high latitudes resulted in the compression of the areas occupied by the sub-polar and subtropical-transitional water masses and left the tropical water masses largely intact. Whilst the islands of the Pitcairn Group may have been subjected to a windier and perhaps stormier glacial climate, with greater contrasts between windward and leeward shores, it is unlikely that water temperatures were appreciably cooler or more inimical to coral growth than at the present time (Figure 12).

HOLOCENE SEA LEVEL CHANGE IN THE SOUTH-CENTRAL PACIFIC

The explanation of global Holocene sea level changes lies in the interaction between the volume, melting history and location of sources of meltwater from the decay of ice sheets and deformation of the earth's crust due to both the unloading of continental ice (glacio-isostasy) and the loading of the oceans by meltwater (hydro-isostasy) (Farrell and Clark 1976). Both these sets of controls have varied considerably in their magnitude at-a-point in time over the globe and at-a-point in space during the Holocene transgression, thus giving rise to a range of Holocene sea level curves not only between different oceanic areas (e.g. Peltier et al. 1978) but also between continental margins and oceanic islands (Walcott 1972) and within ocean basins (e.g. Nakiboglu et al. 1983, Lambeck and Nakada 1985).

There is widespread evidence through the south-central Pacific for a higher-than-present Holocene sea level of $\pm 1.0\text{m}$ between $\sim 6,000$ years B.P. and $\sim 2,000$ years B.P. (Figure 13). Field observations suggest that this event is represented on at least some of the islands of the Pitcairn Group.

At Ducie atoll, the lagoon shore of the largest *motu*, Acadia Island (Rehder and Randall 1975), is characterised by superficially cemented sheets and slabs of coral rubble, estimated to be 0.30-0.45 m above present mean sea level at the lagoon edge. The morphology of these deposits varies along the lagoon shore: at the eastern extremity of the island the deposits appear to represent lithified, coalescing washover fans (Plate 1) whereas on the central lagoon shore extensive areas of exfoliating sheets, comparable to the Holocene conglomerates described from French Polynesia by Montaggioni and Pirazzoli (1984) are characteristic (Plate 2). Towards the western end of the *motu* these sheets are replaced by boulder streams of uncemented coral heads, separated by areas of coral stick gravel (Plate 3).

On the south-western margin of the main wooded island at Oeno atoll (Figure 14) a lower beachrock unit with locally abundant acroporids appears bevelled and overlain by an upper beachrock unit containing corals and *Tridacna* (Plate 4). Massive beachrocks also locally overlie the bevelled beachrock and on the north-west coast appear to provide the basement for island sediments (Figure 14).

By comparison with the atolls, however, there is less evidence for a Holocene high stand on Henderson Island. The reasons for this are partly topographical: the presumed high stand deposits on Ducie and Oeno have accumulated, and subsequently been preserved, on lagoon shores whereas the inner reef environment on Henderson offers no such protection (and see Spencer and Paulay this volume). However, additional tectonic and tectonically-related factors must also be considered.

Attempts to model Holocene sea level change in the Pacific basin have been complicated by the presence of two unknowns: the melting histories of the Late Pleistocene ice sheets (and specifically the relative contributions of Arctic and Antarctic ice) and the flow in the lower mantle, or mantle viscosity. In addition, local hydro-isostatic adjustments take place around islands (Nakada 1986) and on continental shelves (Chappell et al. 1982) because of the differential loading between island interiors and outer margins and inner and outer continental shelves respectively; fortunately, however, such effects are unimportant at scales of 10 km or less. Finally, however, the possible role of migratory geoidal highs in determining the sea level record, first postulated by Mörner (1976), also needs to be evaluated in reef environments (Nunn 1986).

The sea level record can in turn be masked by non-related tectonic processes. One such process is the subsidence of young volcanic islands; subsidence rates of $1-2 \text{ mm yr}^{-1}$, over 8,000 years, and $0.14-0.15 \text{ mm yr}^{-1}$, over 5,000 years, have been suggested for Oahu, Hawaii Islands (Nakiboglu et al. 1983) and Moorea-Tahiti, Society Islands (Pirazzoli and Montaggioni 1985) respectively. Furthermore, Pirazzoli and Montaggioni (1985) have argued that regional variations in Holocene sea level curves between the Society Islands and the N.W. Tuamotu atolls (and perhaps within the Societies themselves (Pirazzoli 1983)) have resulted from lithospheric flexure around the Tahiti-Moorea volcanic load. Presumably such arguments might be used to differentiate sea level histories between flexure-affected Henderson on the one hand and Ducie and Oeno atolls on the other.

Taking these complications into account, and using constrained values for mantle viscosity and favoured melting models from the comparison of model sea level curves with observed records from N. Australia and New Zealand (Lambeck and Nakada 1985, Lambeck and Nakiboglu 1986), it is clear that the Pacific high stand at $\sim 6,000$ years B.P. was largely due to the control of mantle viscosity, with the potential contribution of Antarctic ice controlling the 'peakedness' of the event (Lambeck and Nakada 1985, Nakada and Lambeck 19867). The model fit to observed values provides a useful first approximation, although high sea levels appear to have been sustained into a period when model predictions suggest a gradual fall in sea level (Figure 15). Clearly more detailed observations, with radio-carbon dating control, from the Pitcairn group would be useful in the refining of sea level models for this period.

CONCLUSIONS

This review has shown that while our knowledge of the genesis and evolution of the south-central Pacific, and of the Pitcairn group of islands within this region, has been considerably increased on the last decade, large gaps still remain in our understanding of volcanic island development and attendant coral reef construction.

While remarkable progress has been made through the use of remotely-sensed data many of the research questions which have been generated by these studies now require direct field-testing. This might be achieved by programmes of deep-drilling, to both ocean floor

and island basements, and by more comprehensive geological and geomorphological surveys than have hitherto been achieved.

Although there have been notable exceptions, mid-plate, mid-latitude locations have been relatively neglected in scientific terms by comparison with plate-marginal settings and both high and low latitude environments. However, a much greater knowledge of the processes active at intra-plate settings at intermediate latitudes will be required for forthcoming reconstructions of global tectonic and environmental histories and the predictions at and around remote islands such as those of the Pitcairn group are of significance, both to answer specific research hypotheses and in much broader contexts.

REFERENCES

Abers G A, Parsons B and Weissel J K 1988. Seamount abundances and distributions in the southeast Pacific. *Earth Planet Sci. Lett.* 87 : 137-151.

Aharon P and Chappell J 1986. Oxygen isotopes, sea level changes and the temperature history of a coral reef environment in New Guinea over the last 10^5 years. *Palaeogeog., Palaeoclimatol., Palaeoecol.* 56 : 337-379.

Bloom A L, Broecker W S, Chappell J, Matthews R K and Mesolella K J 1974. Quaternary sea level fluctuations on a tectonic coast: New $^{230}\text{Th}/^{234}\text{U}$ dates from the Huon Peninsula, New Guinea. *Quat. Res.* 4 : 185-205.

Bonatti E and Harrison C G A 1976. Hot lines in the Earth's mantle. *Nature (Lond.)* 263 : 402-404.

Bonatti E, Harrison C G A, Fisher D E, Honnorez J, Schilling J-G, Stipp J J and Zentilli M 1977. Eastern volcanic chain (southeast Pacific): a mantle hot line. *J. Geophys. Res.* 82 : 2457-2478.

Broecker W S and Van Donk J 1970. Insolation changes, ice volumes and the $\text{O}18$ record in deep sea sediments. *Rev. Geophys. Space Phys.* 8 : 169-198.

Brousse R 1985. The age of the islands in the Pacific Ocean: volcanism and coral-reef build up. *Proc. Fifth Int. Coral Reef Congr., Tahiti*, 6 : 389-400.

Brousse R, Phillipet J-C, Guille G and Bellon H 1972. Géochronometrie des îles Gambier (Ocean Pacifique). *C. R. somm. seanc. Acad. Sci. Paris* 274 : 1995.

Buck W R 1985. When does small-scale convection begin beneath oceanic lithosphere? *Nature (Lond.)* 313 : 775-777.

Buck W R and Parmentier E M 1986. Convection beneath young oceanic lithosphere: Implications for thermal structure and gravity. *J. Geophys. Res.* 91 : 1961-1974.

Calmant S and Cazenave A 1986. The effective elastic lithosphere under the Cook-Austral and Society Islands. *Earth Planet. Sci. Lett.* 77 : 187-202.

Calmant S and Cazenave A 1987. Anomalous elastic thickness of the oceanic lithosphere in the south-central Pacific. *Nature (Lond.)* 328 : 236-238.

Canadian Hydrographic Service 1982. General Bathymetric Chart of the Oceans (GEBCO), 5th edn. Ottawa.

Cazenave A and Dominh K 1984. Geoid heights over the Louisville ridge (South Pacific). *J. Geophys. Res.* 89 : 11171-11179.

Chappell J 1974. Geology of coral terraces, Huon Peninsula, New Guinea: A study of Quaternary tectonic movements and sea level changes. *Bull. Geol. Soc. Amer.* 85 : 553-5570.

Chappell J 1983. A revised sea level record for the last 300,000 years from Papua New Guinea. *Search* 4 : 99-101.

Chappell J, Rhodes E G, Thom B G and Wallensky E 1982. Hydro-isostasy and sea-level isobase of 5500 B.P. in north Queensland. *Marine Geol.* 49 : 81-90.

Chappell J and Shackleton N J 1986. Oxygen isotopes and sea level. *Nature* 324 : 137-140.

CLIMAP Project Members 1976. The surface of the ice-age earth. *Science* 191 : 1131-1137.

CLIMAP Project Members 1984. The Last Interglacial ocean. *Quat. Res.* 21 : 123-224.

Cochran J R 1986. Variations in subsidence rates along intermediate and fast spreading mid-ocean ridges. *Geophys. J. Roy. Astron. Soc.* 87 : 421-454.

Delibrias G and Pirazzoli P A 1985. Sea-level changes, Appendix 1 - Lists of Radiocarbon dates from the Society Islands In:- Pirazzoli P A, Leeward Islands - Maupiti, Tupai, Bora Bora, Huahine - Society Archipelago. *Proc. Fifth. Int. Coral. Congr., Tahiti*, 1: 63-65.

Detrick R S and Crough S T 1987. Island subsidence, hot spots and lithospheric thinning. *J. Geophys. Res.* 84 : 4713-4729.

Diament M and Baudry N 1987. Structural trends in the Southern Cook and Austral archipelagoes (South Central Pacific) based on an analysis of SEASAT data: geodynamic implications. *Earth Planet. Sci. Lett.* 85 : 427-438.

Dodge R E, Fairbanks R G, Benninger L K and Maurasse F 1983. Pleistocene sea levels from raised coral reefs of Haiti. *Science* 219 : 1423-1425.

Duncan R A and Clague D A 1985. Pacific plate motion recorded by linear volcanic chains, In:- Nairn A E M, Stehli F G and Uyeda, S, editors, The Ocean Basins and their margins: The Pacific Ocean vol. 7A, Plenum: New York, 89-121.

Duncan R A and Hargraves R B 1984. Plate tectonic evolution of the Caribbean region in the mantle reference frame. In:- Bonini W E, Hargraves R B, and Shagam R, editors, The Caribbean-South America Plate Boundary and Regional Tectonics, Mem., Geol. Soc. America 162 : 81-93.

Duncan R A, McDougall I, Carter R M, and Coombs D S 1974. Pitcairn Island - another Pacific hot spot? *Nature (Lond.)* 251 : 679-682.

Edwards A R 1975. southwest Pacific Cenozoic Paleogeography and an integrated Neogene palaeocirculation model In:- Andrews J E and co-workers, Initial Reports of the Deep Sea Drilling Project, 30, U.S. Govt. Printing Office: Washington D.C., 667-684.

Emiliani C 1955. Pleistocene temperatures. *J. Geol.* 63 : 538-578.

Epstein S, Buchsbaum R, Lowenstam H A and Urey H C 1953. Revised carbonate-water isotopic temperature scale. *Bull Geol Soc. Amer.* 64 : 1315-1326.

Fairbanks R G and Matthews RK 1978. The marine oxygen isotope record in Pleistocene coral, Barbados, West Indies. *Quat. Res.* 10 : 181-196.

Farrell W E and Clark J A 1976. On postglacial sea level. *Geophys. J. R. Astron Soc.* 46 : 647-667.

Fischer K M, McNutt M K and Shure L 1986. Thermal and mechanical constraints on the lithosphere beneath the Marquesas swell. *Nature (Lond.)* 322 : 733-736.

Hamilton W 1979. Tectonics of the Indonesian Region. U.S.G.S. Prof. Pap. 1078 : 1-345.

Handschumacher S W 1975. Post-Eocene plate tectonics of the eastern Pacific. In:- Sutton G H, Maghnani M H and Moberly R, editors, The Geophysics of the Pacific Ocean Basin and its Margin. Geophysical Monograph 19, Amer. Geophys. Un.: Washington D.C., 177-202.

Herron E M 1972. Sea-floor spreading and the Cenozoic history of the East-Central Pacific. *Bull. Geol. Soc. Amer.* 83 : 1671-1692.

Herron E M and Tucholke B E 1976. Sea-floor magnetic patterns and basement structure in the south eastern Pacific. In:- Hollister C D and co-workers, Initial Reports of the Deep Sea Drilling Project, 35, U.S. Govt. Printing Office: Washington D.C., 263-278.

Hey R N, Naar D F, Kleinrock M C, Morgan W J P, Morales E And Schilling J-G 1985. Microplate tectonics along a superfast seafloor spreading system near Easter Island. *Nature (Lond.)* 317 : 320-325.

Imbrie J and Kipp N G 1971. A new micropalaeontological method for quantitative palaeoclimatology: application to a late Pleistocene Caribbean core. In:- Turekian KK, editor, Late Cenozoic Glacial Ages. Yale University Press: New Haven, Conn., 171-181.

Ingle J C 1981. Origin of Neogene diatomites around the North Pacific rim. In:- Garrison R E and co-editors, The Monterey Formation and related siliceous rocks of California. Soc. Econ. Palaeont. Mineral., Spec. Publ., 159-180.

Janecek T R and Rea D K 1985. Quaternary fluctuations in the northern hemisphere tradewinds and westerlies. *Quat. Res.* 24 : 150-163.

Keigwin L D 1978. Pliocene closing of the Panama Isthmus based on biostratigraphic evidence from nearby Pacific Ocean and Caribbean cores. *Geology* 6 : 630-634.

Keigwin L D and Keller G 1984. Middle Oligocene cooling from equatorial Pacific DSDP Site 77B. *Geology* 12 : 16-19.

Kennett J P 1977. Cenozoic evolution of Antarctic glaciations, the Circum-Antarctic Ocean and their impact on global palaeoceanography. *J. Geophys. Res.* 82 : 3843-3860.

Kennett J P 1982. Marine Geology. Prentice-Hall: New Jersey, 813 pp.

Kennett J P, Keller G and Srinivasan M S 1985. Miocene planktonic foraminiferal biogeography and palaeoceanography development of the Indo-Pacific region. In: Kennett J P, editor, The Miocene Ocean: Palaeoceanography and Biogeography. Geol Soc. Amer. Mem. 163 : 197-136.

Konishi K, Omura a and Nakamichi O 1974. Radiometric coral ages and sea level records from the late Quaternary reef complexes of the Ryukyu islands. *Proc. 2nd. Int. Coral Reef. Symp., Brisbane 2* : 595-613.

Lambeck K 1981a. Flexure of the ocean lithosphere from island uplift, bathymetry and geoid height observations: the Society Islands. *Geophys. J. Roy. Astron. Soc.* 67 : 91-114.

Lambeck K 1981b. Lithosphere response to volcanic loading in the southern Cook Islands. *Earth Planet. Sci. Lett.* 55 : 482-496.

Lambeck K and Nakada M 1985. Holocene fluctuations in sea level: Constraints on mantle viscosity and melt-water sources. *Proc. Fifth Int. Coral Reef Congr., Tahiti, 3* : 79-84.

Lambeck K and Nakibolu S M 1986. Recent sea level changes in the South-West Pacific In: Reilly W and Hartford B E, editors, Recent Crustal Movements of the Pacific Region. Bull. Roy. Soc. N.Z. 24 : 485-491.

Leinen M 1979. Biogenic silica accumulation in the central equatorial Pacific and its implications for Cenozoic palaeoceanography. *Bull. Geol. Soc. Amer.* 90 : 1310-1376.

McAdoo D C and Sandwell D T 1985. Folding of oceanic lithosphere. *J. Geophys. Res.* 90 : 8563-8569.

McNutt M K 1984. Lithospheric flexure and thermal anomalies. *J. Geophys. Res.* 89 : 11180-11194.

McNutt M K and Fischer K 1986. The south Pacific superswell. (Abstr). *EOS* 67 (44) : 1220.

McNutt M K and Menard H W 1978. Lithosphere flexure and uplift atolls. *J. Geophys. Res.* 83 : 1206-1212.

McNutt and Menard H W 1982. Constraints on yield strength in the oceanic lithosphere derived from observations of flexure. *Geophys. J. Roy. Astron. Soc.* 71 : 363-394.

Mammerickx J, Heron E M and Dorman L M 1980. Evidence for two fossil ridges in the southeast Pacific. *Bull. Geol. Soc. Amer.* 91 : 263-271.

Matthews R K 1973. Relative elevation of late Pleistocene high sea level stands: Barbados uplift rates and their implications. *Quat. Res.* 3 : 147-153.

Matthews and Poore R Z 1980. Tertiary ¹⁸O record and glacio-eustatic sea-level fluctuations. *Geology.* 8 : 501-504.

- Miller K G and Fairbanks R G 1985. Oligocene to Miocene carbon isotope changes and abyssal circulation changes. In:- Sundquist E T and Broecker W A, editors, The Carbon Cycle and Atmospheric CO₂: Natural variations Archean to Present. Geophys. Monogr. 32, Amer. Geophys. Un.: Washington D.C., 469-486.
- Molina-Cruz A 1977. The relation of the southern tradewinds to upwelling processes during the last 75,000 years. *Quat. Res.* 8 : 324-338.
- Montaggioni L F and Pirazzoli P A 1984. The significance of exposed coral conglomerates from French Polynesia (Pacific Ocean) as indicators of recent relative sea-level changes. *Coral Reefs* 3 : 29-42.
- Montaggioni L F, Richards G, Gabrie C, Monteforte M, Naim O, Payri C and Salvat B 1985. Les recifs coralliens de l'île de Makatea, archipel des Tuamotu, Pacifique central: geomorphologie et repartition des peulements. *Ann. Inst. Oceanogr.*, Paris 61 : 1-15.
- Montaggioni L F and co-workers 1985. Geology and marine biology of Makatea, an uplifted atoll, Tuamotu Archipelago, central Pacific Ocean. *J. Coastal Res.* 1 : 165-171.
- Moore T C Jr, Burckle L H, Geitzenaur K, Luz B, Molina-Cruz A, Robertson J H, Sachs H, Sancetta C, Theide J, Thompson P and Wenkam C 1980. The reconstruction of sea surface temperatures in the Pacific Ocean of 18,000 yr B.P. *Marine Micropalaeontology* 5 : 215-247.
- Moore T C Jr, Hutson W H, Kipp N, Hays J D, Prell W, Thompson P and Boden G 1981. The biological record of the ice-age ocean. *Palaeogeog., Palaeoclimatol., Palaeoecol.* 35 : 357-370.
- Moore W S 1982. Late Pleistocene sea-level history. In:- Ivanovich M and Harmon R S, editors, Uranium series disequilibrium: applications to environmental problems. Clarendon Press: Oxford, 481-496.
- Morgan W J 1978. Rodriguez, Darwin, Amsterdam, ..., a second type of hotspot island. *J. Geophys. Res.* 83 : 5355-5360.
- Mörner N-A 1976. Eustasy and geoidal changes. *J. Geol.* 84 : 123-151.
- Nakada M 1986. Holocene sea levels in oceanic islands: implications for the rheological structure of the earth's mantle. *Tectonophys.* 121 : 263-276.
- Nakada M and Lambeck K 1987. Glacial rebound and relative sea-level variations: a new appraisal. *Geophys. J. Roy Astron Soc.* 90 : 171-224.
- Nakiboglu S M, Lambeck K and Aharon P 1983. Postglacial sea levels in the Pacific: implications with respect to deglaciation regime and local tectonics. *Tectonophys.* 91 : 335-338.
- Nunn P D 1986. Implications of migrating geoid anomalies for the interpretation of high-level fossil coral reefs. *Bull. Geol. Soc. Amer.* 97 : 946-952.
- Okal E A 1984. Intraplate seismicity of the southern part of the Pacific plate. *J. Geophys. Res.* 89 : 10053-10071.

- Okal E A and Bergeal J-M 1983. Mapping the Miocene Farallon Ridge jump on the Pacific plate: a seismic line of weakness. *Earth Planet. Sci. Lett.* 63 : 113-122.
- Okal E A and Cazenave A 1985. A model for the plate evolution of the east-central Pacific based on SEASAT investigations. *Earth Planet. Sci Lett.* 72 : 99-116.
- Okal E A, Talandier J, Sverdrup K A and Jordan T H 1980. Seismicity and tectonic stress in the south-central Pacific. *J. Geophys. Res.* 85 : 6479-6495.
- Parsons B and Sclater J G 1977. An analysis of the variation of ocean floor bathymetry and heat flow with age. *J. Geophys. Res.* 82 : 803-827.
- Peltier W R, Farrell W E and Clark J A 1978. Glacial isostasy and relative sea level; a global finite element model. *Tectonophys.* 46 : 81-110.
- Pirazzoli P A 1983. Mise en evidence d'une flexure active de la lithosphere dans l'archipel de la Société (Polynésie française) d'après la position des rivages de la fin de l'Holocène. *C. R. somm. Seanc. Acad. Sci., Paris*, 296 : 695-698.
- Pirazzoli P A 1985. Sea level changes in Rangiroa. *Proc Fifth Int. Coral Reef Congr., Tahiti*, 1 : 195-198.
- Pirazzoli P A 1987a. A reconnaissance and geomorphological survey of Temoe Atoll, Gambier Islands (South Pacific). *J. Coastal Res.* 3 : 307-323.
- Pirazzoli P A 1987b. French Polynesia. In:- Yoshikawa T, editor, Inventory of Quaternary Shorelines: Pacific and Indian Oceans Region (INQUA Commission on Quaternary Shorelines), NODAI Research Institute, Tokyo University of Agriculture, Tokyo, Japan, 128-130.
- Pirazzoli P A, Delibrias G, Montaggioni L F, Saliège J F and Vergnaud-Grazzini C 1978a. Vitesse de croissance laterale des platiers et évolution morphologique récente de l'atoll de Reao, Iles Tuamotu, Polynésie française. *Ann. Inst. Oceanogr., Paris*, 63 : 57-68.
- Pirazzoli P A and Montaggioni L F 1985. Lithosphere deformation in French Polynesia (French Pacific) as deduced from Quaternary shorelines. *Proc. Fifth. Int. Coral Reef Congr., Tahiti*, 3 : 195-200.
- Pirazzoli P A and Montaggioni L F 1986. Late Holocene sea-level changes in the Northwest Tuamotu Islands, French Polynesia. *Quat. Res.* 25 : 350-368.
- Pirazzoli P A, Montaggioni L F, Delibrias G, Faure G and Salvat B 1985. Late Holocene sea-level changes in the Society Islands and in the northwest Tuamotu atolls. *Proc. Fifth. Int. Coral Reef Congr., Tahiti*, 3 : 131-136.
- Pirazzoli P A, Montaggioni L F, Vergnaud-Grazzini C and Saliege J F 1987b. Late Holocene sea levels and coral reef development in Vahitahi Atoll, Eastern Tuamotu Islands, Central Pacific. *Marine Geol.* 76 : 105-116.
- Pirazzoli P A and Veeh H H 1987. Age $^{230}\text{T}_{\text{H}}/^{234}\text{U}$ d'une encoche émergée et vitesses de soulèvement quaternaire à Rurutu, îles Australes. *C. R. somm. Seanc. Acad. Sci, Paris*, 305 (II) : 919-923.

Pontoise and co-workers 1986. Levés SEABEAM dans l'Archipel des Iles Australes: confirmation d'une nouvelle méthode de localisation des monts sous-marins basée sur l'analyse des données SEASAT. C. R. somm. Seanc. Acad. Sci., Paris, 303 : 563-568.

Rea D K and Bloomstine M K 1986. Neogene history of the South Pacific Tradewinds: Evidence for hemispherical symmetry of atmospheric circulation. Palaeogeogr., Palaeoclimatol., Palaeoecol. 55 : 55-64.

Rea D K and Janecek T R 1982. Late Cenozoic changes in atmospheric circulation deduced from North Pacific eolian sediments. Marine Geol. 49 : 149-167.

Rehder H A and Randall J E 1975. Ducie Atoll; its history, physiography and biota. Atoll Res. Bull. 183 : 1-40.

Rhind D and Peteet D 1985. Terrestrial conditions at the last glacial maximum and CLIMAP sea-surface temperature estimates: are they consistent? Quat. Res. 24 : 1-22.

Richter F M 1973. Convection and the large scale circulation of the mantle. J. Geophys. Res. 78 : 8735-8745.

Roe K K and Burnett W C. Uranium geochemistry and dating of Pacific island apatite. Geochim. cosmochim. acta. 49 : 1581-1592.

Romine K 1982. Late Quaternary history of atmospheric and oceanic circulation in the eastern equatorial Pacific. Marine Micropalaeontology 7 : 163-187.

Romine K and Moore T C Jr 1981. Radiolarian assemblage distributions and palaeoceanography of the eastern equatorial Pacific Ocean during the last 127,000 years. Palaeogeog., Palaeoclimatol., Palaeoecol. 35 : 281-314.

Sailor R V and Okal E A 1983. Applications of SEASAT altimeter data in seismotectonic studies of the south-central Pacific. J. Geophys. Res. 88 : 1572-1580.

Sandwell D T 1984. A detailed view of the South Pacific geoid from satellite altimetry. J. Geophys. Res. 89 : 1089-1104.

Savin S M 1977. The history of the earth's surface temperature during the past 100 million years. Ann. Rev. Earth Planet. Sci. 5 : 319-355.

Savin S M and co-workers 1981. Miocene benthic foraminiferal isotope records: A synthesis. Marine Micropalaeontology. 6 : 423-450.

Schilling J-G, Sigurdsson H, David A N and Hey R N 1985. Easter microplate evolution. Nature. (Lond.) 317 : 325-331.

Schofield J C 1970. Notes on late Quaternary sea levels, Fiji and Rarotonga. NZ J. Geol. Geophys. 13 : 199-206.

Shackleton N J 1967. Oxygen isotope analyses and palaeotemperatures reassessed. Nature. (Lond.) 215 : 15-17.

- Shackleton N J 1984. Oxygen isotope evidence for Cenozoic climatic change. In: Brenchley P J, editor, Fossils and Climate. Wiley : Chichester, 27-34.
- Shackleton N J 1986. Palaeogene stable isotope events. *Palaeogeog., Palaeoclimatol., Palaeocol.* 57 : 91-102.
- Shackleton N J and co-workers 1984a. Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region. *Nature (Lond.)* 307 : 620-623.
- Shackleton N J, Hall M A and Boersma A 1984b. Oxygen and carbon isotope data for leg 74 foraminifers. In: Moore T C Jr and co-workers, Initial Reports of the Deep Sea Drilling Project, 74 U.S. Govt. Printing Office: Washington D.C., 599-612.
- Shackleton N J and Kennett J P 1977. Palaeotemperature history of the Cenozoic and the initiation of Antarctic glaciation; oxygen and carbon isotope analyses in DSDP Sites 277, 279 and 281. In: Kennett J P and co-workers, Initial Reports of the Deep Sea Drilling Project, 29, U.S. Govt. Printing Office: Washinton D.C., 743-755.
- Shackleton N J and Matthews R K 1977. Oxygen isotope stratigraphy of late Pleistocene coral terraces in Barbados. *Nature (Lond.)* 268 : 618-620.
- Shackleton N J and Opdyke N D 1973. Oxygen isotope and palaeomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 and 10^6 year scale. *Quat. Res.* 3 : 39-55.
- Shackleton N J and Opdyke N D 1977. Oxygen isotope and paleomagnetic evidence for early northern hemisphere glaciation. *Nature (Lond.)* 270 : 216-219.
- Spencer T, Stoddart D R and Woodroffe C D 1987. Island uplift and lithospheric flexure; observations and cautions from the South Pacific. *Zeits. für Geomorph. NF., Suppl. Bd.* 63 : 87-102.
- Stoddart D R 1973. Coral reefs; the last two million years. *Geography* 58 : 313-323.
- Stoddart D R, Spencer T and Scoffin T P 1985. Reef growth and karst erosion on Mangaia, Cook Islands: a re-interpretation. *Zeits. für Geomorph., Suppl. Bd.* 57 121-140.
- Ten Brink U S and Watts A B 1985. Seismic stratigraphy of the flexural moat flanking the Hawaiian Islands. *Nature (Lond.)* 317 : 421-424.
- Thompson P R 1981. Planktonic foraminifera in the western north Pacific during the last 150,000 years; Comparison of modern and fossil assemblages. *Palaeogeog., Palaeoclimatol., Palaeocol.* 35 : 241-279.
- Turner D L and Jarrard R D 1982. K-Ar dating of the Cook-Austral Island chains; a test of the hot-spot hypothesis. *J. Vol. Geotherm. Res.* 12 : 187-220.
- Van Andel T J H, Heath G R, and Moore T C Jr 1975. Cenozoic history and palaeoceanography of the Central Equatorial Pacific Ocean. *Mem. Geol. Soc. Amer.* 143 : 1-134.

- Veeh H H 1966. $\text{Th}^{230}/\text{U}^{238}$ and $\text{U}^{234}/\text{U}^{238}$ ages of Pleistocene high sea level stand. *J. Geophys. Res.* 71 : 3379-3386.
- Walcott R L 1972. Past sea levels, eustasy and deformation of the earth. *Quat. Res.* 2 : 1-14.
- Watts A B 1978. An analysis of isostasy in the world's oceans. 1. Hawaiian-Emperor Seamount chain. *J. Geophys. Res.* 83 : 5989-6004.
- Watts A B, Bodine J H and Ribe N M 1980. Observations of the flexure and the geological evolution of the Pacific Ocean basin. *Nature (Lond.)* 283 : 532-537.
- Watts A B and Ribe N M 1984. On goeid heights and flexure of the lithosphere at seamounts. *J. Geophys. Res.* 89 : 11152-11170.
- Watts A B, Ten Brink U S, Buhl P and Brocher T M 1985. A multi-channel seismic study of lithospheric flexure across the Hawaiian seamount chain. *Nature (Lond.)* 315 : 105-111.
- Williams C A 1986. An oceanwide view of Palaeogene plate tectonic events. *Palaeogeog., Palaeoclimatol., Palaeoecol.* 57 : 3-25.
- Winterer E L and Sandwell D T 1987. Evidence from en-echelon cross-grain ridges for tensional cracks in the Pacific plate. *Nature (Lond.)* 329 : 534-537.
- Woodruff F, Savin S M and Douglas R G 1981. Miocene stable iostope record: a detailed Pacific Ocean study and its palaeoclimatic implications. *Science* 212 : 665-668.
- Yonekura N Y, Matsushima Y, Meade E and Kayanne H 1984. Holocene sea-level changes in the southern Cook Islands. Sea level changes and tectonics in the Middle Pacific. Report of the HIPAC Project in 1981, 1982 and 1983 (Department of Earth Sciences, Kobe University, Japan), 113-136.
- Yonekura N Y, Ishii T, Saito Y, Matsumoto E and Kayanne H 1986. Geologic and geomorphic development of Holocene fringing reefs of Mangaia Island, the South Cooks. Sea level changes and the tectonics in the Middle Pacific. Report of the HIPAC Project in 1984 and 1985 (Second Research) (Department of Earth Sciences, Kobe University, Japan), 44-69.

Core	Location	SST _w (°C) ¹			SST _s (°C) ¹		
		5e ²	Atlas ³	Core Top ⁴	5e ²	Atlas ³	Core Top ⁴
V19-53 (Moore et al. 1980)	17°01'S 113°31'W	27.0 ± 3.0	23.1	23.9	29.3 ± 1.5	25.4	28.3
V32-126 (Thompson 1981)	35°19'N 117°55'E	18.9 ± 1.5	16.0	16.8	27.0 ± 1.5	24.6	25.2
V21-146 (Moore et al. 1980)	37°41'N 163°02'E	17.3 ± 2.4	14.5	13.5	24.8 ± 1.8	24.0	22.0

¹ SST_w = winter sea surface temperature in February (N. Hemisphere) or August (S. Hemisphere).

SST_s = summer sea surface temperature in August (N. Hemisphere) or February (S. Hemisphere).

² Last Interglacial seasonal sea surface temperature (with standard error of estimate) from transfer functions applied to deep sea core.

³ Modern SST from atlas values.

⁴ Holocene SST from core top sediments

TABLE 1 SEA SURFACE TEMPERATURES IN SELECTED MID-LATITUDE DEEP-SEA CORES:
ISOTOPIC STAGE 5e COMPARED TO MODERN ESTIMATES

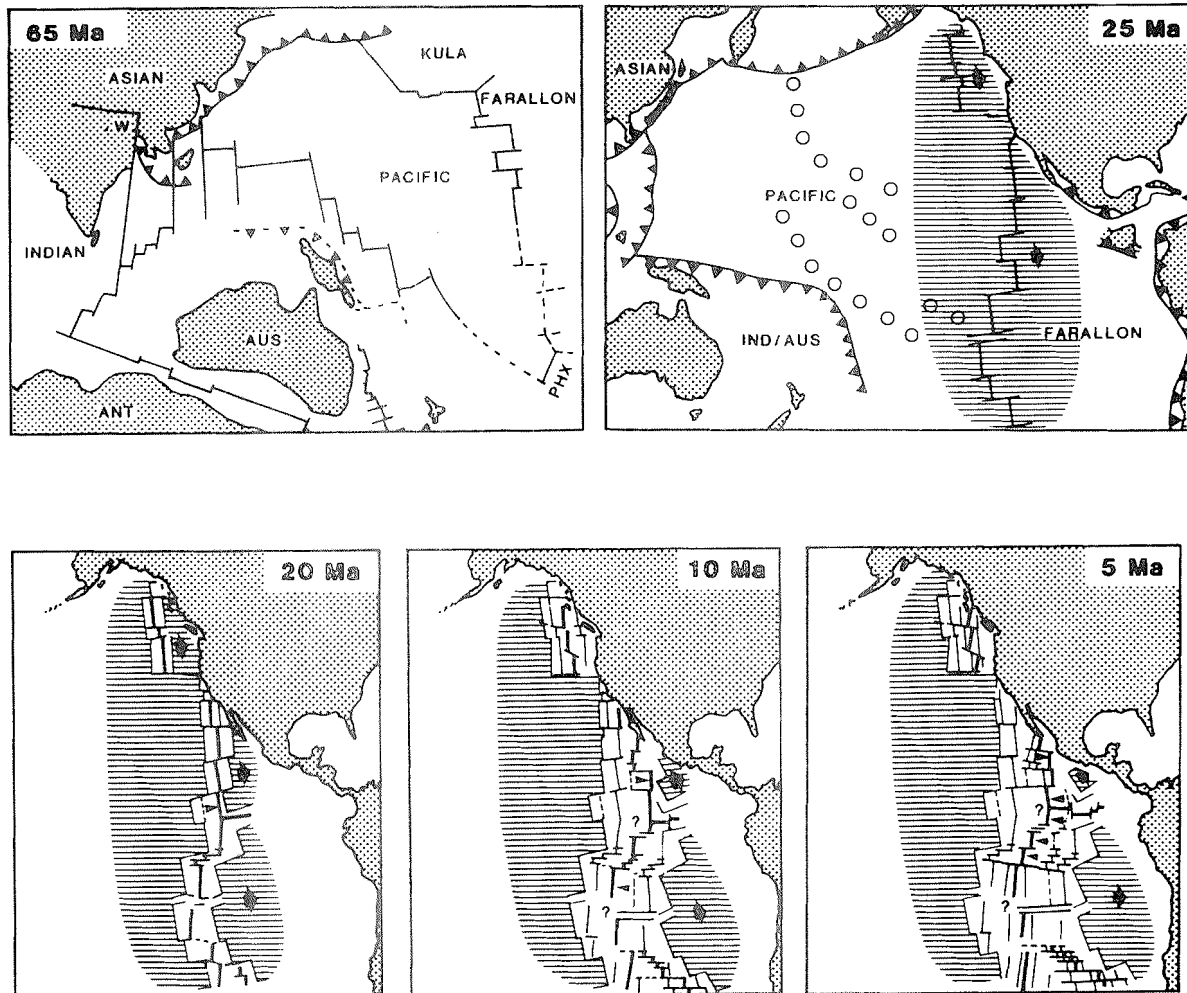


Figure 1. The Pacific at the beginning (65 million years B.P.) and end (25 million years B.P.) of the Palaeogene (after Williams 1986) and subsequent plate boundary reorganisation 20 to 50 million years B.P. (after Handschumacher 1975 and other sources). Open circles indicate island chains of volcanoes and seamounts.

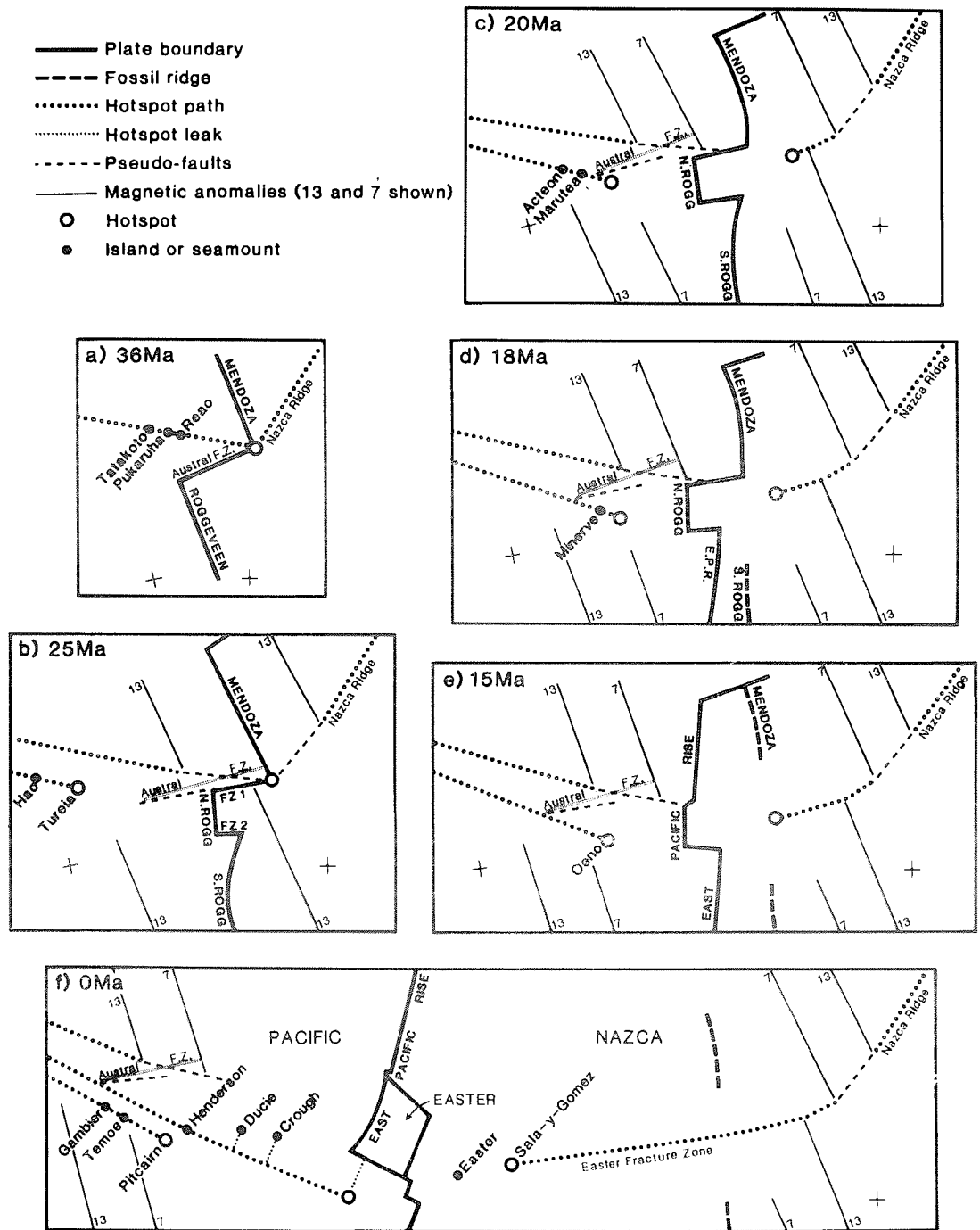


Figure 2. Plate tectonic evolution of the south-central Pacific 36 millions years B.P. to present (after Okal and Cazenave 1985). Island/Seamount names noted only on first time-frame appearance.

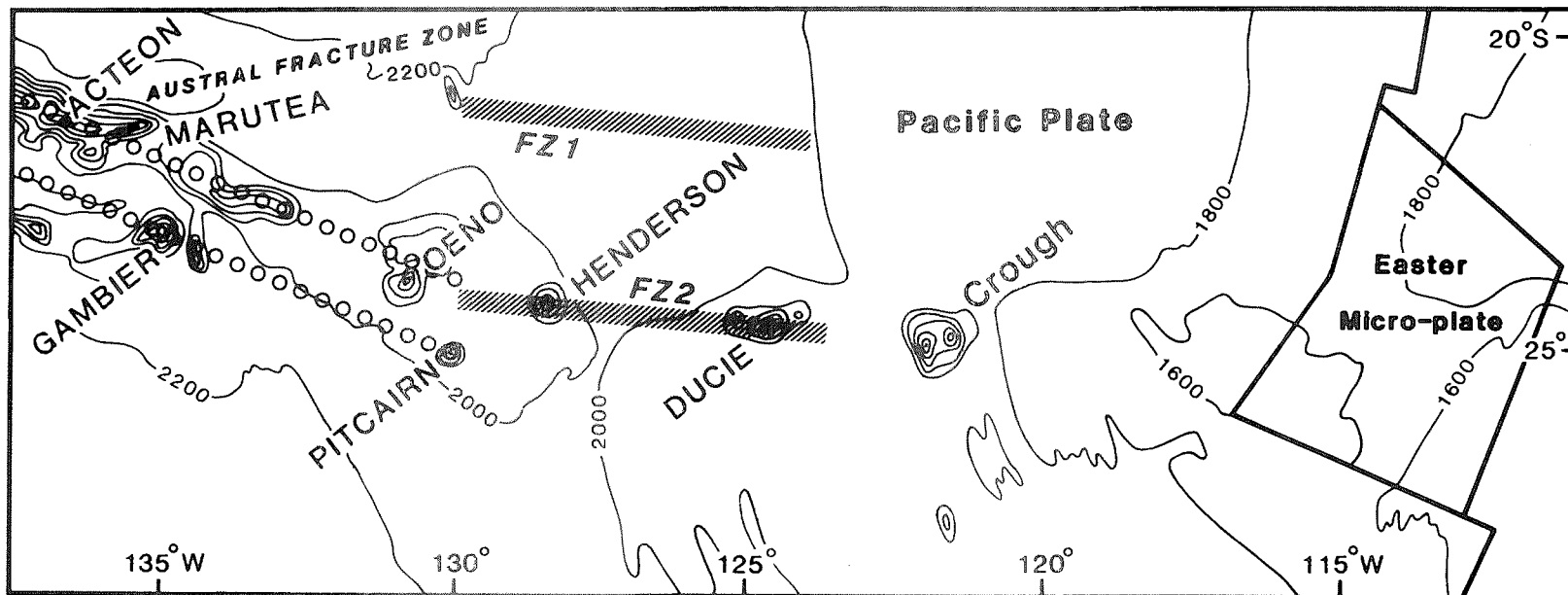


Figure 3. Regional bathymetry; and principal tectonic features and lineations (after Okal and Cazenave 1985).

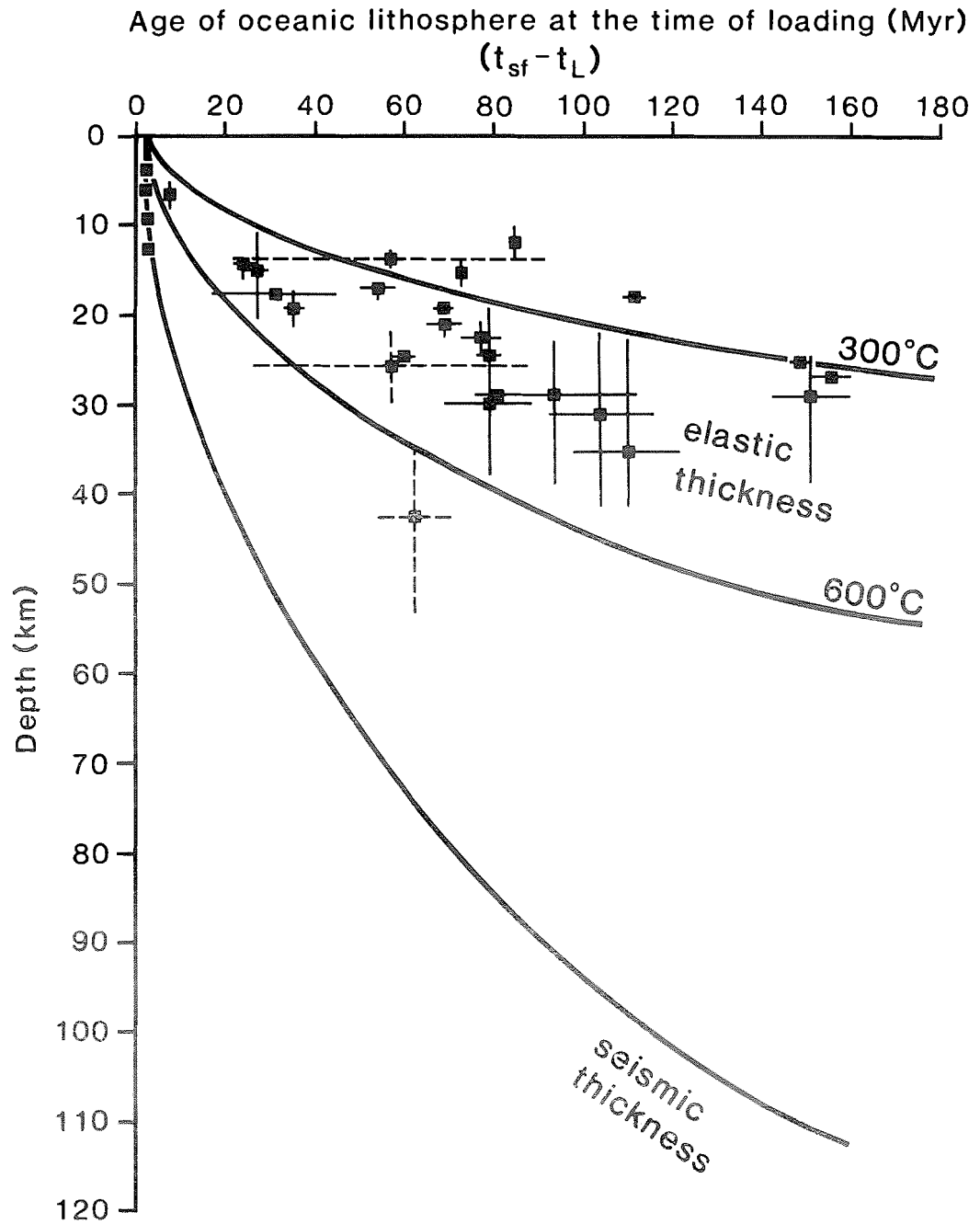


Figure 4. Elastic thickness of oceanic lithosphere, T_e , as a function of the age of lithosphere at time of loading t_{sf} = age of seafloor; t_L = age of load (after Watts and Ribe 1984). 300° and 600° isotherms from cooling plate model of Parsons and Sclater (1977).

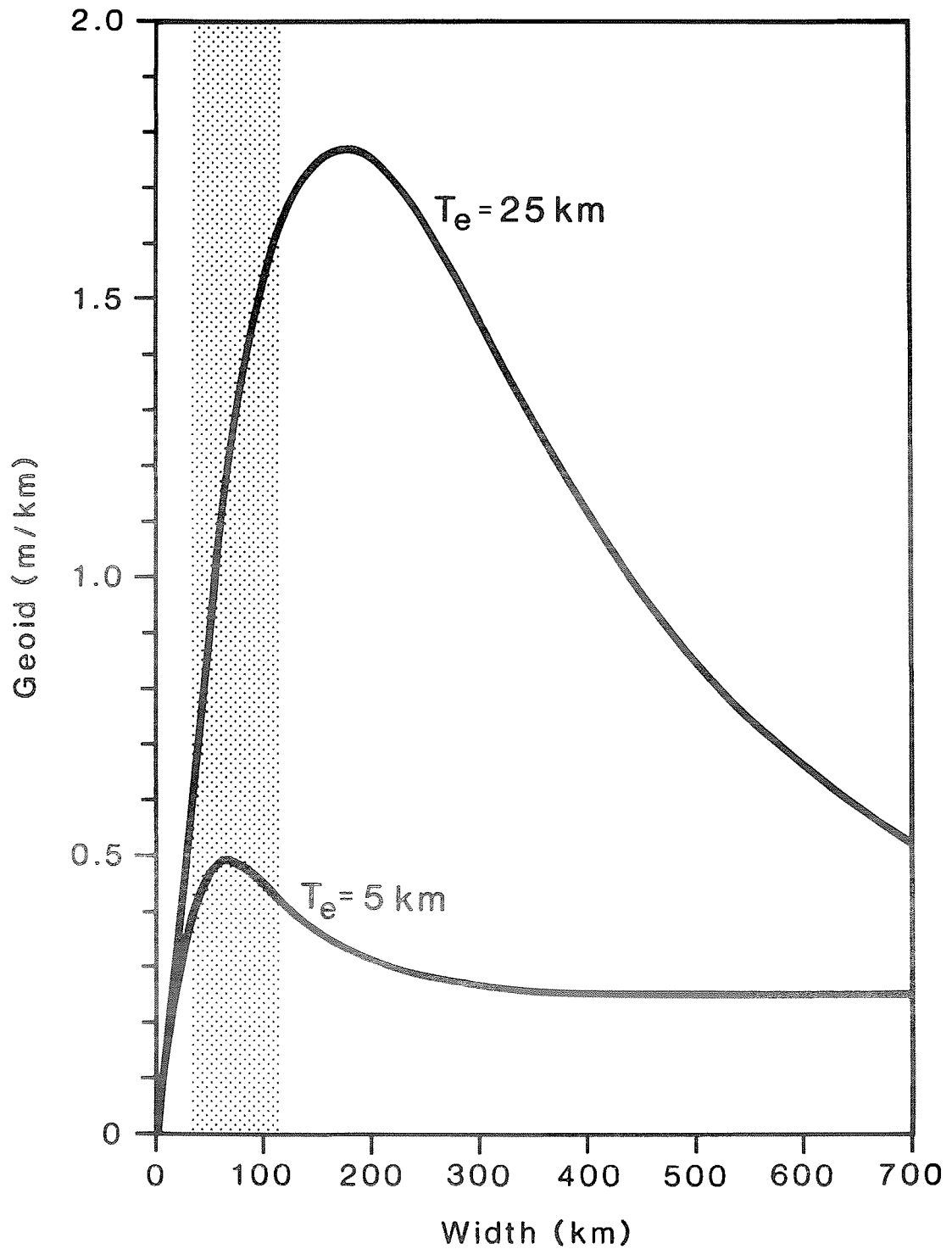


Figure 5. Maximum amplitude of geoid anomaly associated with two-dimensional seamount model. Upper curve for emplacement off-ridge ($T_e = 25 \text{ km}$), lower curve for on-ridge origin ($T_e = 5 \text{ km}$). Abscissa represents e^{-1} width of Gaussian seamount topography and vertical band indicates range of widths for typical oceanic seamount. Ordinate shows maximum geoid anomaly amplitude over feature 1 km high (after Watts and Ribe 1984).

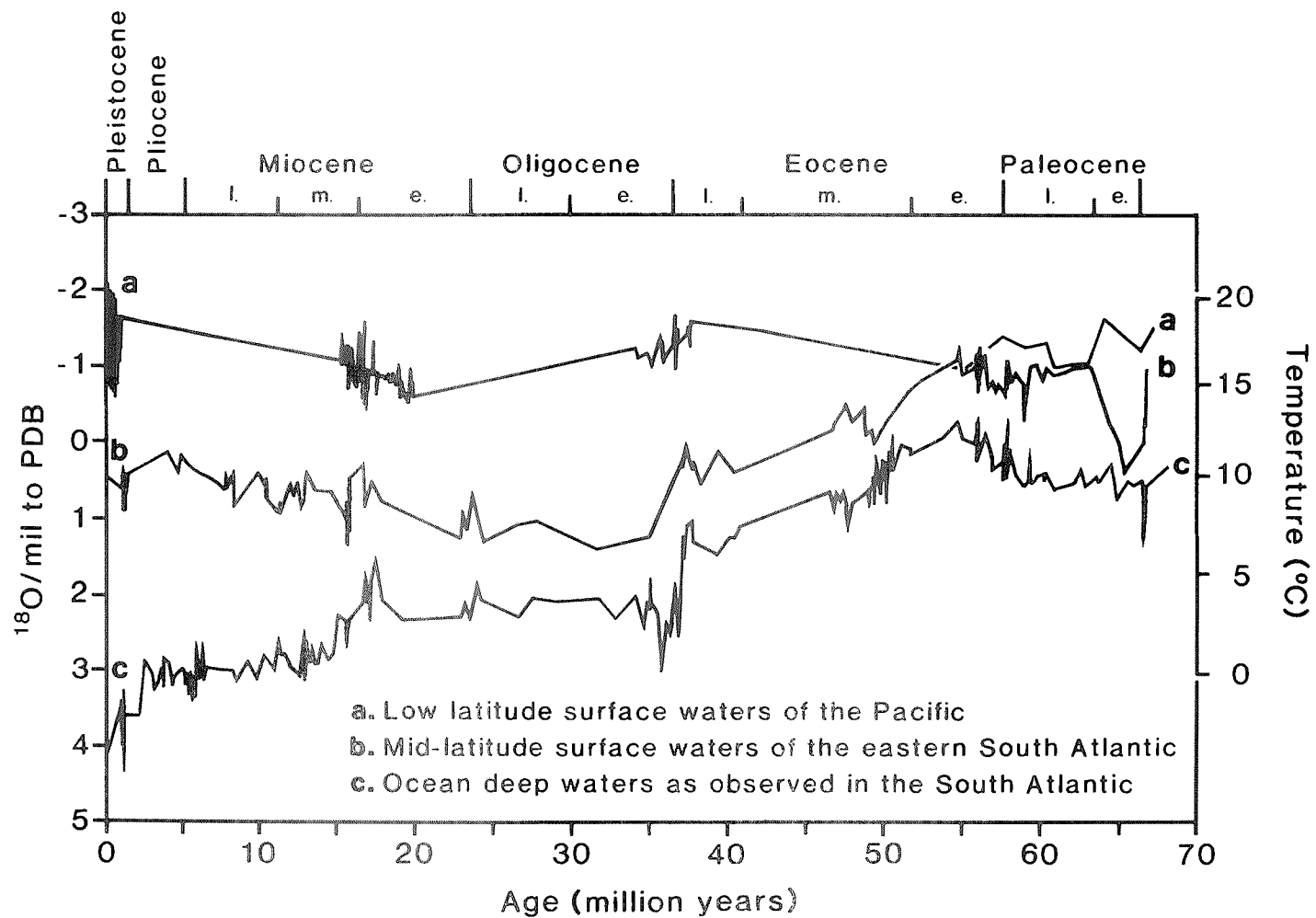
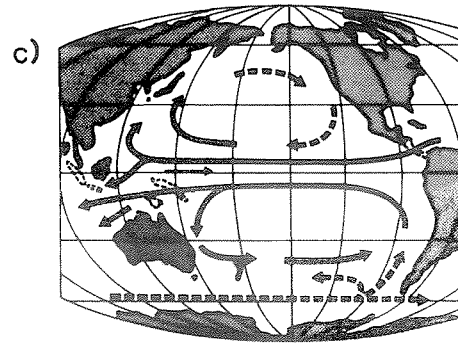
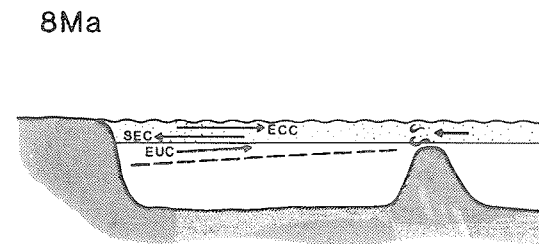
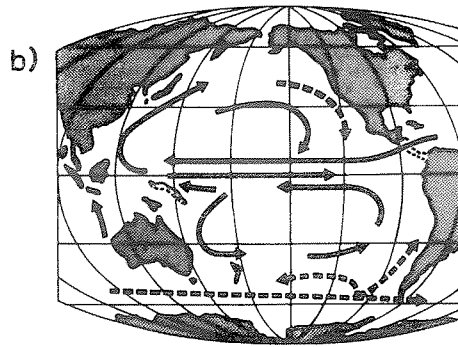
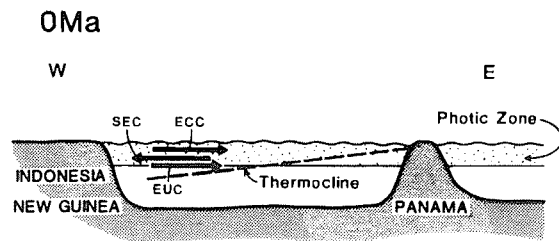
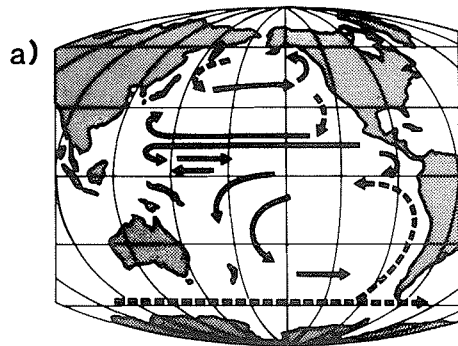
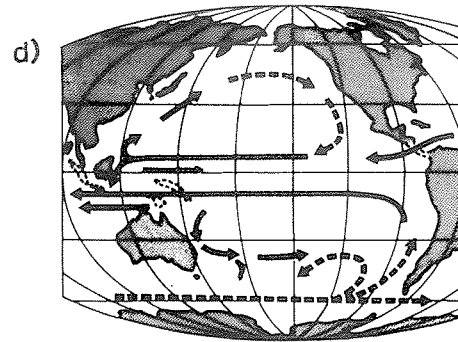


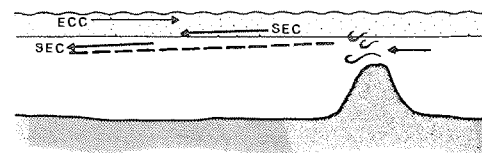
Figure 6. Oxygen isotope records for the last 70 million years (after Shackleton 1984). Temperature scale applicable only in the absence of Antarctic ice sheet before Middle Miocene; presence of Antarctic ice prior to this period would yield temperatures slightly higher than indicated on the temperature scale.



16Ma



22Ma



SEE PRECEDING PAGE

Figure 7. Inferred circulation patterns of surface and near surface waters in the Pacific Ocean at 22, 26 and 8 million years B.P. and suggested changes in surface water-mass structure in the equatorial Pacific, early Miocene to present (after Kennett et al. 1985).

On maps: broken arrows indicate cold currents; solid arrows indicate warm currents.

On sections: SEC South Equatorial Current; ECC Equatorial Countercurrent; EUC Equatorial Undercurrent.

22-16 million years B.P.: Both Indonesian and C. American seaways open to surface waters; thermocline relatively deep in absence of undercurrent; ECC weak.

8 million years B.P.: Indonesian seaway closed and formation of EUC; thermocline shallower; ECC moderately strong.

Modern ocean: C.American seaway closed; all surface water circulation more vigorous; further raising of thermocline into photic zone, particularly in E. Pacific.

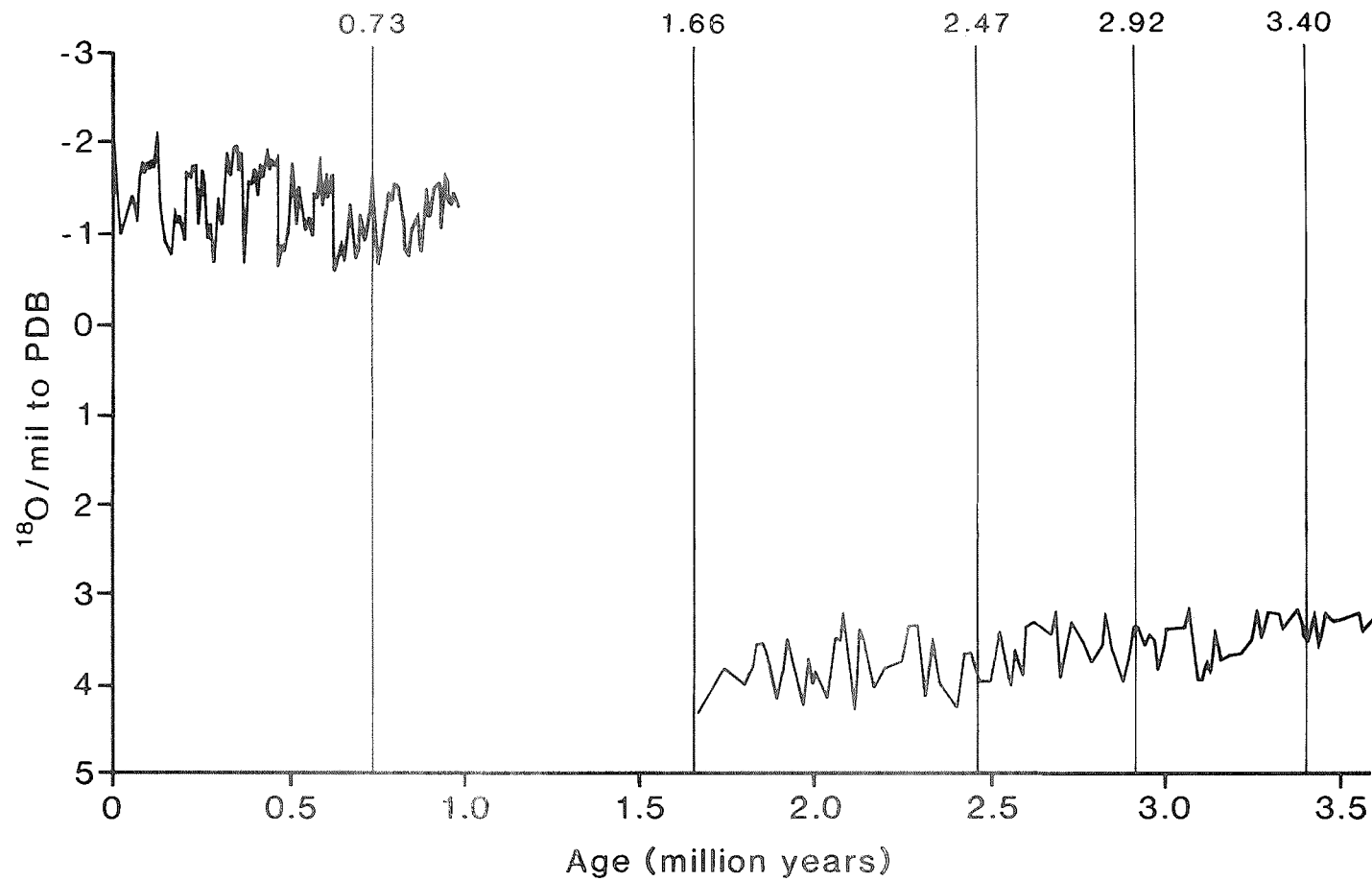


Figure 8. Oxygen isotope records from the Pacific Ocean. Top left: Pacific deep-sea core V28-238 (after Shackleton and Opdyke 1973). Bottom right; core V28-179 (after Shackleton et al. 1984a). Time control horizons: Brunhes normal chron (0.73 Myr), Olduvai normal subchron (1.66 Myr), Matuyama reversed chron (2.47 Myr), Kaena reversed subchron (2.92 Myr) and Gauss chron (3.40 Myr).

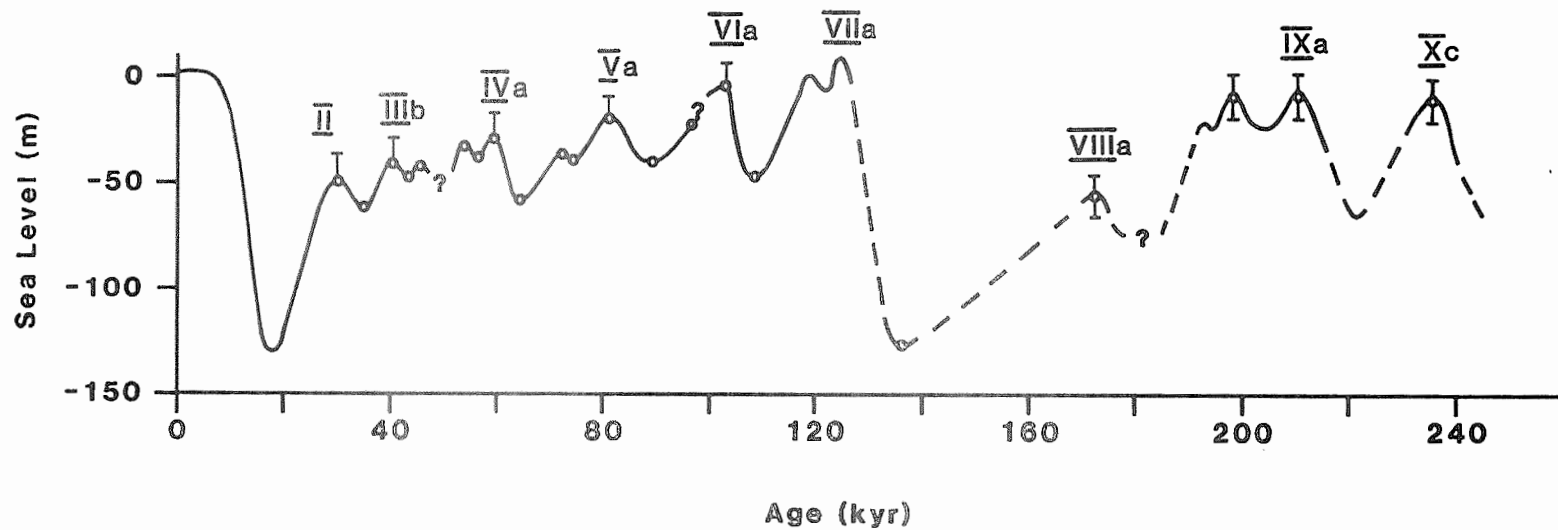


Figure 9. Latest sea-level curve for Huon Peninsula, New Guinea, with re-calculations after detailed correlation with ^{18}O record of Pacific core V19-30 (after Chappell and Shackleton 1986).

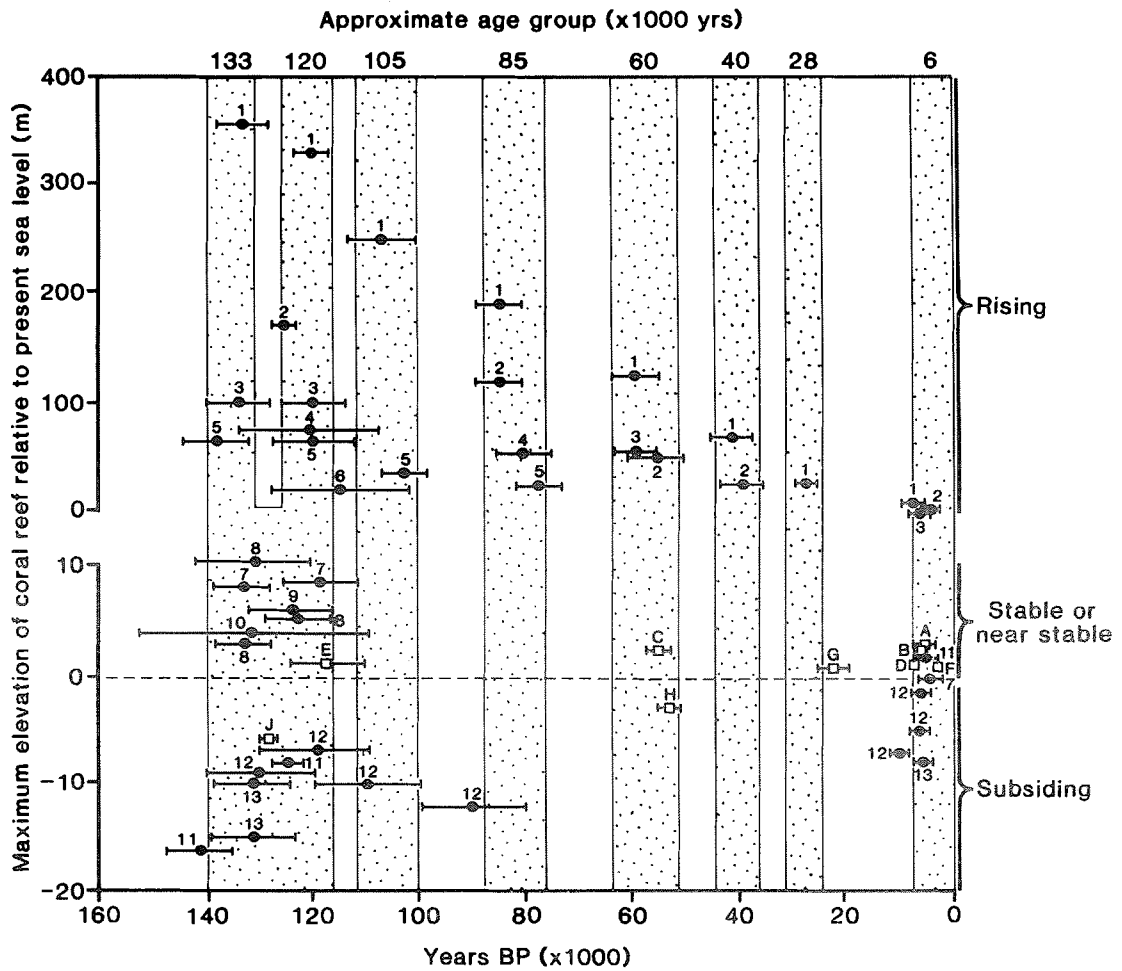


Figure 10. Chronology of late Quaternary coral reef occurrences (closed circles; after Aharon and Chappell 1986) and island phosphates (open squares; after Roe and Burnett 1985) from the Indo-Pacific biogeographical province plotted against the present position relative to mean sea level. Bars represent 1 S.D error of multiple dates.

Coral reef database: 1. Huon/New Guinea; 2. Ryukyu/Minami Daito; 3. Efate/Vanuatu; 4. Torres Island/Vanuatu; 5. Atauro/Timor; 7. Oahu/Hawaii; 8. Rottneest/W. Australia; 9. Aldabra; 10. Tuamotu Archipelago; 11. Great Barrier Reef; 12. Mururoa; 13. Enewetak.

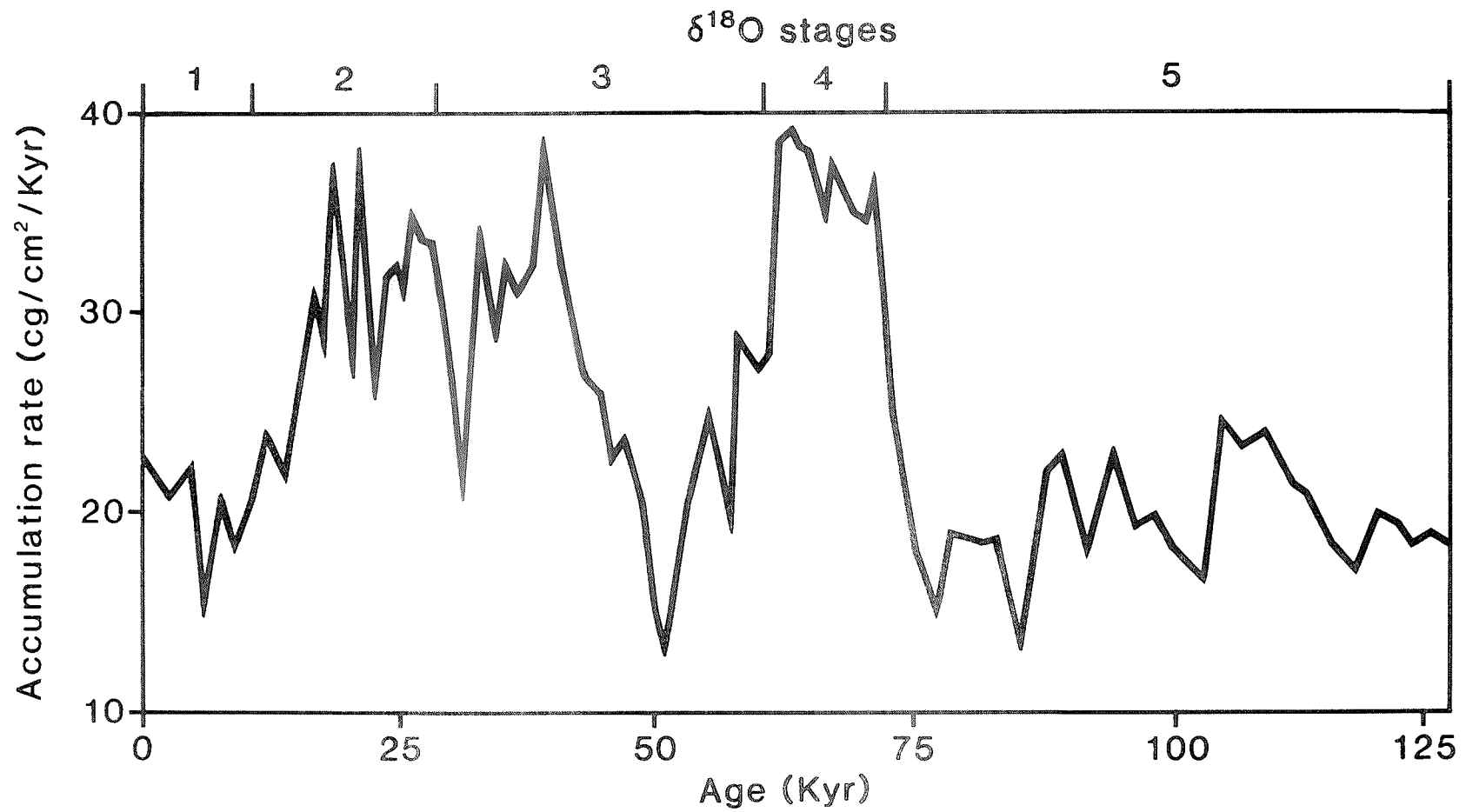


Figure 11. Downcore record of quartz accumulation in Pacific core V19-29 (3°35'S 83°56'W). (After Molina-Cruz 1977 and Romine and Moore 1981).

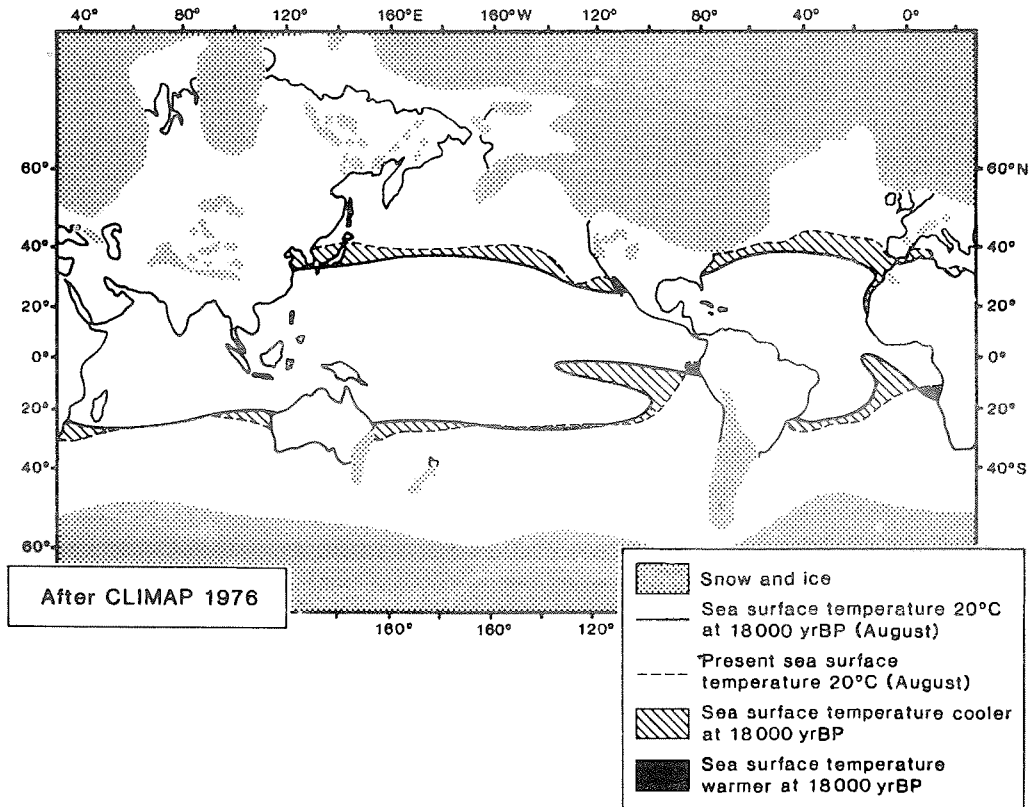
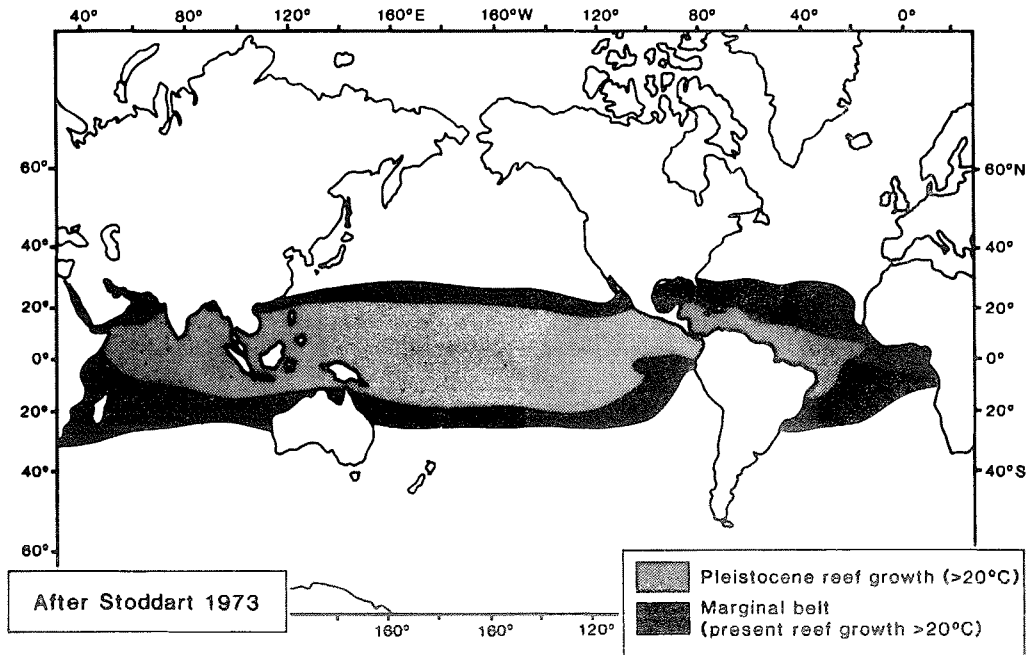


Figure 12. Possible changes in the extent of the reef seas, 18×10^3 yr B.P. v. present. In each case the isotherm of 20°C is taken as the effective limit of reef formation (Stoddart 1973. CLIMAP Project Members 1976).

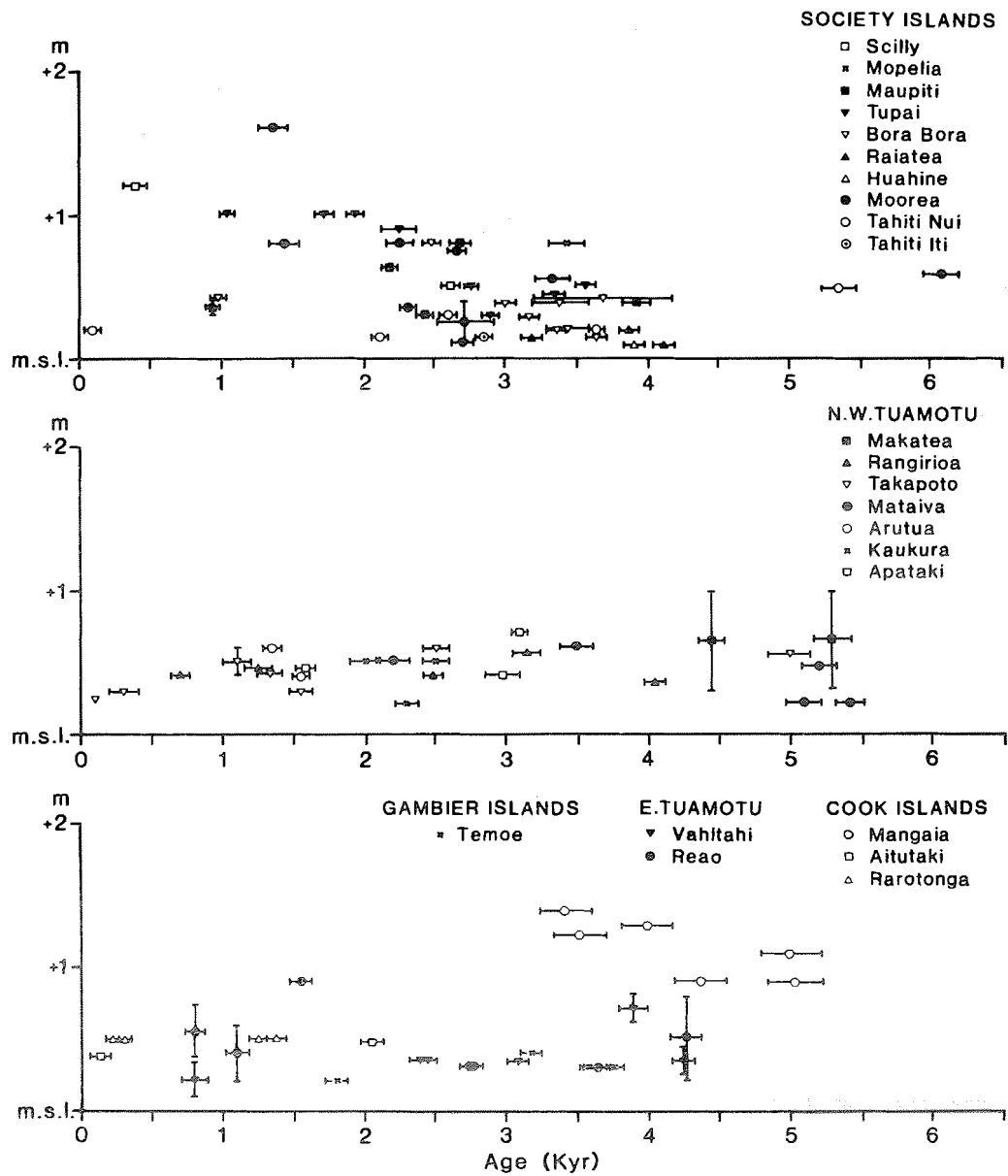


Figure 13. Time - elevation plot of sea levels in Polynesian archipelagoes, 6×10^3 yr B.P. to present.

Society Islands : Delibrias and Pirazzoli 1985; N.W. Tuamotu Pirazzoli 1985, Montaggioni et al. 1985, Pirazzoli and Montaggioni 1986; Gambier Islands: Pirazzoli 1987a, 1987b; E. Tuamotu: Pirazzoli et al. 1987a, 1987b; Cook Islands: Schofield 1970, Yonekura et al. 1984, 1985. Austral Islands: Pirazzoli 1987b.

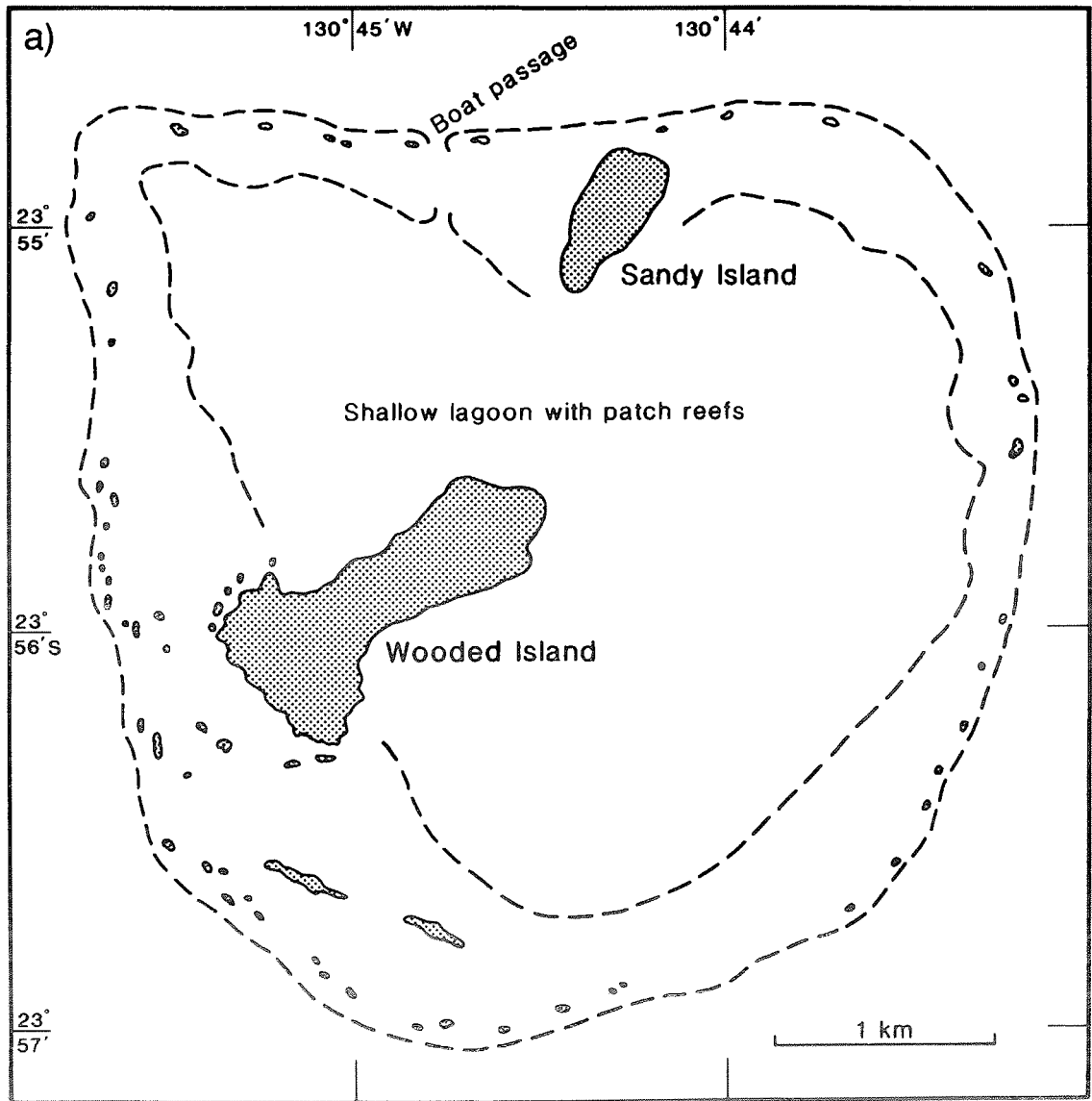
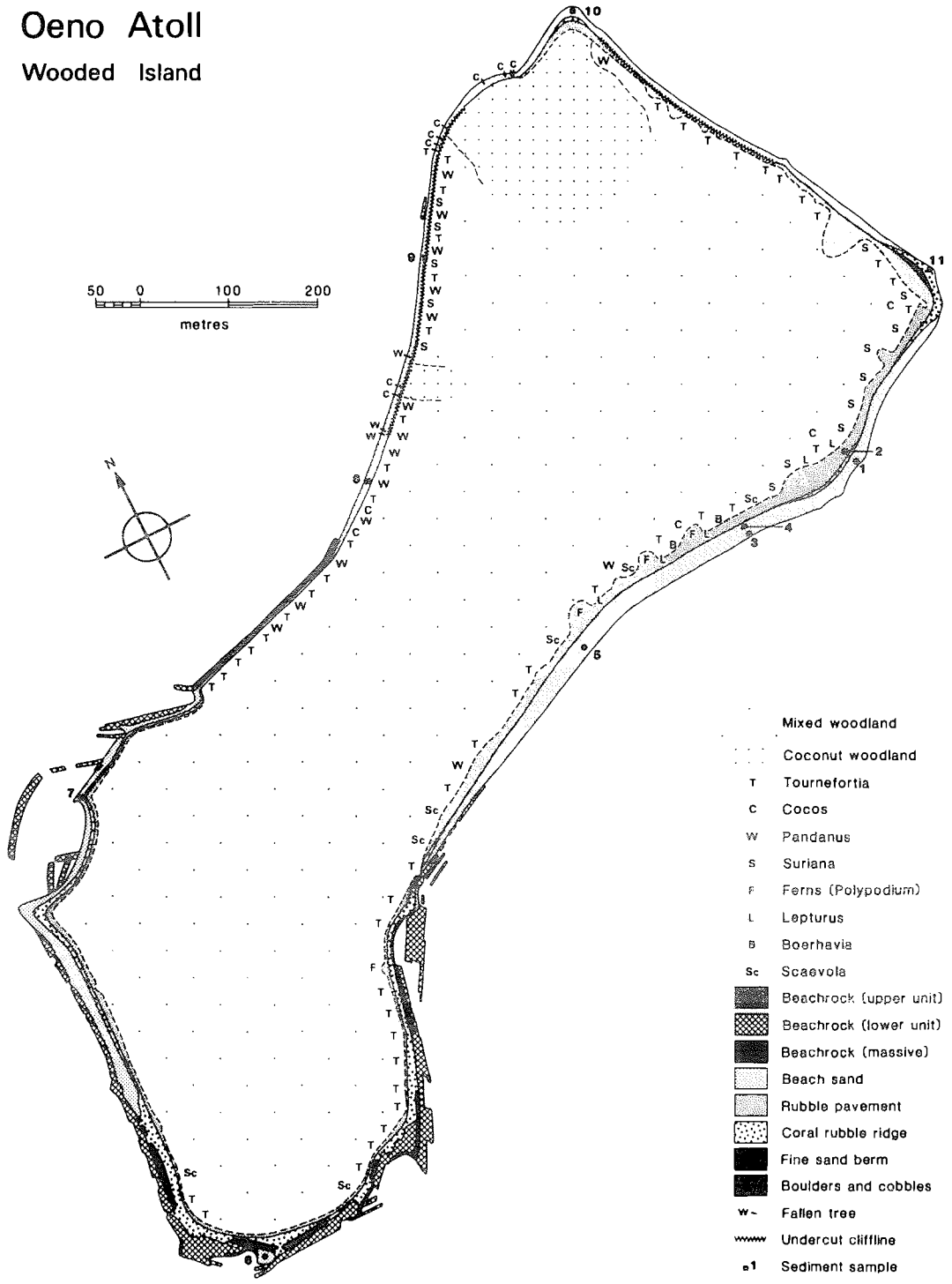
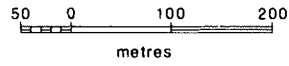


Figure 14. Geomorphology of Oeno Atoll, Pitcairn Islands. a) General plan form (after Admiralty Chart) b) Geomorphological map of wooded island. (see next page)

Oeno Atoll

Wooded Island



- Mixed woodland
- Coconut woodland
- T Tournefortia
- c Cocos
- w Pandanus
- s Suriana
- f Ferns (Polypodium)
- L Lepturus
- B Boerhavia
- Sc Scaevola
- Beachrock (upper unit)
- Beachrock (lower unit)
- Beachrock (massive)
- Beach sand
- Rubble pavement
- Coral rubble ridge
- Fine sand berm
- Boulders and cobbles
- w~ Fallen tree
- ~~~~~ Undercut cliffline
- a1 Sediment sample

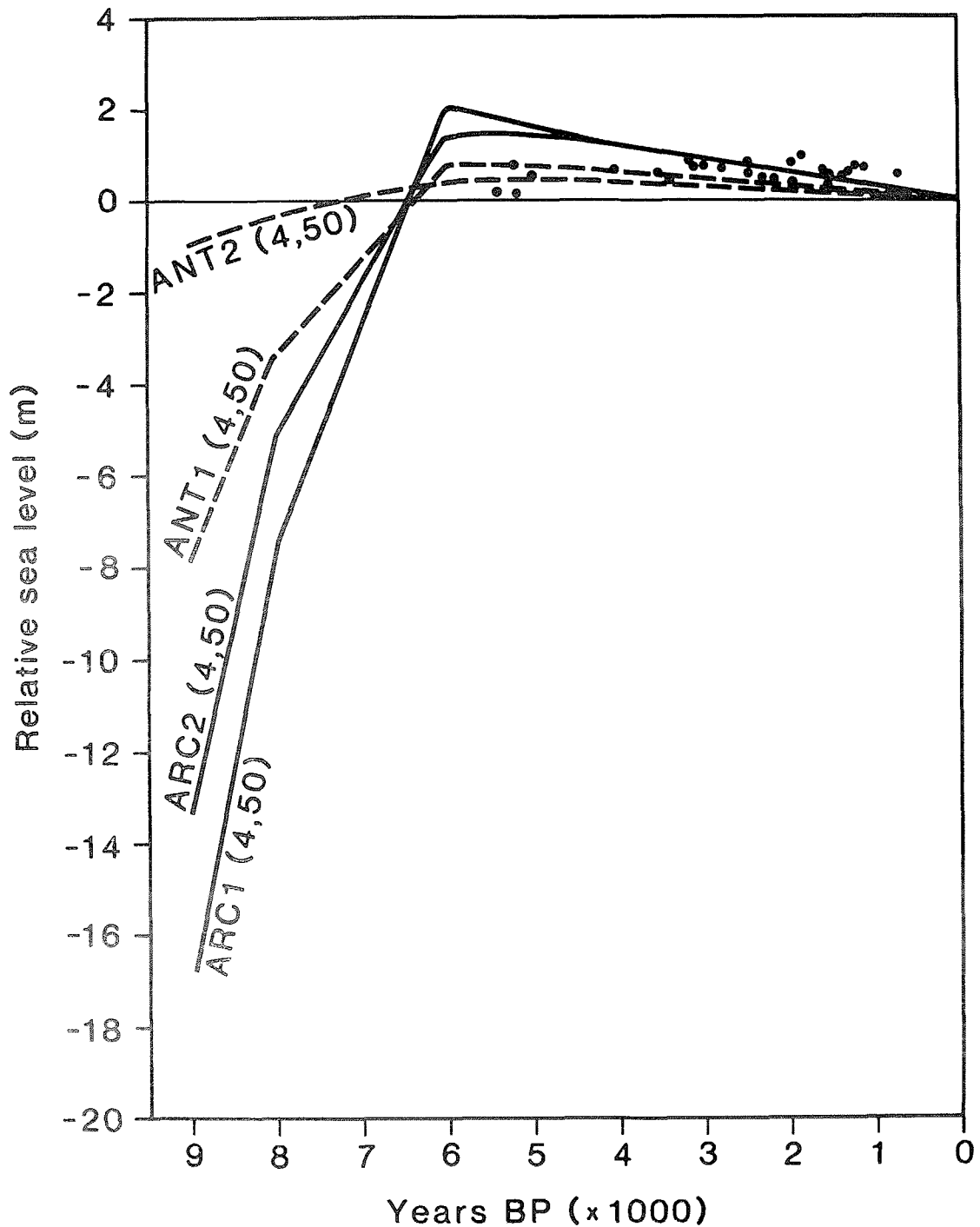


Figure 15. Relative sea-level models for the 'far field' as a function of ice load, mantle viscosity and lithospheric thickness (after Nakada and Lambeck 1987). Ice-melt models: ARC1,2: Arctic ice only; ANT1,2: Arctic and Antarctic ice. (4, 50): Upper mantle viscosity = 10^{21} Pa, lower (>670 km) mantle viscosity = 10^{23} Pa; 50 km thick lithosphere. Observed sea levels (closed circles) = N.W. Tuamotu archipelago (Pirazzoli and Montaggioni 1986).

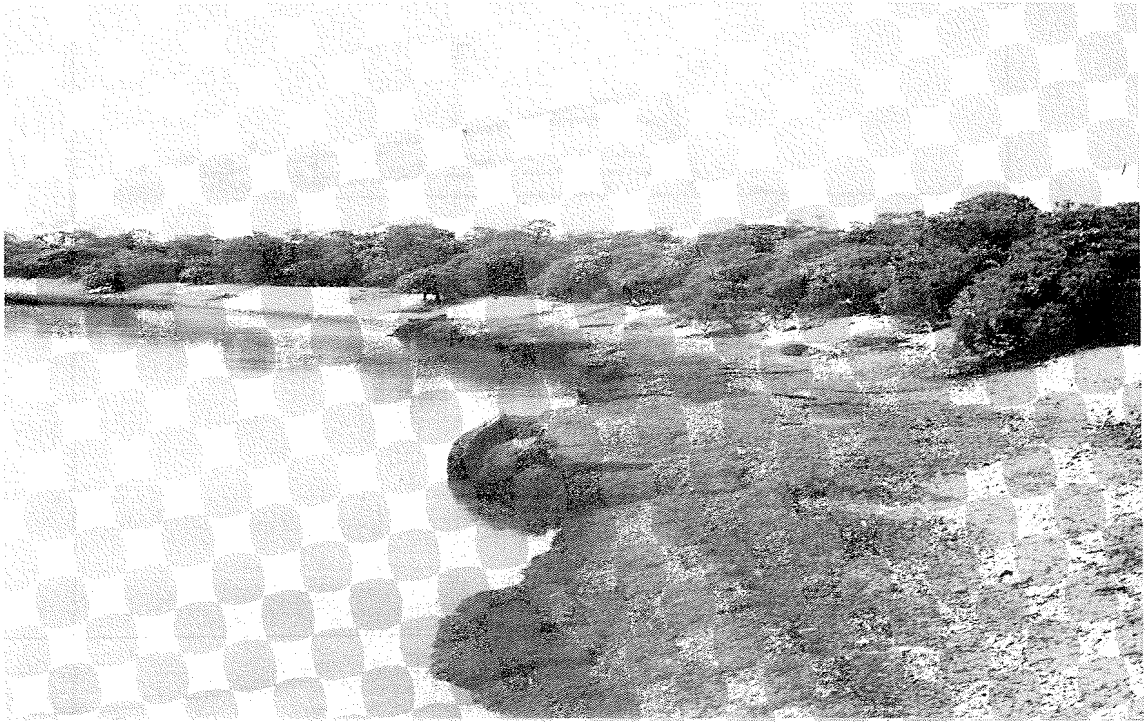


Plate 1. Cemented washover fans of coral stick rubble, eastern end of lagoon shore, Acadia Island, Ducie atoll.

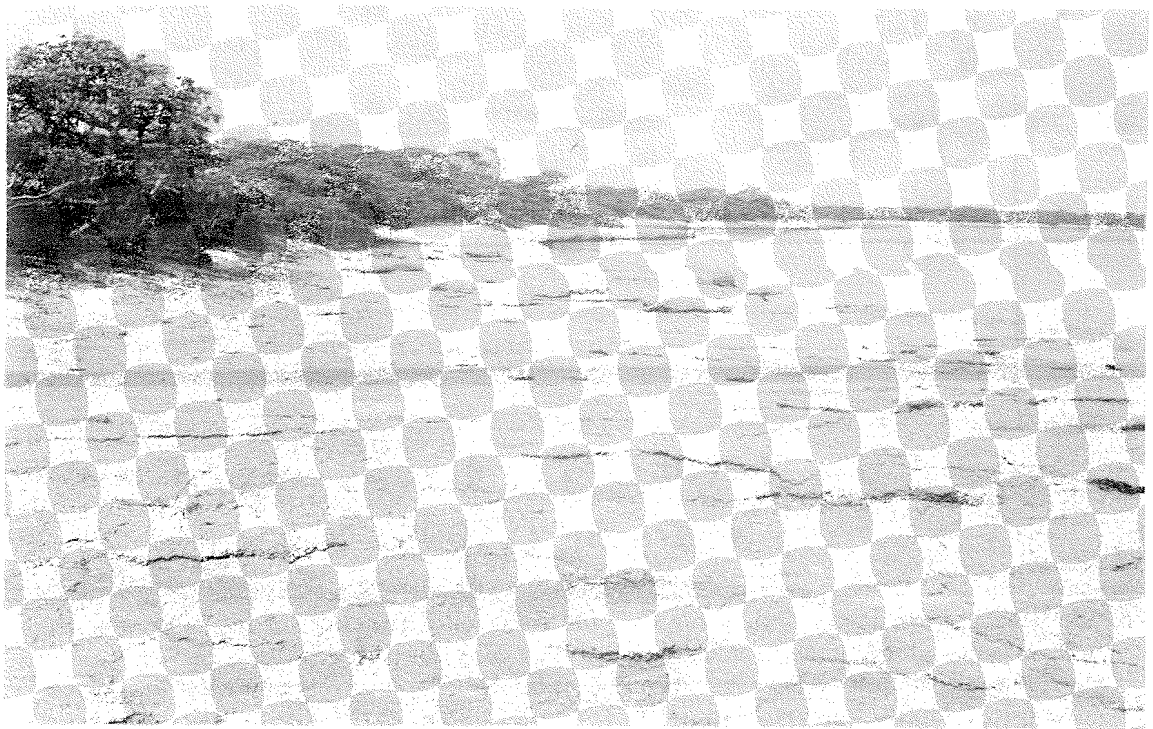


Plate 2. Pitted sheets of lagoon conglomerate spreads, central section of lagoon shore, Acadia Island, Ducie atoll. Lagoon margin at +0.30 - +0.45 m above mean sea level.

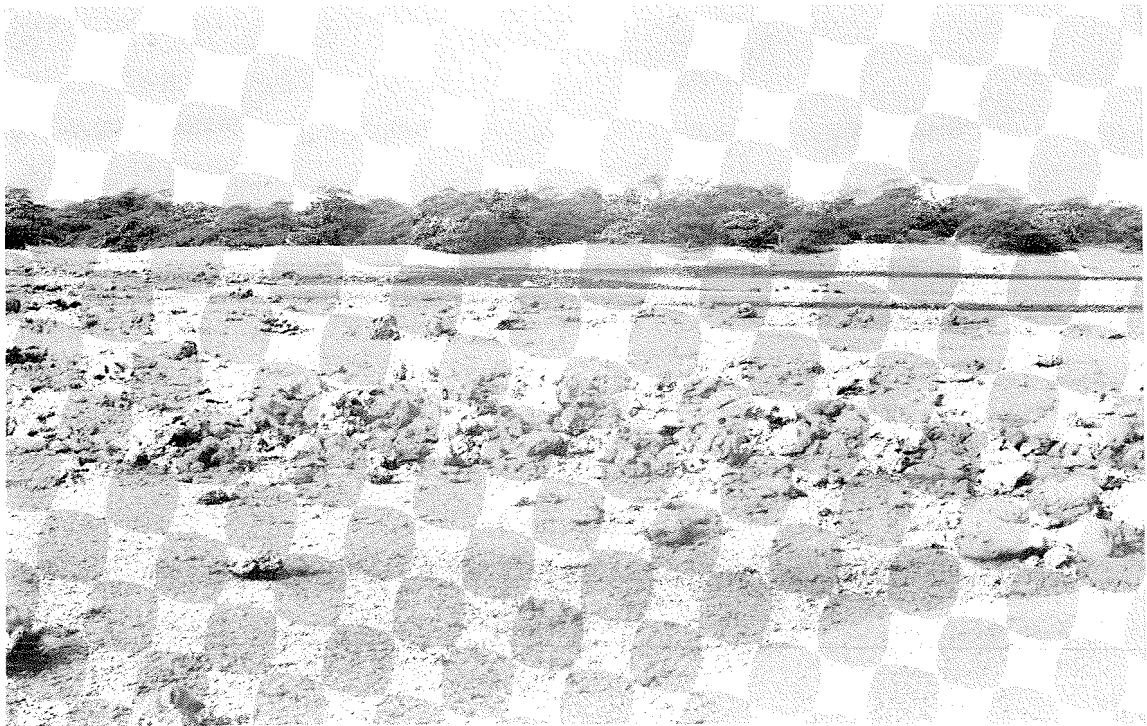


Plate 3. Boulder streams of coral heads with intervening flats of coral sticks, western end of lagoon shore, Acadia Island, Ducie atoll.



Plate 4. Lower bevelled beachrock unit overlain by more massive upper beachrock unit at S.W. point of wooded island, Oeno atoll.