Colloquium
Climatic Change and Geomorphology in tropical environments
Brussels, 6 May 1992
Royal Académy of Overseas Sciences
pp. 49 - 99 (1994)

Geo-Eco-Trop 16, (1 - 4), 49 - 99

GEODYNAMICS, OCEAN DYNAMICS AND ISLAND GEOMORPHOLOGY ON THE PACIFIC PLATE

Géodynamique, dynamique océanique et géomorphologie des îles sur la plaque du Pacifique

SPENCER, T.*

RÉSUMÉ

La reconstitution de la géodynamique régionale et des changements environnementaux sur la plaque du Pacifique peut être réalisée grâce à une meilleure connaissance à la fois de la géologie, de l'hydrologie et de la climatologie des océans assortie d'une interprétation nouvelle des stratigraphies, des topographies et des dépôts superficiels qui leur sont associés. A l'encontre des vues traditionnelles, une synthèse moderne montre que l'Océan Pacifique est géologiquement jeune, dynamique du point de vue de la géophysique et comprend des régions susceptibles d'enregistrer avec une certaine sensibilité des changements à grande échelle dans l'environnement. L'histoire géologique de l'Océan Pacifique depuis le début du Paléogene est l'évolution d'une grande région composée de plusieurs plaques vers une autre dominée par une seule plaque - celle du Pacifique - où la paléoocéanographie caractérisée par une zone océanique circum-équatoriale a été remplacée par un système circum-antarctique avec la mise en place des courants océaniques actuels. Dans les mers des bassins modernes, la succession darwinienne des types de récifs coralliens peut être expliquée par des processus à l'échelle du bassin océanique, tels que le refroidissement d'une plaque et la subsidence des îles à long terme. Toutefois, dans certains archipels et dans certaines chaînes montagneuses, la juxtaposition d'îles d'âge et de morphologie très différents fait ressortir l'importance des processus néo-tectoniques associés au nouveau réchauffement de la plaque et à la surcharge due aux apports volcaniques. Dans des cas semblables, la géomorphologie peut jouer un rôle important dans l'identification des zones de surélévation ou de subsidence. Toutefois, les effets de la tectonique demandent à être isolés de ceux des changements généralisés du niveau marin. La preuve de l'existence de hauts niveaux marins a été obtenue grâce à la datation d'une succession de récifs coralliens sur des îles en surrection tandis qu'une chronologie des positions basses du niveau de la mer a été établie grâce aux isotopes de l'oxygène, du carbone et du strontium de certains prélèvements effectués sur les plates-formes calcaires du centre de l'océan. Grâce à cette méthodologie, les variations du niveau de l'océan ont pu être établies pour le Tertiaire, le Pléistocène et l'Holocène. Dans la reconstitution des différents environnements, il a fallu prendre garde aux conséquences des modifications de la dynamique océanique et notamment des renversements périodiques de l'ENSO.

* Department of Geography, University of Cambridge, Downing Place, Cambridge CB2 3EN, U.K.

ABSTRACT

The reconstruction of regional geodynamics and environmental change on the Pacific plate is possible from the combination of improved knowledge of ocean basin geology, oceanography and climatology with re-interpretations of island stratigraphies, topographies and superficial deposits. Contrary to traditional views, a modern synthesis shows that the Pacific Ocean is geologically young, geophysically dynamic and contains regions which sensitively record large-scale environmental change. The geological history of the Pacific Ocean since the early Palaeogene is one of evolution from a large, multi-plate region to an area dominated by one plate, the Pacific plate, and where a Tertiary palaeooceanography dominated by a circum-equatorial seaway has been replaced by a circum-Antarctic ocean circulation system, establishing modern ocean current patterns. In the reef seas of the modern basin, the Darwinian sequence of coral reef types can be explained by ocean-wide processes of plate cooling and long-term island subsidence. However, in some island groups and chains, the inter-mixing of islands of widely differing ages and island morphologies points to the importance of neo-tectonic processes associated with plate re-heating and lithospheric loading by young volcanoes. In such circumstances, geomorphology provides an important role in establishing accurate patterns of regional uplift and subsidence. However, these tectonic signals need to be disentangled from the record of sea-level change. High sea-level stand evidence has been obtained from precise radiometric dating of coral reef terrace sequences on uplifting islands whilst the sea-level low stand chronology has been derived from the oxygen, carbon and strontium isotope record from mid-oceanic carbonate platforms. Using these methodologies sea-levels can be reconstructed for Tertiary, Pleistocene and Holocene timescales. Environmental reconstructions also need to be aware, however, of the role of ocean dynamical processes, and their quasi-periodic reversal in ENSO events, in generating additional signals in the record of environmental change.

INTRODUCTION

Improved knowledge of ocean basin geologies - through recovery and analysis of deep sea cores and the application of sea surface altimetry and seismic technologies - allied to modern plate tectonic theory - has revolutionised the way the scientific community sees the ocean basins. Contrary to traditional views which emphasised stability and antiquity, a contemporary synthesis shows that the ocean floors and their islands are geologically young and geophysically dynamic. Furthermore, recent studies which have shown rapid and largescale excursions in atmospheric carbon dioxide across interglacial/glacial boundaries have focussed attention on the role of the oceans in regulating CO_2 levels and have generated an interest in defining and modelling three-dimensional ocean dynamics rather than reliance on static maps of average circulation characteristics.

Nevertheless in some quarters the view persists that on shorter timescales the mid- and low-latitude oceans have been relatively unaffected by the large-scale climatic and environmental changes that have affected terrestrial and high-latitude marine environments over the last three million years and that they can be used as baselines against which to measure change elsewhere. Such a view is still prevalent in the area of sealevel studies where tropical islands have often been seen as stable "dipsticks" recording true eustatic variations in sealevel which can subsequently be subtracted from more complex records of sealevel elsewhere.

This paper, in choosing sealevel as a key environmental variable, attempts to draw attention to i) the potential of new technologies for the improved reconstruction of the record of environmental change and its extension back from the Pleistocene;

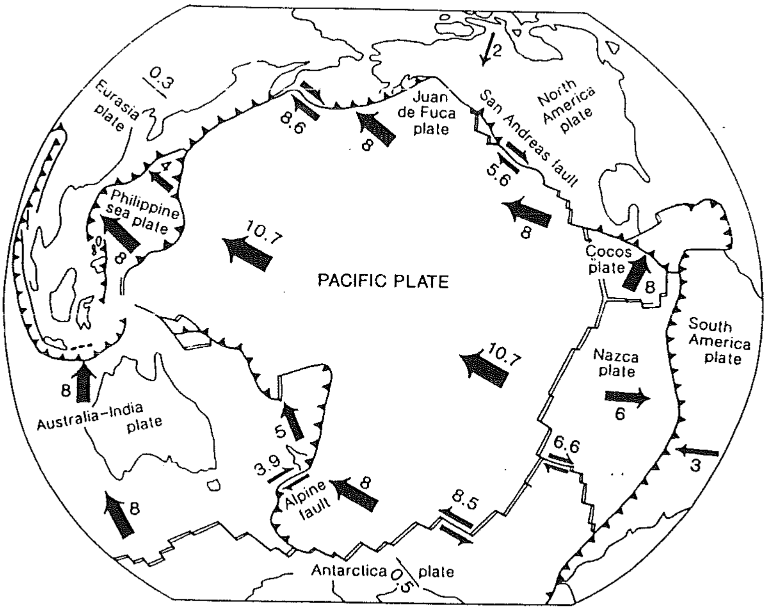

Fig. 1 : The Pacific plate and neighbouring plates of the Circum-Pacific region (modified from OTA & KAIZUKA, 1991). Spreading centres : double lines; transform faults : single lines; barbed lines : subduction zones, with barbs in direction of subducting plate. Long arrows show absolute rate of plate motion in cm a^{-1}. Half arrows show relative direction and rates of motion at transform boundaries.

ii) the need for, and difficulties of, isolating the climatically- and tectonically-driven elements of the record of environmental change and iii) the role of geomorphology, in re-interpreting island stratigraphies, topographies and superficial deposits, in aiding this process of improved environmental reconstruction. Its focus is that part of the Pacific Ocean associated with the Pacific plate and its boundaries (Fig. 1), the greater part of an ocean basin that covers 35 % (166 M. km^2) of the Earth's surface.

ESTABLISHMENT OF THE MODERN PACIFIC OCEAN : PLATE GEOMETRY AND OCEAN-ATMOSPHERE CIRCULATION SYSTEM

The Pacific Ocean has been subjected to major changes in plate geometry over the Cenozoic. Such changes have had major implications for global climates as a result of the re-organization of regional oceanography from a circum-equatorial ocean-atmosphere circulation system to one dominated by circum-polar processes and a strongly meridional circulation pattern (KENNETT, 1982).

At the beginning of the Palaeogene (65 Ma BP), 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). 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 became the dominant plate, exhibiting varying rates and directions of large-scale motion with changing boundary constraints. Plate dynamics can be reconstructed from the hot-spot traces of intra-plate islands and seamounts. The best example of such a trace is the Hawaiian Islands - Emperor Seamounts island chain which stretches nearly 6000 km across the North Pacific Ocean and comprises at least 107 individual volcanic centres. The chain is age-progressive with the most southerly islands actively forming and the most northerly dated at 70 - 85 Ma; the bend in the chain between the emergent islands and reef shoals and the older seamounts has been dated at 43.1 ± 1.4 Ma and probably indicates major plate reorganization consequent upon the collision of India with Eurasia (CLAGUE & DALRYMPLE, 1987), most notably by the "bend" in the Hawaiian Islands - Emperor Seamounts chain. In the South Pacific, the collision of the Pacific-Kula-Farallon boundary with the Farallon-Americas trench at ~ 26 Ma BP stopped all seafloor spreading and subduction in this region and initiated the direct coupling between the Pacific and Americas plate. This fusion created a series of accommodations over next 20 Ma, including both clockwise rotation of the southern

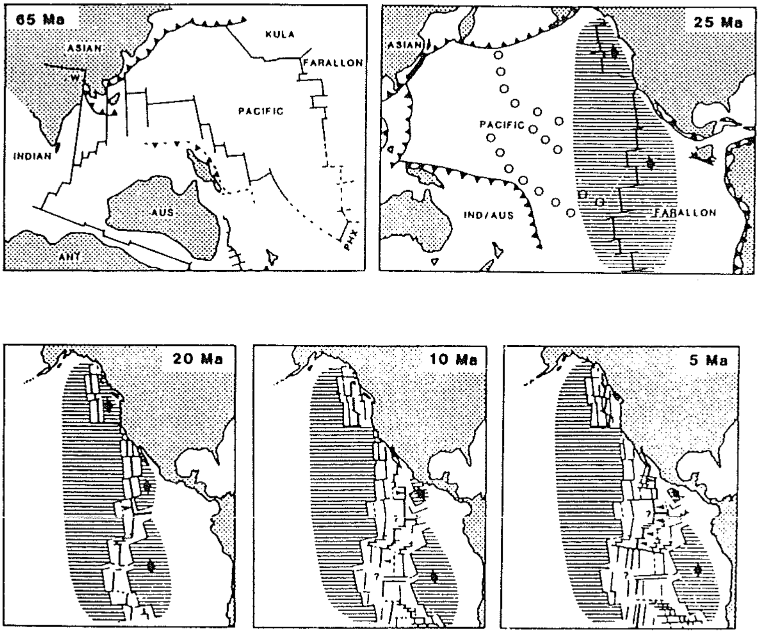

Fig. 2 : The Pacific Ocean at the beginning (65 Ma BP) and end 25 Ma BP) of the Palaeogene and subsequent plate re-organisations at 20, 10 and 5 Ma BP (after SPENCER, 1989). Open circles indicates island chains of subaerial volcanoes and seamounts.

portion of the Pacific-Farallon ridge, development of the Galapagos rift and episodic, eastward mid-ocean ridge 'jumps' in the South Pacific (Fig. 2; SPENCER, 1989).

Oxygen isotope data (SHACKLETON, 1984) suggests that the early Cenozoic oceans were warm and possessed relatively small equator-to-pole and surface-to-bottom temperature gradients. It seems likely that such conditions were a continuation of the warm, high sea-level state of the Cretaceous oceans. Numerical ocean circulation models indicate that the palaeo-circulation system during the mid-Cretaceous (~ 100 Ma) was dominated by strong westerly flows near the equator (BARRON & PETERSON, 1990, 1991; FÖLLMI & DELAMETTE, 1991). Interestingly, reef corals dredged from the Mid-Pacific Mountains, which were near the equator 91-119 Ma BP, and Upper Cretaceous (84-65 Ma) shallow water foraminifera recovered from the Line and Marshall Islands, which would have been south of the equator but in tropical waters at this time, all show, as expected by this reconstruction, zoogeographical affinity with the

Caribbean portion of the former Tethyan seaway (GRIGG, 1988); this phase of increased west-east dispersal appears to have come to an end after the Eocene (GRIGG & HEY, 1992).

Dramatic cooling of both mid-latitude and deep-ocean waters occurred at the Eocene-Oligocene boundary with further cooling downturns in the middle Miocene (SHACKLETON & KENNETT, 1977; SAVIN *et al.*, 1981; WOODRUFF *et al.*, 1981; KEIGWIN & KELLER, 1984). This trend appears due to palaeogeographic changes in the Pacific Ocean basin, the result of the re-arrangement of plate boundaries, completely altering patterns of ocean circulation. The Oligocene development of the Circum-Antarctic Current, as Australia (with the subsidence of the Tasman Rise) and later, South America (through the opening of the Drake Passage), 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). Furthermore, the associated formation of Antarctic bottom water, from ~ 36 Ma, effectively began the separation of surface and deep ocean water masses and created the modern thermohaline circulation system. Allied to the establishment of this high-latitude, circum-global circulation was the modification and ultimate demise 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 changes resulted in the establishment of stronger latitudinal temperature gradients and stronger wind and ocean currents. Climatically, evidence for increasingly vigorous atmospheric circulation from the early Miocene, at both 15-16 Ma BP and 9-5 Ma BP, associated with climatic deterioration, is provided by the appearance of diatomites in Pacific rim sedimentary sequences (INGLE, 1981) increasing biogenic silica accumulation (from 16 Ma BP, peak at 8 Ma BP : LEINEN, 1979) and rising calcium carbonate supply rates (peaking 14-15 Ma BP : 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 Ma BP imply a significant intensification of the southern (REA & BLOOMSTINE, 1986) and northern hemisphere tradewinds (REA & JANECEK, 1982). Within the ocean gyral circulation was strengthened, not only in magnitude by increased wind stress but also meridionally by the diversion of large volumes of water into westerly boundary currents. Reef corals first appear in seamount sediments from the Emperor seamounts chain ~ 34 Ma, replacing earlier carbonate accumulations composed of bryozoan algal facies; GRIGG (1988) hypothesizes that their arrival can be correlated with the increased transport of coral larvae from the diversity centre of the Indo-Pacific biogeographic region, centred over the Philippines, under a stronger circulation system.

PLATE TECTONIC PROCESSES AND ISLAND PATTERNS ON THE MODERN PACIFIC PLATE

As newly-formed lithosphere moves away from the spreading centre of the East Pacific Rise to form new Pacific Ocean seafloor, it progressively cools and thickens as upper asthenospheric material accretes to the base of the plate. Isostatic adjustments result in lithosphere subsidence and a deepening seafloor beneath sea-level; these relationships are consistent and can be modelled using a simple heat loss equation (e.g. PARSONS & SCLATER, 1977), at least until plate ages of ~ 70 Ma. Thus at the East Pacific Rise the average water depth is 2 780 m, on 30 Ma old lithosphere 4 350 m, and on 75 Ma old lithosphere 5 610 m (LE PICHON *et al.*, 1973). The Pacific Ocean therefore progressively deepens towards the area of the oldest lithosphere, the seafloor off the Kamchatka/Kurile subduction zone, taking volcanoes formed at the Rise into progresively deeper water through time. Typical subsidence rates, usefully summarised by PAULAY and McEDWARD (1990), have been calculated at 0.22 - 0.29 cm ka^{-1} at Enewetak Atoll, 0,14 - 0.16 cm ka^{-1} at Midway Atoll; and 0.8 - 1.25 cm ka^{-1} at younger Mururoa Atoll.

These thermal processes now provide a mechanism for DARWIN'S (1842) theory of the origin of coral reefs. Volcanic islands in the reef seas (where water temperature in the coldest month of the year > ~ 20° C) are likely to acquire a fringing reef of coral and coralline algae. With progressive subsidence of the central volcano but continued upward growth of coral to sea-level, fringing reef islands develop into barrier reef islands, possessing a lagoon between the island and the reef. Embayed volcanic shorelines and reef passages indicate the position of former subaerial valleys now drowned by subsidence (DAVIS, 1928). Ultimately, with the complete disappearance of the central volcano, an atoll is formed, exhibiting a ring of coral surrounding a single central lagoon. Darwin's map of the distribution of these island types shows broad global patterns : barrier reefs and atolls, indicative of subsidence, are characteristic of the ocean basins, whilst fringing reefs, representative of stable or uplifted coasts, are associated with the ocean margins. Superimposing the positions of modern plates and

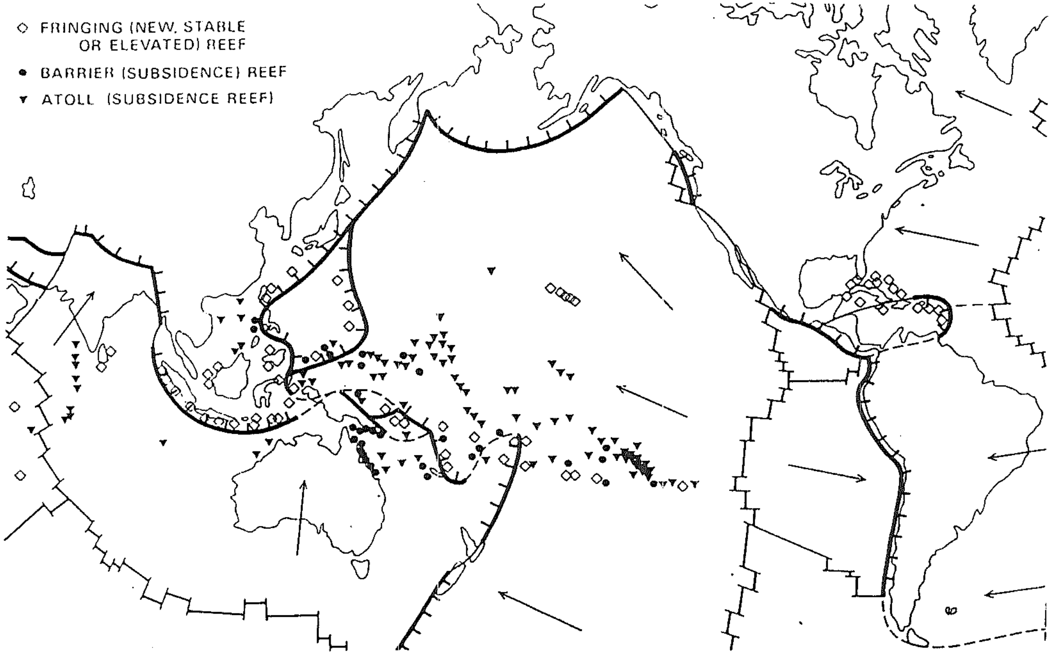

Fig. 3 : Darwinian reef types in relation to modern plate boundaries (after ROSEN, 1982).

their boundaries shows that barrier reefs and atolls are typically found at mid-plate setting whereas fringing reefs are more characteristic of plate margins (Fig. 3, ROSEN, 1982).

However, coral reefs are shallow water (< 50 m) constructions and the distribution of magma production along the East Pacific Rise generally results in ~ 2.5 km of water above the spreading ridge. Thus most coral islands are not associated with aseismic ridge genesis but with mid-plate sites of high magma production consequent upon plate re-heating at asthenospheric swells and fixed plume melting points, or "hotspots", within the asthenosphere (MENARD, 1973). Some of these areas are extensive : the South Pacific "superswell" for example, characterizes the region between ~ 10° S to ~ 30° S and from 120° to ~ 160° W and is associated with high levels of volcanic activity (CALMANT & CAZENAVE, 1986, 1987; FISCHER *et al.*, 1986). With continued lithospheric plate motion, and de-coupling from the asthenospheric heat sources which are fixed in position, the plate cooling-seafloor deepening relationshhip is re-established, although at accelerated rates of subsidence, as thermal rejuvenation imposes the characteristic behaviour of young lithosphere (CROUGH, 1978; DETRICK & CROUGH, 1978). Subsidence rates of 1.2 - 4.1 mm a^{-1} at the present time; 1 - 2 mm a^{-1} over 8,000 years; and 0.14 - 0.15 mm a^{-1} over 5 ka have been suggested for Hawaii (MOORE & FONARI, 1984); Oahu (NAKIBOGLU *et al.*, 1983) and Moorea-Tahiti (PIRAZZOLI & MONTAGGIONI, 1985) respectively.

Of particular importance to palaeo-environmental reconstructions, yet relatively neglected, are the implications of lateral plate motion. Current rates of Pacific plate motion have been established from the study of ocean floor sediments, from palaeomagnetic measurements and from the radiometric dating of exposed volcanics along various island chains. These studies indicate an average rate of plate motion of ~ 11 cm a^{-1} to the north-west (Tab. I). This rate is rapid in geological terms and considerably greater than the changes in vertical island position described above. The nature of reef growth (although not necessarily the growth potential) associated with this lateral movement is related to biogeographic diversity patterns of reef-building corals which are themselves determined by plate tectonics : there is a rapid decline in the number of reef-building coral genera east of Fiji and Tonga (COUDRAY & MONTAGGIONI, 1983) and, even more strikingly, an absence of reef-associated seagrass and mangrove communities (STODDART, 1992). Lateral migration is particularly significant for established coral reefs which are migrating out the reef seas.

Tab. I : Estimates of the rate of Pacific plate motion.

Rate of plate motion (cm a^{-1})	Location	Method	Source
10.7 ± 1.6	Austral -Cook Is.	K-Ar dating of exposed volcanics	DUNCAN & CLAGUE (1985)
11.4 ± 2.3	Austral -Cook Is.	K-Ar dating of exposed volcanics	BROUSSE (1985)
10.4 ± 1.8	Marquesas	K-Ar dating of exposed volcanics	DUNCAN & CLAGUE (1985)
10.9 ± 1.0	Society Is.	K-Ar dating of exposed volcanics	DUNCAN & CLAGUE (1985)
12.7 ± 15.5	Pitcairn-Gambier Is.	K-Ar dating of exposed volcanics	DUNCAN & CLAGUE (1985)
10.7 ± 11.0	Pitcairn-Gambier - Mururoa	K-Ar dating of exposed volcanics	BROUSSE (1985)
9.2 ± 0.3	Hawaiian Islands	K-Ar dating of exposed volcanics	CLAGUE and DALRYMPLE (1987)
7.2 ± 1.1	Emperor Seamounts	K-Ar dating of exposed volcanics	CLAGUE and DALRYMPLE (1987)

Progressively more inimical conditions for coral growth ultimately result in levels of reef accretion which cannot balance continued coral bioerosion and subsidence effects (discussed above). Here atolls drown to become submerged seamounts : GRIGG (1982) terms this threshold the "Darwin Point" and places its position at 29° N, near Kure Atoll at the northern end of the Hawaiian Archipelago. On the other hand, largely reef-less islands, like Pitcairn Island at 24° S, are moving towards more favourable water temperatures (STODDART, 1976).

Volcanic islands are respeatedly formed and then detached from mid-plate magma sources with continued plate motion. Thus Pacific islands chains, aligned over the last 40 Ma in the north-westerly direction of plate motion, should be characterized by i) progressive increases in volcanic basement ages and island denudation with distance from hotspot and ii) fringing reefs at their south-easterly ends and barrier reefs and then atolls at their north-western extremities. Although such patterns of islands are found in the reef seas, many Pacific chains show more complex mixes of island ages and topographies. In particular, the simple division of islands into high volcanic islands and low coral atolls can be complicated by the transformation of atolls, and other reef types, into a third category of island, high limestone or *makatea* islands, by intra-plate neotectonics. How can the record of environmental change be reconstructed from the geology and geomorphology of these different island types ?

RECONSTRUCTING THE RECORD OF ENVIRONMENTAL CHANGE

HIGH SEA-LEVEL STANDS : CORAL REEFS EXPOSED BY TECTONIC UPLIFT

Volcanic islands and coral reefs associated with the margins of the Pacific plate show complex patterns of uplift and subsidence. Trench-arc classification differentiates between two contrasting modes of subduction as the end members of a continuum of deformation types : compressional or high stress (Andean/Chilean) and extensional or low stress (Mariana) regimes. Compressional systems are characterized by volcanic arcs divided into ridges produced by folding and/or thrust faulting and fronted by a prominent fore-arc often supporting uplifted terraces and fronted itself by an accretionary prism of material derived from the subducting plate. The extensional type of trench-arc system does not have a massive volcanic or fore-arc and is characterized by a spreading back-arc basin. These relatively simple models are further complicated when oblique subduction occurs (OTA, 1991). Using age-altitude relationships for Last Interglacial and Holocene (6 ka) terraces around the Pacific margin, and controlling for glacio-eustatic sea-level fluctuations, OTA (1986) has calculated rates of uplift by plate boundary type. Uplift rates are highest on coasts adjacent to boundaries between continental and subducting oceanic plates (e.g. Santos and Malekula Is. Vanuatu; Huon Peninsula, New Guinea) : here uplift rates are generally in excess of 1 m ka^{-1} over the last 125 ka and over 1.5 m ka^{-1} since 6 ka. Maximum rates are 2.6 m ka^{-1} for the Huon Peninsula (CHAPELL, 1974) and 2.0 m ka^{-1} in Vanuatu (TAYLOR *et al.*, 1980). Holocene uplift rates on Vanuatu have reached 4.0 m ka^{-1} (JOUANNIC *et al.*, 1980).

It is staircases of raised reef terraces on the most rapidly uplift plate margins that have provided precise recods of sea-level change : the difference between the altitude of a reef of known age and present sea level provides a precise measure of past sea level if the tectonic component can be estimated and substracted. The general applicability of this methodology and the resulting good correlations between sea-level records from widely separated locations around the globe (e.g. CHAPPELL & POLACH, 1991) gives confidence that a true eustatic record of sea level change is being recorded. In the Pacific basin, 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 m ka^{-1}) coastline. The sea-level curve from this locality has been recently refined, following comparison with the deep-sea core oxygen isotope record (CHAPPELL & SHACKLETON, 1986). The curve shows, for the last glacial-interglacial cycle, rapidly rising sea levels (up to 8 m ka^{-1}) during the major post-glacial transgressions, culminating in interglacial high sea level stands between 118-138 ka BP (reef complex VII) and from 8.2 ka BP (reef I). The Last Interglacial sea-level on the Huon Peninsula (124 ka BP; reef VIIa) is assumed to have reached ~ 6.0 m above present sea level. Slower rising sea levels (up to 2.5 m ka^{-1}) were recorded by Huon fringing reef development at 100 (reef VIa), 81 (Va), 59 (IVa), 45 - 40 (IIIa - IIIb) and 28 (II) ka BP, representing a series of high sea level interstadials falling progressively further below present sea level at - 9 ± 3 m, -19 ± 5 m, -28 ± 3 m, -41 ± 4 m and -44 ± 2 m respectively (Fig. 4). This sequence has subsequently been confirmed at several other localities on the western margins of the Pacific basin (e.g. Timor, Indonesia (see below) and the Ryukyu Islands (KONISHI *et al.*, 1974). However, despite these valuable insights into the nature of the sea-level, record, coral terrace sequences provide no comparable information on the exact magnitude and precise timing of low sea-level stands. Furthermore, only a comparatively small portion of the complete raised reef sequences can be dated because of the problems of diagenesis of exposed marine carbonates. This has largely limited the dating of the record to the last 350 ka. Electron spin resonance (ESR) dating methods now allow the extension of the U-series record to ~ 600 ka BP, but with lower levels of precision. Beyond this timescale, and assuming that uplift rates have remained linear, extrapolation to undated reef terraces has allowed

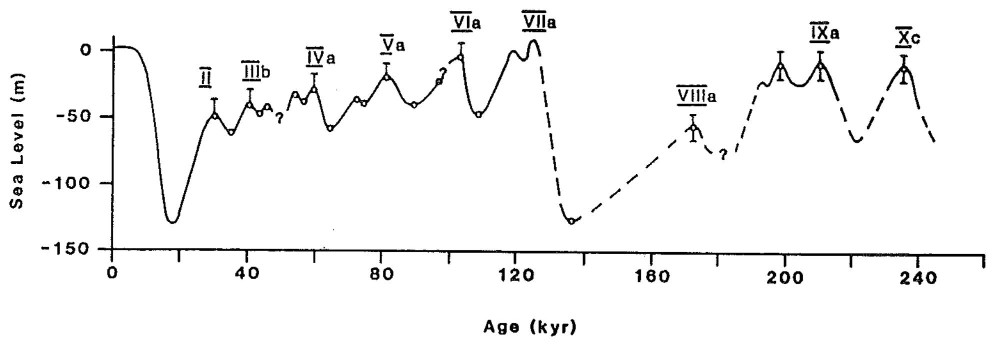

Fig. 4 : Sea level curve and high sea-level stand reef units, Huon Peninsula, New Guinea (after CHAPPELL & SHACKLETON, 1986).

sea-level reconstruction to 700 ka BP at Atauro, E. Timor (CHAPELL & VEEH, 1978) and to 900 ka BP at Sumba, Indonesia (PIRAZZOLI *et al.*, 1993), although clearly these sea-level reconstructions are less secure than the Huon record.

LOW SEA-LEVEL STANDS : CORAL REEFS SUBJECTED TO TECTONIC SUBSIDENCE

By comparison, the recovery of long sequences of shallow water carbonates from atoll basements offers the possibility of reconstructing a record of sea-level history which includes low sea-level stands. It has long been known that atoll cores contain alternations of limestones and unconsolidated sands which can be correlated between islands, suggesting periods of island emergence and submergence respectively and therefore a record of sea-level fluctuations (e.g. Enewetak and Bikini Atolls : LADD & SCHLANGER, 1960; SCHLANGER, 1963). Some attempts were made during the 1980s to quantify the magnitude of sea-level fall represented by some of these unconformities : thus SCHLANGER and PREMOLI SILVA (1986) estimated a Middle Oligocene (~ 30 Ma BP) sea-level fall of at least 100 m below present sea-level was responsible for the creation of the - 825 m solution unconformity in Enewetak core E-1 and the associated turbidite deposits of the Nauru Basin and the Line Islands. At Midway Atoll a Late Miocene sea-level fall of - 75 to - 125 m was suggested (LINCOLN & SCHLANGER, 1987). More recently, the development of stable isotope and new radiometric dating techniques, allied to improved petrographic and biostratigraphic analyses, have allowed the resolution of these coarse differences in core composition and isolated estimates of sea-level change into a much more precise and complete sea-level record.

Rising or stable sea-level on a subsiding platform yields an uninterrupted depositional sequence; if sea-level falls but at less than the subsidence rate then deposition will not be interrupted but the reef sequence will be compressed. If eustatic sea-level fall exceeds the subsidence rate then the platform will be subaerially exposed. Experience at Enewetak Atoll suggests that this fall must be in excess of 30 m before an event is preserved in both reef and lagoon stratigraphies (LINCOLN & SCHLANGER, 1991). Exposure results in the development of rugged karst terrains, with reddish-brown palaeosols, on exposed surfaces. The water table on such surfaces is generally within + 1 - 2 m of present sea-level. Within this freshwater layer, diagenesis of marine carbonates results in particular cement characteristics and oxygen and carbon isotope signatures (e.g. MOORE, 1989; ALLAN & MATTHEWS, 1982). Subsequent sea-level rise and saltwater intrusion shifts the isotope curves back to more enriched values (Fig. 5;

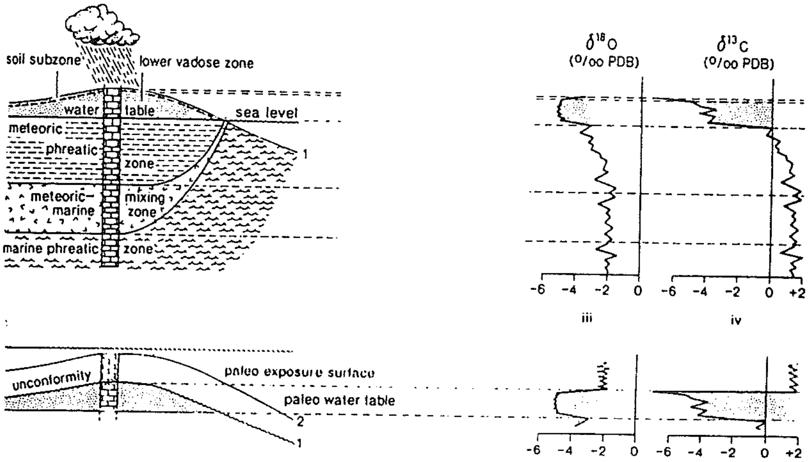

Fig. 5 : Upper : Groundwater system of an exposed carbonate platform and isotopic signatures related to freshwater diagenesis within the groundwater lens. Lower : Submergence and deposition of marine carbonates shifts isotope curves towards more enriched values (after WHEELER & AHARON, 1991).

WHEELER & AHARON, 1991) and preserves the exposure surface under new reef carbonates.

These studies need to be tied to absolute timescales to establish the exact timing and duration of sea-level change. Curently, poor-resolution biostratigraphic methods of age determination (e.g. ADAMS, 1984; BERGGREN *et al.*, 1985) are being replaced by the use of Strontium isotope techniques. It is known that the seawater $^{87}Sr/^{66}Sr$ ratio has shown a progressive increase since the Eocene; strontium incorporated into the crystal lattice of marine carbonates preserves this changing ratio. Absolute ages are calculated by comparing the Strontium isotope composition of samples to an established seawater reference curve. The latter has been derived from deep-sea sediments whose ages are inferred from biostratigraphy and magnetochronology (e.g. BURKE *et al.*, 1982; KOAPNICK *et al.*, 1985, 1988; CAPO & DE PAOLO, 1990; SALLER & KOEPNICK, 1990; HODELL *et al.*, 1991). The particular value of the Strontium dating technique in sea-level studies is that lateral shifts in the Sr ratio at specific levels in the record indicate periods of erosion when part of the records has been lost; the strength of the shift should indicate the duration of erosion and thus exposure (Fig. 6; WHEELER & AHARON, 1991).

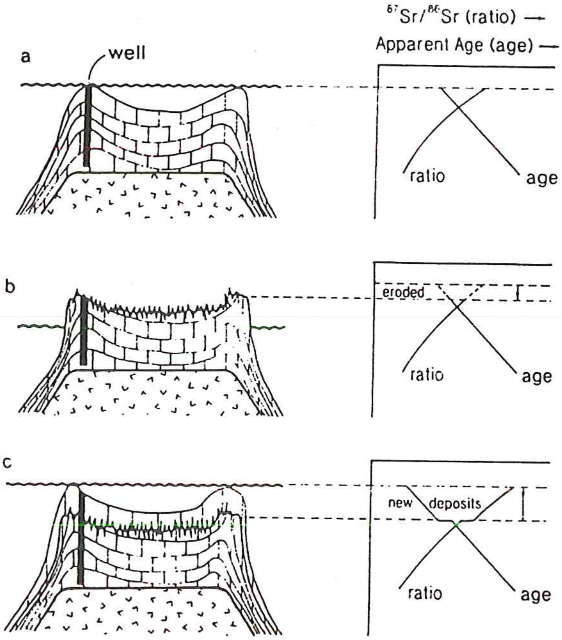

Fig. 6 : Cartoon (after WHEELER and AHARON, 1991) to show development of discontinuity in strontium ratios and ages as a result of the removal of part of the reef growth record (middle), prior to a further phase of marine sedimentation (bottom).

RECONSTRUCTION OF A COMPREHENSIVE SEA LEVEL RECORD FROM ATOLL CORES

Correct identification of low sea-level stands forms only part of the process of sea-level reconstruction from atoll core records. A full synthesis requires a consideration of several methodological and interpretive issues (LUDWIG *et al.*, 1988; QUINN *et al.*, 1991); some of these address geomorphological questions.

If the subsidence rate of the atoll basement can be reconstructed (for methods see : LINCOLN & SCHLANGER, 1991; WHEELER & AHARON, 1991), then any subsurface reef facies of known age can be restored to its depositional position. The use of non-reef facies is more difficult as reconstructions must allow for the depth of deposition : at Enewetak, for example, there is a difference of 60 m between the reef flat and the lagoon floor (EMERY *et al.*, 1954). Simple "backtracking" of this nature provides only a partial reconstruction of the sea-level record. Firstly, new sediments are not deposited on top of the exposure surface until relative sea level rise (from eustatic change, subsidence, continued downwearing or a combination of these processes) floods the surface. Secondly, a fundamental problem is that with sea-level fall those sediments closest to former higher sea-levels are removed first and therefore the record of high stands is preferentially lost. This puts a premium on the correct estimation of erosion rates in order to try to reconstruct former high stands.

The erosion problem

What were the likely rates of denudation on exposed platforms ? Clearly at one extreme the rate of downwearing cannot have exceeded the sea-level fall. Several authors, however, have favoured the other extreme and argued for no erosion on exposure or at least a degree of erosion which can be ignored for sea-level reconstructions (LINCOLN & SCHLANGER, 1987; MAJOR & MATTHEWS, 1983). This is despite the fact that modelling has shown that erosion effects are not trivial : thus LINCOLN and SCHLANGER (1991) show that a Late Miocene unconformity will appear - 50 m lower at an erosion rate of 0.35 cm ka^{-1} compared to a zero erosion rate. Furthermore, such an erosion rate would add five additional unconformities to the whole core record (QUINN, 1991b).

The erosion problem has been highlighted in a debate between QUINN and GRAY. QUINN and MATTHEWS (1990), in using a forward model of diagenetic alteration to determine subsidence and sea-level histories at Enewetak Atoll, argued for no subaerial erosion on exposure as erosion rates greater than zero generated more expected unconformities than were present in the study core. In comment, GRAY (1991) suggested that the number of preserved unconformities should be less than the expected number for several reasons : "weak" unconformities go unrecognized; large erosional events remove previously recorded events; and insufficient dating control does not adequately define high-frequency sea-level cycles. In favouring significant erosion, she drew attention to the measurements of erosion rates on modern raised reefs of 0.37 cm ka^{-1} on Grand Cayman Island, West Indies (SPENCER, 1985) and 0.26 cm ka^{-1} on Aldabra Atoll, Indian Ocean (TRUDGILL, 1976) and to the 6 - 24 m of sedimentation commonly observed on Holocene reefs compared to the ca. 3 m of sedimentation expected from subsidence processes alone. In reply (QUINN, 1991a) rightly pointed out the dangers of extrapolating short-term micro-erosion meter results to the long-term geological record and that high erosion rates require not only a period of platform exposure but also the availability of high rainfall inputs to drive solutional processes. Interestingly, elsewhere, and most notably in the Maldives, relationships have been established between rainfall and lagoon depths (MENARD, 1982). At Enewetak, on the basis of the presence of caliche horizons, QUINN (1991a) has argued for semi-arid glacial climates and little available moisture for downwearing of the exposed reef. A further points is that erosion rates and the rapidity of surface downwearing are not synonymous : many actively eroding karst surfaces develop an intricately dissected karst micro-topography which contains both foci of erosion and areas where the original depositional surface is preserved (SPENCER, 1983). This debate, although inconclusive, does point to the need for more accurate specification of erosion rates on emergent platforms; as yet there are no micro-erosion meter studies of contemporary erosion on island limestones of greater than Pleistocene age.

Synthesis

In spite of these difficulties, considerable progress has been made in the reconstruction of long-term, comprehensive records of Pacific sea-level change. The principal application of this methodology has been at Enewetak Atoll where fifteen subaerial exposure surfaces have been recognised in the uppermost 180 m of core KAR-1 (published research includes LUDWIG *et al.*, 1988; SALLER & KOEPNICK, 1990; LINCOLN & SCHLANGER, 1991; QUINN *et al.*, 1991). The reconstructed record of sea-

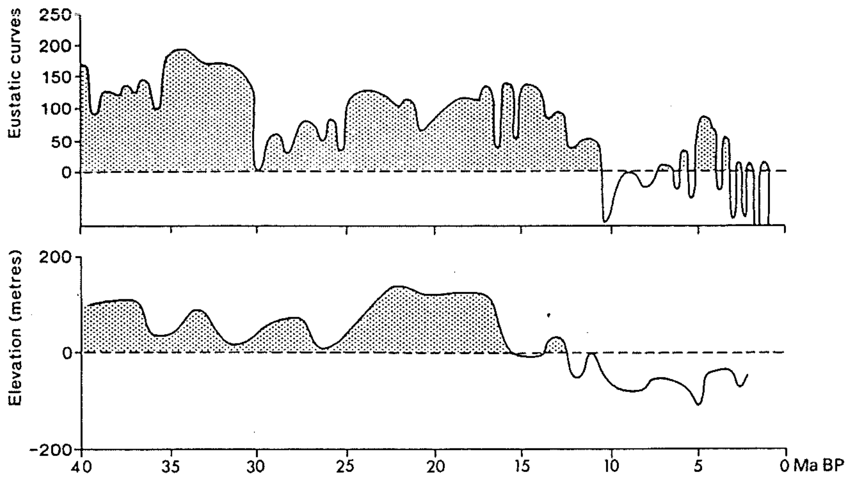

Fig. 7 : Sea-level change at Enewetak Atoll. Upper : Sea level curve of HAQ *et al.* (1987) for passive continental margins. Lower : Reconstructed sea-level curve for Enewetak (after LINCOLN & SCHLANGER, 1991).

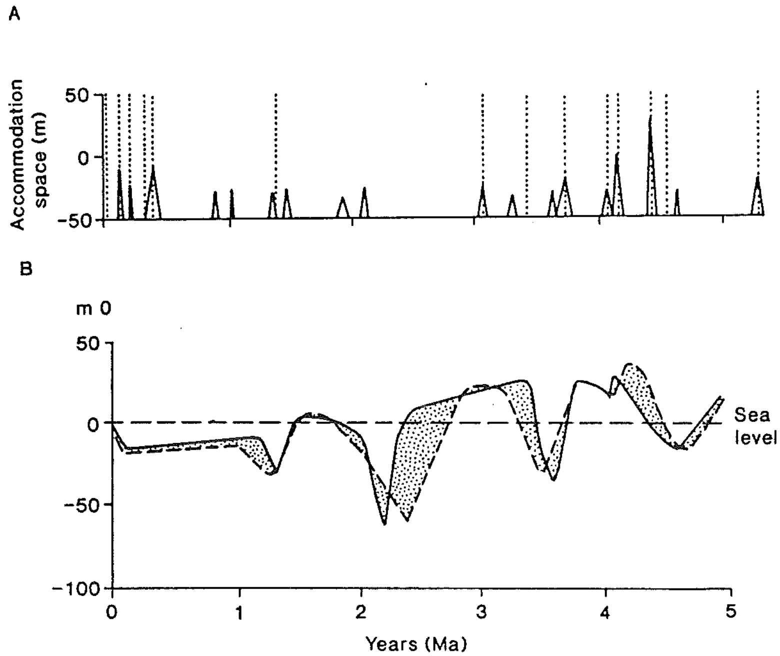

Fig. 8 : Pliocene sea-level change at Enewetak Atoll. a : Model prediction of timing and magnitude of flooding events (after QUINN & MATTHEWS, 1990), b : Sea-level curve envelope from isotopic (solid line) and sequence (dotted line) models (after WARDLAW & QUINN, 1991).

Tab. II : Ages of sealevel events (modified from OHDE & ELDERFIELD, 1992).

Kita-Daito-Jima[1]	Niue[2]	Enewetak[3]	Deep-sea cores[4]	Continental[5] margins
16.4				16.5
15.5				15.5
		13.6	13.6	13.8
				12.5
10.9		11.5	11.3	10.5
		7.9		8.2
				6.3
5.2	5.4 - 4.3	5.5		5.5
	3.7 - 3.2			3.8
				2.9
1.8		1.2		1.6
				0.8

1 : OHDE & ELDERFIELD (1992) ; 2 : WHEELER & AHARON (1991); 3 : LINCOLN & SCHLANGER (1991 and QUINN *et al.*, (1991); 4: MILLER *et al.* (1991); 5: HAQ *et al.*, (1987).

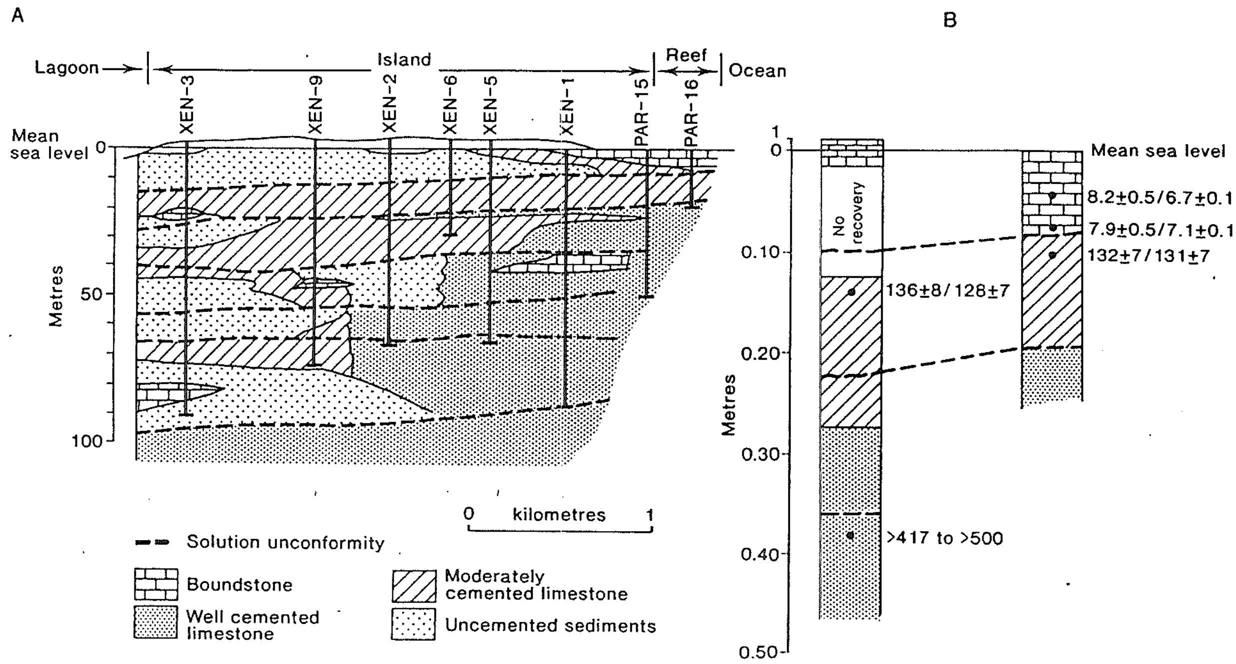

Fig. 9 : Diagrammatic section from Enjebi Island, Enewetak Atoll, showing five Quaternary solution unconformities (after LINCOLN & SCHLANGER, 1987).

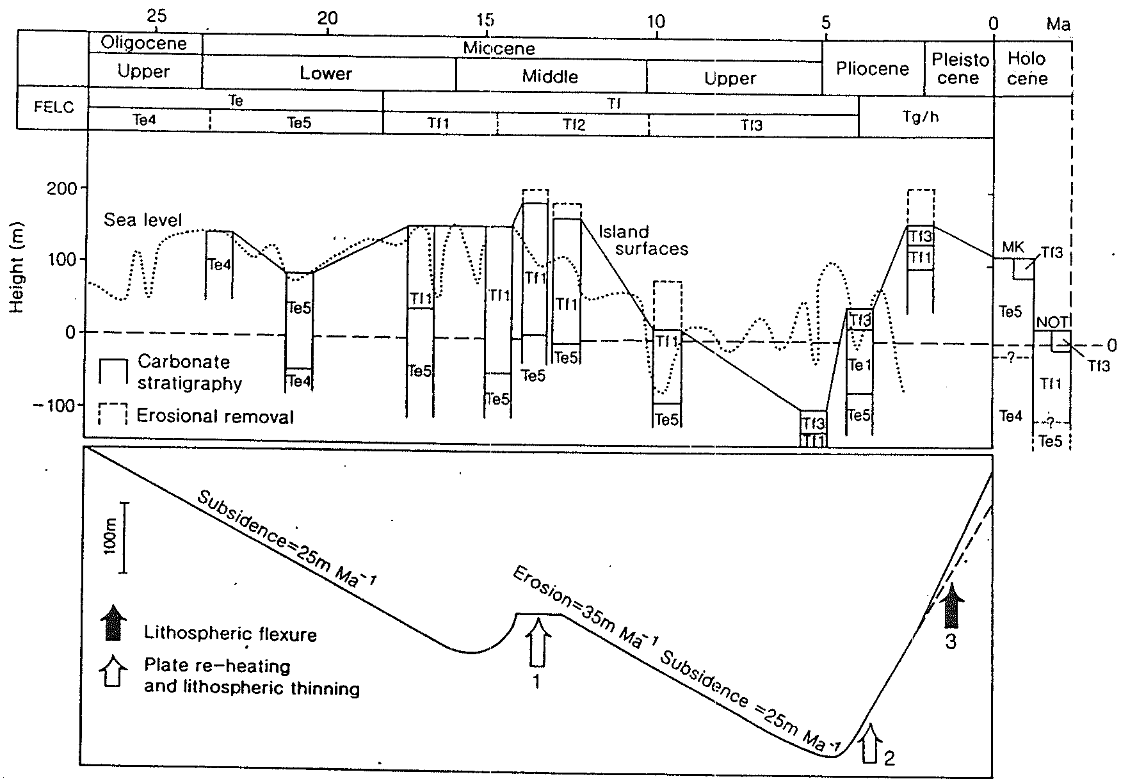

Fig. 10 : Geological history of the NW Tuamotus, based on the stratigraphy of Makatea Island (after MONTAGGIONI, 1989). Biostratigraphy from ADAMS (1970); sea-level curve from HAQ *et al.* (1987); erosional removal after LINCOLN & SCHLANGER, (1987).

level change since the Oligocene for Enewetak Atoll is shown in Fig. 7. Fig. 8 shows the detail of the Piocene record. This suggests a maximum high stand of + 36 m, and a maximum low sea-level stand of - 63 m, relative to present mean sea-level (WARDLAW & QUINN, 1991). Fig. 9 indicates the six intervals of reef growth of the Quaternary record (SZABO *et al.*, 1985).

These records show reasonably good correspondence (Tab. II) to the sea-level record established for Kita-diato-jima (OHDE & ELDERFIELD, 1992) and Niue (WHEELER & AHARON, 1991). Perhaps more importantly - because they provide estimates of sea-level change which have been arrived at independently using different principles - they also show a fair correlation with the sea-level record from passive continental margins (HAQ *et al.*, 1987, 1988) and the reconstruction of sea-level events provided by oxygen isotope stratigraphy (MILLER *et al.*, 1991).

INTRA-PLATE TECTONICS AND SEA-LEVEL RECONSTRUCTION

One reason why the correlation of sea-level events within the ocean basins, and between ocean basins and continental margins, remains difficult is that the sea-level record of many islands cannot be arrived at by the simple subtraction of a subsidence terms. One effect which is difficult to quantify is the migration of geoidal anomalies across the ocean basins (NUNN, 1986). A more tractable but nevertheless difficult problem concerns intra-plate tectonics. Even at Enewetak Atoll some authors have argued for multiple episodes of regional and volcanism from Cretaceous times (e.g. LINCOLN *et al.*, 1989). Thus at Makatea Island, North west Tuamotu Archipelago, currently a raised atoll reaching 113 m above present sea-level, subsidence appears to have been offset by two periods of uplift, the first resulting from plate re-heating and the second by a process known as lithospheric flexure (Fig. 10; MONTAGGIONI, 1989). This latter process has recently been recognised as of considerable significance in modifying simple explanations of sea-level change. Flexing of the lithosphere as it approaches the plate subduction zones has resulted in the uplift of coral atolls. The lithospheric bulge associated with such plate margins has been described from the Loyalty Islands, near the New Hebrides subduction zone, and from Niue Island which is approaching the Tonga trench (DUBOIS *et al.*, 1975). The raised atoll of Kita-daitojima, on the Philippine plate and approaching the Ryukyu trench, is estimated to have been uplifted ~ 43 m over ~ 2 Ma, at an uplift rate of 22 m Ma^{-1} (OHDE & ELDERFIELD, 1992).

Lithospheric flexure and isolated loads : island geomorphology from the South Pacific.

Some islands chains, the Society Islands for example, exhibit Darwin's temporal sequence of atoll development over space as expected (MORHANGE, 1992); elsewhere, however, this is not the case. The Southern Cook - Austral islands archipelago consists of two sub-parallel chains of volcanic islands, seamounts and coral reef islands extending over 2000 km across the south-central Pacific. The simplest model for the archipelago's evolution suggests island formation and subsequent northwesterly drift from volcanically-active McDonald Seamount. However, K-Ar dating of terrestrial volcanics within the group have shown that whilst distance from hotspot / plate age relationship hold in the Austral Islands, in the Southern Cooks only Mangaia fits the expected trend with at least four islands (Mauke, Mitiaro, Atiu and Rarotonga) being substantially younger than expected. Furthermore, simple age-distance relationships are complicated by more than one phase of volcanism being recorded on the islands of Rurutu and Aitutaki (TURNER & JARRARD, 1982; STODDART & SPENCER, 1987).

Tab. III : Island ages and altitudes of volcanic and reef units, Austral - southern Cook Islands chain (after SPENCER *et al.*, 1988).

Location	Age of Volcanics (Ma BP)	Maximum Altitudes (m) Volcanics	Tertiary Makatea	Pleistocene Reef Limestones
Rapa	5.0 - 5.2	670		
Raivavae	5.5 - 7.4	437		
Tubuai	8.6 - 10.4	422		
Rurutu	12.3 0.6 - 1.9	389	100	15
Rimatara	>28.6 >21.2 >14.4 >4.8	92	11	
Mangaia	17.4 - 19.4	169	73	14.5
Mauke	>6.0	24.4	14.7	10.0
Mitiaro	>12.3	8.9	10.9	9.8
Atiu	8 - >10	72	22.1	12.2
Rarotonga	>1.6 - 2.3	653	-	3.5
Aitutaki	>8.1 0.7 - 1.9	124		

Theory suggests that the islands at the northwestern end of the chain should be coral atolls in form. However, whilst Palmerston and Manuae can be classifield as of this island type, Aitutaki retains island volcanics, and is known as an "almost-atoll", while Mangaia, Atiu, Mitiaro and Mauke are all *makatea* islands, possessing volcanic cores surrounded by uplifted limestones (STODDART *et al.*, 1990). Most anomalously of all, Rarotonga is a high volcanic island with contemporary fringing reefs and small, low outcrops of Pleistocene raised reef (STODDART *et al.*, 1985; SPENCER *et al.*, 1988). In the Austral Islands, raised limestones are also present on Rimatara and Rurutu (PIRAZZOLI & VEEH, 1987; PIRAZZOLI & SALVAT, 1992) and throughout the Cooks - Austral chain the juxtaposition of different island types with widely differing elevations point to considerable differential vertical movements (Tab. III). The association of young (< 2 Ma old) volcanoes with raised coral reefs is, however, characteristic of much of the South Pacific Ocean (Fig. 11) and led McNUTT and MENARD to argue in a classic paper in 1978 that island uplift is the result of a process of lithospheric flexure : loading of the lithosphere by a volcanic load causes an island-marginal moat and a compensatory uparching at some distance from the load which may convert sea-level coral atolls into raised reef islands (Fig. 12). Although near-island moats have been described around some mid-plate volcanoes (Hawaii : TEN BRINK & WATTS, 1985; WATTS *et al.*, 1985; Marquesas : FISCHER *et al.*, 1986) most bathymetric charts are insufficiently detailed to reveal such features. As a result, most studies of this process have concentrated upon defining the magnitude and rate of flexure from a knowledge of the timing of the load, from K-Ar dating of supposed volcanic loads, and from the determination of the degree of uplift from accurate measurements of raised limestone altitudes at varying radii from the load. Clearly such studies assign a key role to the careful geomorphological investigation of raised reef topographies. In the Southern Cook Islands, Late Pleistocene uplift rates upon flexure have been calculated at between 4.4 cm ka^{-1} (for Mangaia : ~ 200 km from Rarotonga and on the arch crest) and 1.0 cm ka^{-1} (for Mitiaro and Mauke : 260 - 280 km from the load) (SPENCER *et al.*, 1988). There is reasonable agreement between such field-controlled calculations and geophysical models which predict uplift as a function of load density, degree of marginal moat infill and flexural rigidity of the lithosphere (Fig. 13; LAMBECK, 1981b), although it is difficult to model the close association of large-scale uplift with much more modest emergences at small distance increments from the load (LAMBECK, 1981a).

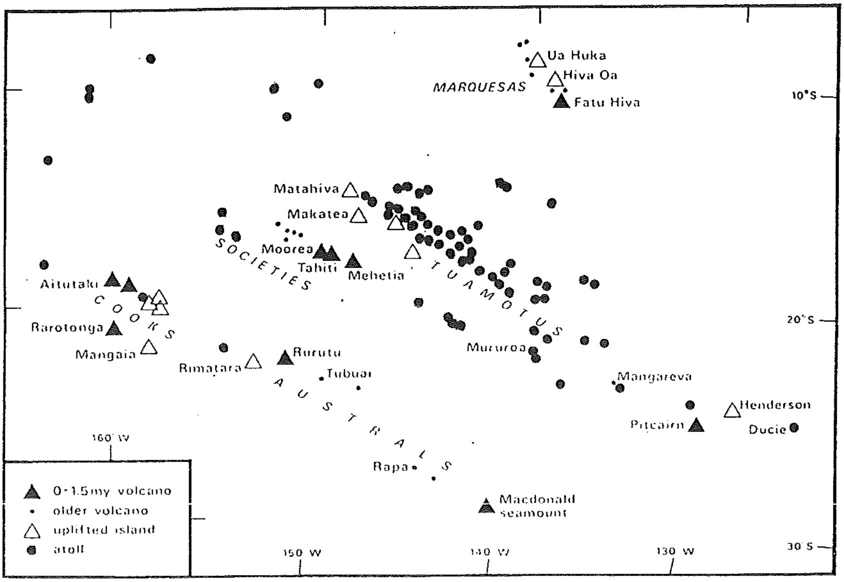

Fig. 11 : Islands types in the South Pacific. Note the association of uplifted islands with young volcanoes (after McNUTT & MENARD, 1978).

It is possible to quantify flexure-controlled uplift more precisely by the topographic survey and radiometric dating of Pleistocene limestones which veneer and surround the older Tertiary core limestones ("makatea" *sensu stricto*) on raised reef limestones. This requires, however, a separation of tectonic effects from those resulting from glacio-eustatic fluctuations in sea level. Thus tectonically-effected Makatea Island has cliff-veneering apron reefs, bounded at their upper margin by notch lines and caves 5 - 8 m above present sea level (MONTAGGONI *et al.*, 1985) which have been assigned to the high sea-level stand of 100-140 ka BP on the basis of limited uranium-series age determinations (VEEH, 1966).

In the Southern Cook Islands, presumed Last Interglacial (i.e. reef complex VII) reef limestones reach 3.48 m above present sea-level at Ngatangiia and 2.2 m at Nikao Rarotonga and have been attributed to eustatic fluctuations in sea-level (STODDART *et al.*, 1985). By comparison dated Last Interglacial reef limestones on Mangaia (101 - 135 ka BP : VEEH, 1966; SPENCER *et al.*, 1988) reach 14.5 m above present sea level and may relate to a sea level of + 20 m above present (STODDART et al., 1985). On neighbouring Atiu, Mitiaro and Mauke, there are two Pleistocene reef units, a

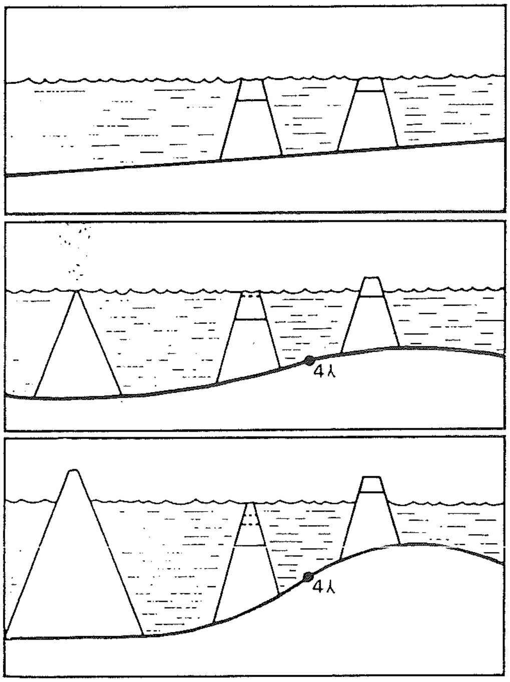

Fig. 12 : Process of lithospheric flexure (after McNUTT & MENARD, 1978). Coral islands at the radius of the lithospheric bulge will be uplifted; those in the moat will show continued reef growth.

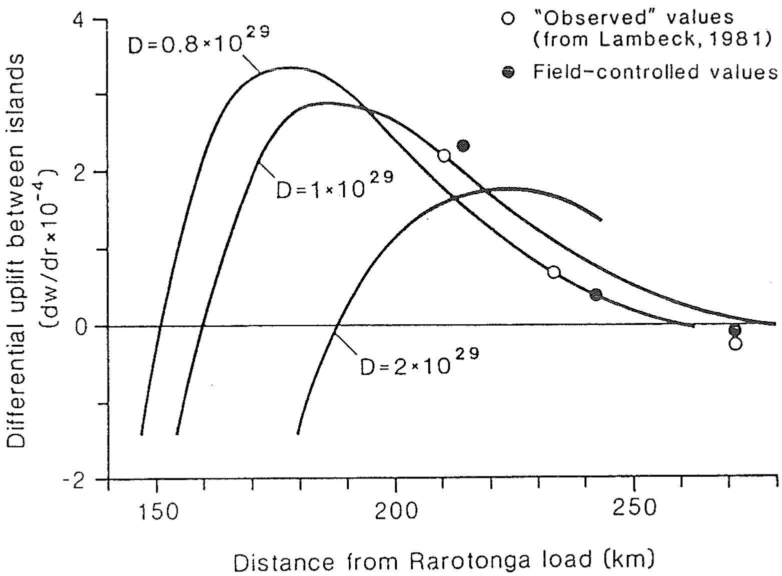

Fig. 13 : Comparison of geophysical profiles (from LAMBECK, 1981b) with field data on uplift with distance from load in the southern Cook Islands (after SPENCER *et al.*, 1988).

Tab. IV : Elevation of sea level features and calculation of differential uplift rates, southern Cook Islands

Location	Distance from Load (km)	Highest Makatea (m)	Uplift Rate cm ka^{-1}	Highest Stage 5 e (Last interglacial) reef (m)	Uplift cm ka^{-1}	Highest Holocene (m) *****
Rarotonga	0	-	-	3.5*		0-1 (1.0)
Mangaia	202	73	4.4	14.5 (20.0)**	7 - 10	1.7
Atiu	222	22.1	1.3	12.2	5	0 - 1
Mitiaro	264	10.9(12.3)***	0.7	9.8	3	1.2
Mauke	281	14.7	0.9	10.0(12.7)****	3 - 5	< 1

* Presumed age; see WOODROFFE *et al.* (1991);
** 14.5 m is highest observed 5 e; but highest sea-level may have been 20 m (see STODART *et al.*, 1985)
*** Maximum spot height of 12.3 m from S. SWABEY (pers. comm. 1992)
**** 10.0 is general level of 5e reef; 12.7 is maximum observed (see STODDART *et al.*, 1990)
***** For details see WOODROFFE *et al.* (1990) and text.

stratigraphy which appears to be replicated elsewhere in the south Pacific Ocean (VEEH, 1966; Tonga : TAYLOR & BLOOM, 1977; Loyalty Islands : BOURROUILH-LE JAN, 1985; Makatea Islands : MONTAGGIONI *et al.*, 1985). The lower, fossiliferous unit provides Uranium-series ages of 187 - 205 ka BP (stage 7) at Atiu and 204 - 243 ka BP on Mauke. The upper limestone unit on these islands is clearly of stage 5e age : 111 - 135 ka BP at Atiu, 112 - 128 ka BP at Mitiaro, and 119 - 136 ka BP at Mauke (WOODROFFE *et al.*, 1991), with an average age of 115 ka BP at Mangaia (SPENCER *et al.*, 1988).

Table IV summarizes key altitudinal data and calculated uplift rates from the Southern Cook Islands. The rates of differential movement indicate that Mangaia, on the crest of the lithospheric bulge, has experienced the most rapid uplift whereas those islands at greater distances from the the Rarotongan load have experienced lesser degrees of emergence. Clearly, however, none of these islands provide a stable platform for the estimation of eustatic sealevel fluctuations through the late Pleistocene.

Lithospheric flexure with continued lithospheric loading; island geomorphology from the North Pacific.

Complications to these lithospheric flexure processes occur in linear island archipelagoes where there are sequential additions of new volcanoes at the active ends, of the chain. In this situation, large-scale flexing of pre-existing "downstream" volcanoes accompanies the appearance of new volcanic material. Pre-existing volcanoes in the near-load flexural moat experience enhanced subsidence while volcanoes on the more distant compensatory arch undergo uplift. This process is further complicated by the migration of these new loads in the direction of plate motion; thus over time and over space, island chains show flexural responses to both new and pre-existing loads. The magnitude and timing of such adjustments are poorly known at present but some clues are emerging from studies within the southeastern Hawaiian Islands, whose large mass (estimated at ~ 6 x 10^{17} kg; WATTS & TEN BRINK, 1989), relatively isolated location in the North Pacific away from plate marginal or other island chain influences and considerable archive of cognate research make them appropriate for study.

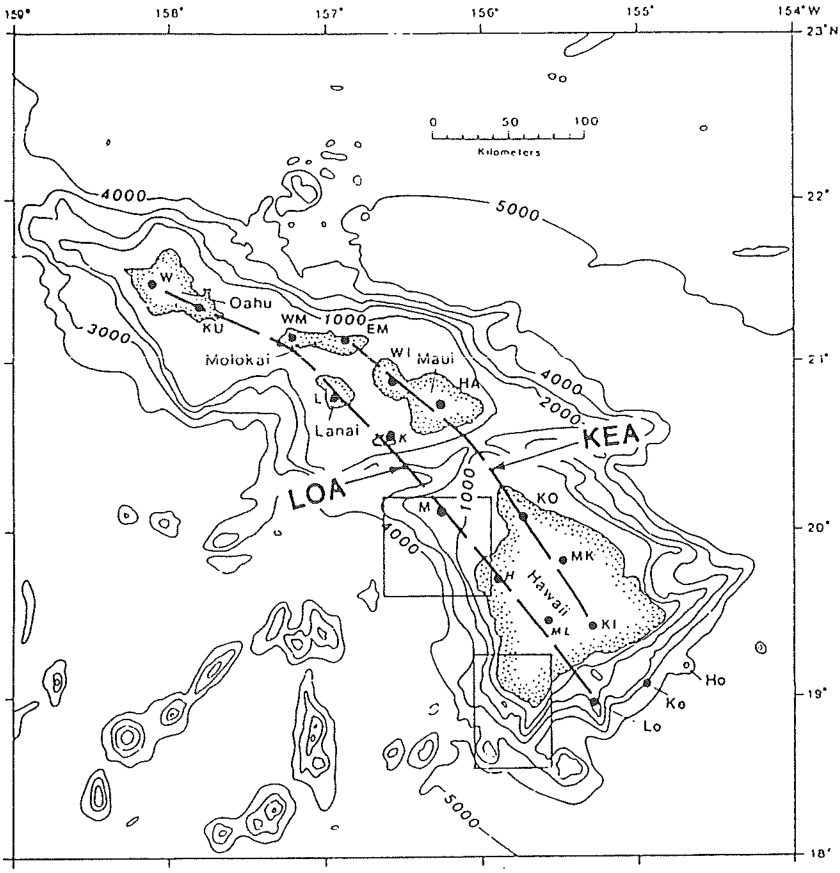

Fig. 14 : Bathymetry of southeastern section of Hawaiian Ridge (contour interval = 1000 m). Heavy lines indicate Loa and Kea lineations; closed circles indicate volcanic centres (after MOORE & CLAGUE, 1992).

The southeastern Hawaiian volcanoes generally occur at a spacing of 40 to 60 km on one of two lines, the Loa and Kea ridges (DANA, 1890), 30-40 km apart (Fig. 14). Using evidence from volcanic histories on and adjacent to the island of Hawaii, MOORE and CLAGUE (1992) suggest that the time from the start of seafloor volcanism to the end of the shield-building phase is ~ 600 ka. Thus the youngest volcano along the Hawaiian chain, in the process of replacing Mauna Loa and Kilauea as the main centres of volcanism, is Loihi seamount, 100 ka old and currently 1 km below sea level, on the Loa line; MOORE and CLAGUE (1992) estimate that it will reach sea level in ~ 250 ka and complete its shield building phase in 500 ka. Volcano growth in the southeastern Hawaiian Islands is accompanied by rapid subsidence caused by the withdrawal of large volumes of magma from beneath the crust (total rate of magma generation calculated at 6.5 $m^3 s^{-1}$, 2 - 3 % of the overall global rate; WATTS & TEN BRINK, 1989), and crustal loading. MOORE (1987) has calculated that the bases of the volcanoes eventually subside by 5 - 8 km and that most of this subsidence is completed within a million years of the end of shield-building. The zone of subsidence is cone-like and has an estimated radius of 100 km : tide gauge records suggest that a subsidence effect of 2.4 mm a^{-1} can be extracted from the gauge at Hilo, Hawaii whereas no such effect can be seen in the record at Kahului, Maui, 180 km distant (MOORE, 1987). Furthermore, this rate appears to have characterised the last 450 ka (LUDWIG *et al.*, 1991).

Processes of individual island formation and subsequent evolution can be calibrated from the study of island marginal bathymetry and drowned coral reefs. During the main shield-building phase of island construction, the shoreline is extended subaerially by lava flows; when these flows enter the sea the offshore slope is mantled by rapidly chilled volcanic materials. This difference in the behaviour of lava above and below sea level leads to differences in characteristic slope-angle, typically 4° in subaerial and 13° in submarine environments respectively (MARK & MOORE, 1987). On completion of shield-building this slope change is maintained but, with the dominance of island subsidence over construction, the slope break is progressively drowned over time.

The cessation of shield-building also allows for the growth and preservation of coral reefs at and above this change in island slope. The detailed sequencing of reefs during the Pleistocene has been controlled by the interaction of progressive subsidence and episodic glacio-eustatic sea-level fluctuations at approximately 100 ka intervals. In the glacial periods, the offsetting of subsidence by similar rates of sea-level fall produced relatively stable sea-levels and the development of extensive carbonate platforms (CAMPBELL, 1984). During the interglacial periods, the combination of subsidence and sea-level rise produced 10 ka bursts of rapid sea-level rise exceeding vertical reef growth potential (in this region estimated at 10 mm a^{-1}; GRIGG & EPP, 1989). Modelling of these processes suggests reef "drown-out" events at 18, 130, 245, 340 and 430 ka (MOORE & CAMPBELL, 1987; LUDWIG *et al.*, 1991). Stepped reef sequences, analogous to the New Guinea sequences, have been mapped around the islands of Hawaii, Maui, Lanai and Molokai; the terraces are typically 5 km (range : 1 - 9 km) in width and each terraces is bounded by a steep face, 100 - 250 m high, which intersects the inward margin of the next terrace below (MOORE & CAMPBELL, 1987). The two highest reefs, at - 150 m and - 430 m, are particularly well developed. These reefs have been sampled by submersible and dredging methods and samples dated using Uranium-series (LUDWIG *et al.*, 1991) and Strontium techniques (LUDWIG *et al.*, 1992). A further old reef is known from the slopes of Maui (MOORE *et al.*, 1990). Agreement between model and field depth/age relationships is good for the two youngest reefs but less secure on deeper terraces (Tab. V) : beyond the usual radiometric dating problems of contamination and system closure, ages too young for their depth (e.g. reef 4) may indicate downslope translation of younger reefal material

Tab. V : Age/depth relationships of drowned coral reef terraces, northwest Hawaii

Oxygen Isotope Stage	Depth (m)*	Age (ka BP)	Predicted Age (ka BP)**	Source and Comments ***
2	- 150	13.25 ± 0.38	13	A; ^{14}C age
2	- 150	17 ± 5 ; 19 ± 5	13	B; Mass-spectrometric U234/U238
	- 360	120 ± 5		C; Alpha-spectrometric 230Th/234U
6	- 430	133 ± 10	128	B; Mass-spectrometric U234/U238
8	- 693	225 ± 12; 226 ± 13; 276 ± 9	251	B; Oldest date shows Th230/U234 and U234/U238 discrepancy : open system ?
10	- 945	314 ± 10; 287 ± 10	347	B; Calcite = 11 % in older date
12	- 1146	406 ± 12; 475 ± 38 360 ± 12	440	B
14	- 1336	463 ± 8	502	B
	ca.- 1738	750 ± 13	730	D

* Location : northwest Hawaii; ** After MOORE and CAMPBELL (1987)
*** Sources : A : MOORE and FONARI (1984) B : LUDWIG *et al.* (1991)
C : SZABO and MOORE (1986) D : MOORE *et al.* (1990)

from above whereas ages too old for their depth (e.g. one date in reef 5) may reflect sampling of diagenetically altered reef core, exposed and altered by freshwater leaching following exposure by subaerial solution or landsliding, rather than the minimum age preserved in the reef veneer (LUDWIG *et al.*, 1991).

As subsidence is greatest closest to an active volcano, horizontal reef platforms have become tilted towards the volcano with time; thus for six reef terraces between Molokai and northwest Hawaii this gradient has been calculated at 4 - 6 m km^{-1} (MOORE & CAMPBELL, 1987). The deeper continental shelf break is similary tilted and shows deformation over a wider area. The most prominent of these terraces is the Haleakala, or "H", terrace which is thought to have been formed as a horizontal feature at 0.85 - 1 Ma BP when the Haleakala volcano on Maui was forming (MOORE, 1987; MOORE & CAMPBELL, 1987). The terrace is presently strongly warped to the southeast : it can be traced for over 150 km from north of west Maui, where it is found in water depths of 400 m, to north of Hawaii where the same feature is at over - 2.000 m; the regional slope is 10 - 20 m km^{-1} to the S/SE (MOORE *et al.*, 1990). Interestingly, independent attempts to model this deformation have been remarkably successful : MOORE's (1987) model of a 100 km radius cone subsiding at 2.5 mm a^{-1} at its centre and migrating at 10 cm a^{-1} over 3 Ma "freezes" a final terrace slope beyond the

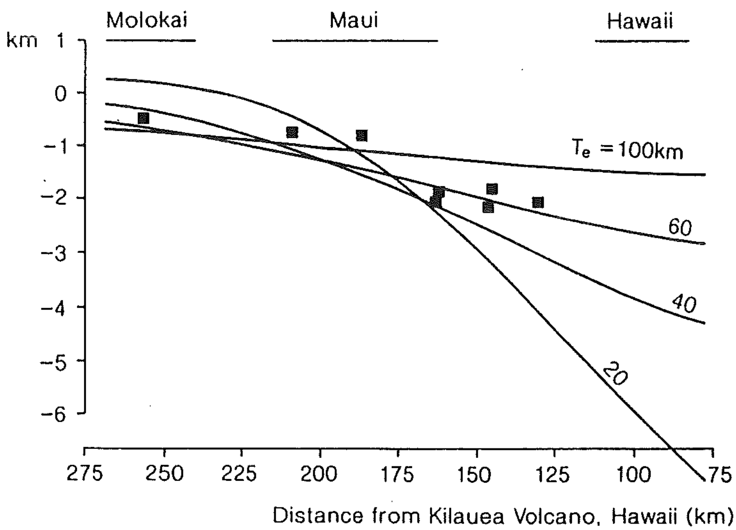

Fig. 15 : Comparison of suite of computed flexure curves due to the load of Hawaii to observed depths of the "H" terrace (after WATTS & TEN BRINK, 1989).

influence of further subsidence at an angle close to the observed tilt of the "H" terrace. Similarly geophysical modelling of lithospheric flexure provides a good fit between flexural profiles and "H" terrace bathymetry with distance from the Hawaiian load (WATTS & TEN BRINK, 1989; Fig. 15). One explanation for the remaining discrepancies between predicted and actual terrace depths is that the adjustment to loading is not yet complete. Tide gauge evidence from Oahu and Maui indicates that Maui is experiencing a relative rate of subsidence of 2 mm a^{-1}, suggesting that a further 0.2 Ma will be required before full adjustment to the load (MOORE, 1970; WATTS & TEN BRINK, 1989).

Loading effects do not only characterise actively growing islands but also have regional effects : thus Hawaii has produced distinct zones of regional subsidence and uplift in its vicinity. The effects of loading normal to the island chain axis are well illustrated by a seismic reflection profiling transect north and south of Oahu (WATTS *et al.*, 1985) showed that the top of the crust dips from depths of 4.5 km at distances of ~ 200 km from the Hawaiian Islands to ~ 6.6 - 9.5 km beneath the islands. In topographic terms, this loading translates into 500 - 600 m of relief between the moat and the flexural arch crest, the latter having a relief of 200 - 300 m above the regional seafloor elevation (WATTS & TEN BRINK, 1989). The moat is infilled by a thick wedge of material. Seismic reflection profiling (TEN BRINK & WATTS, 1985) shows onlapping sequences of sedimentary units within the moat which thicken and increase in dip towards the load. In the plane of plate motion, lithospheric flexure forms a moat up to 2 km deep to the southeast of Hawaii; it is in this depression that the new volcanic island of Loihi is forming. To the northwest, a broad area of subsidence reaching depths of 3 km extends from Hawaii to West Molokai. Beyond this zone there is a broad flexural bulge which reaches its maximum amplitude on Oahu (Fig. 16).

Again, preserved reefs and submarine terraces allow this geometry to be more closely defined. Of particular interest is the fact that only Oahu has extensive coral reef deposits above sea level; the reefs are at altitudes of 7.5 m (Waimanalo Reef) dated to 120 ka, 21.5 m (Laie Reef) and at 29 - 30 m (Kaena Reef), dated at ~ 65 ka BP (STEARNS, 1978). MOORE (1987) has argued that these levels reflect periods of higher sea level in the past whereas STEARNS (1978) used this evidence to suggest these deposits are part of a series of large scale tectonic uplifts of the islands. It seems most likely that the levels result from the combination of regional flexure and sea level change and it is clear that these deposits need more detailed re-examination, and careful

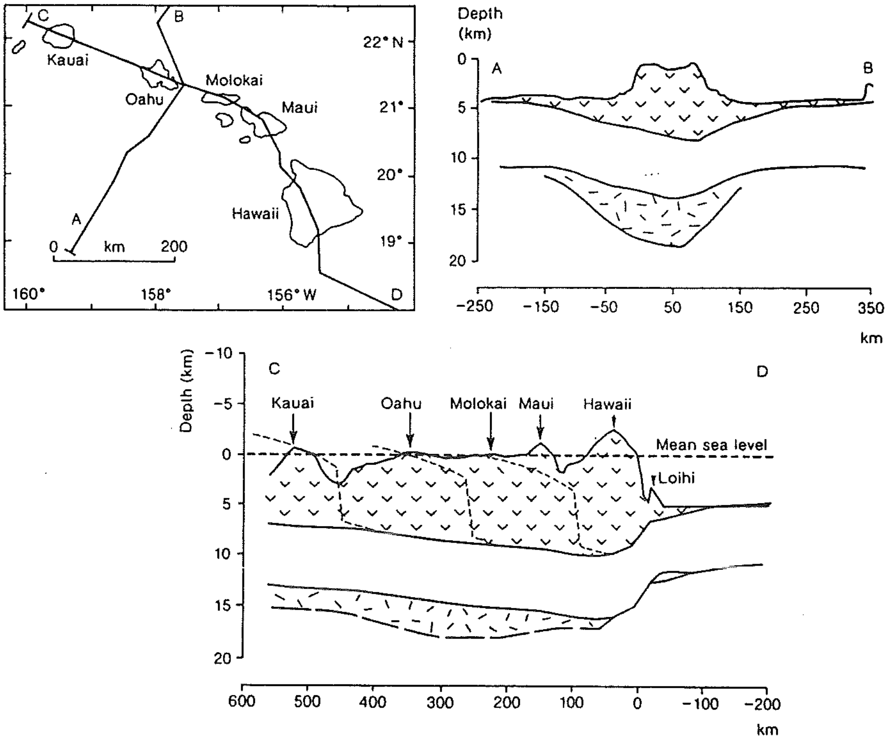

Fig. 16 : Plot of surface topography and top and base of oceanic crust normal to and along the axis of the Hawaiian Islands (after WATTS & TEN BRINK, 1989). Note crustal thickening beneath islands. Dashed lines indicate progressive flexing of the lithosphere as new load is added.

radiometric dating, to define the relative contributions of these two sets of processes over time. Such reconstructions will need to bear in mind one possible type of complication. The volcanoes attain their greatest size at the end of the shield building phase and thereafter are reduced in extent, partly as a result of subsidence and partly as a result of erosional processes. Marginal slopes are steepest, highest and least stable at the end of shield building and this period is characterised by largescale landsliding. It seems highly likely that such activity is preserved by event markers in the sea-level record; thus the youngest failure on the western flank of Mauna Loa, Hawaii, the Alika 2 debris avalanche, has been suggested as the propagator of a tidal wave deposit on Hawaii and adjacent islands (MOORE *et al.*, 1989). The wave associated with this slide is thought to have attained a height of 362 m on Lanai where coral boulder shoreline deposits have been dated at 105 Ka BP (MOORE & MOORE, 1984).

Finally, such studies need also to be combined with reappraisals of terrestrial geomorphology. Several authors established schemes for island ages based upon the degree of island dissection, following the pioneering studies of WENTWORTH (1925, 1927); in the future it might be profitable to look at patterns of river incision and valley alluviation in the context of patterns of flexural subsidence near young loads and flexural uplift at greater distances along the Hawaiian chain.

ADDING OCEAN DYNAMICS TO GEODYNAMICS

Shield building in the southern Hawaiian Islands has been sufficiently voluminous for the islands to have experienced Pleistocene glaciation. On Mauna Kea, at least three phases of glaciation have been recognised : the Pohakaloa (150 ka, probably predating the end of shield-building and relating to a summit elevation of ~ 4,500 m); Waihu (70 - 65 ka); and Makanaka (40 - 16 ka) phases (Fig. 17; PORTER *et al.*, 1987). Snowline depression, and corresponding depression of vegetation zones (PORTER, 1979), have been calculated at 935 ± 190 m on Mauna Kea. On the basis of modern mountain lapse rates, this change in snowline elevation suggests a temperature reduction of 4 - 6.5° C. CLIMAP reconstructions of sea surface temperatures, however, indicate changes of only 2° C. It is difficult to see how this discrepancy may be resolved but part of the answer may lie in changes in the intensity of ocean-atmosphere circulation systems.

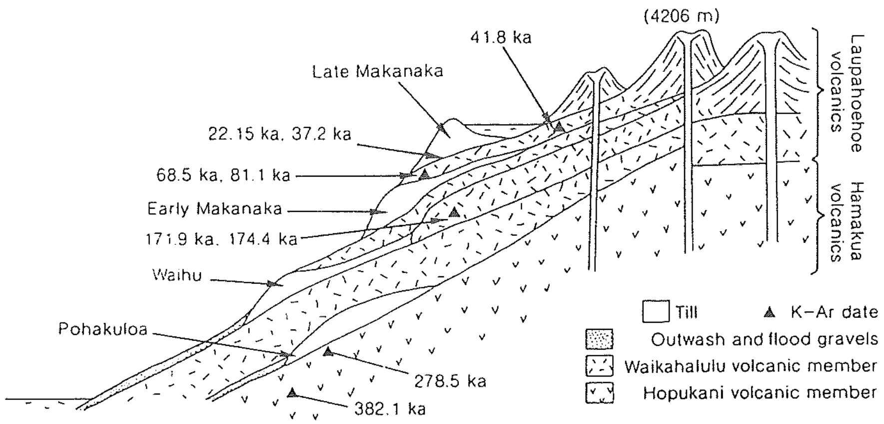

Fig. 17 : Stratigraphic relations and radiometric ages for glacial and volcanic units, Mauna Kea (after PORTER, 1987).

The surface wind stress of the easterly tradewinds coupled to the change in the sign of the Coriolis force at the equator produces a strong divergence in low latitude surface transport in the eastern Pacific. This surface divergence is balanced by shallow upwelling north and south of the equator. The windstress is greatest in the central Pacific and upwelling is greatest here (PHILANDER, 1990). However, the equatorial thermocline slopes upwards towards the east and thus brings deep water nearest to the surface in this region. This upwelling water is characterised by low temperatures (as cold as 14° C at 7° S compared to usual surface temperatures of > 22° C; CANE 1986), high pCO_2 levels and high nutrient concentrations; thus the upwelling zone is characterized by high ocean productivity. It has been estimated that the Pacific Ocean north of 50° S contributes 44 % of total global mean productivity and that the eastern rim of the ocean accounts for half of this total.

This ocean-atmosphere system appears to have been intensified during glacial periods (PEDERSEN, 1983). Studies of quartz abundance distribution in core V19-29 (3°' S, 83°56' W) have suggested strengthened Tradewinds between 73 - 61 ka B.P. and 43 - 16 ka B.P. (MOLINA-CRUZ, 1977). Changing distributions of radiolarian assemblages in deep-sea cores (MOORE *et al.*, 1981) indicate that the increased intensity 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 upwelling reaching its most westerly at the glacial maximum (ROMINE & MOORE, 1981) What might have been the ocean productivity changes (LYLE *et al.*, 1992) associated with these processes ?

One possible means of assessing changes in ocean productivity over time is through the study of island phosphate deposits. Dissolved phosphorous in ocean waters is concentrated up marine food chain and ultimately transported to oceanic islands by seabirds. Phosphate rock results from the interaction of seabird excrement - guano - with island carbonates (HUTCHINSON, 1950). Coral reef islands in the equatorial arid zone of the Pacific Ocean are sites of guano deposition and accumulation for several reasons. The presence of equatorial upwelling supports large seabird populations; island aridity limits vegetation growth and allows nesting by ground-dwelling seabirds; and lack of rainfall inhibits the decomposition and leaching of guano once deposited (STODDART & SCOFFIN, 1983). However, cemented phosphate accumulations are associated with a wide range of rainfall totals and vegetation types on islands in the Pacific basin, suggesting the possibility of either a wider range of environmental tolerance to formation and preservation or changing environmental conditions on islands. There are strong relationships between mean annual rainfall and latitudinal vegetation zones in the central equatorial Pacific basin (STODDART & WALSH, 1992); and an association between the appearance of woodlands of *Pisonia grandis* (a tree which produces an acid raw humus which aids the release of phosphate ions) and cemented phosphates (FOSBERG, 1957). The presence, therefore, of phosphate rock on currently wet Pacific islands at locations devoid of *Pisonia* has been used to argue for climatic change (STODDART & SCOFFIN, 1983). Besides these relationships, phosphate deposits of great volumetric significance are associated with rugged karst terrains on elevated islands. In particular the presence of massive deposits on the islands of Banaba and Nauru suggests the prolonged longitudinal extension of equatorial upwelling from its present more easterly limit. Such a pattern would be consistent with a more intense atmospheric circulation system as outlined above. However, many islands with large phosphate deposits lie outside the equatorial belt and cannot be so readily explained. As an alternative explanation, both BRAITHWAITE (1980) and TRACY (1980) have argued for greater phosphate accumulations on restricted island areas at times of high interglacial sea levels. These competing hypotheses can be tested to some degree (for difficulties see VEEH, 1985) by Uranium-series dating of island phosphates. Summary results (Tab. VI) show three groupings. The first group of dates indentifies a series of

Tab. VI : Uranium series dating of island phosphate deposits : summary data from Pacific islands (after ROE and BURNETT, 1985).

Island or island type	Type of phosphate deposit	Age or age range (ka BP)
"Low" islands	Superficial cemented coral rubble and unconsolidated phosphorite on atoll rim	1.5 - 8.0
	submerged atoll rim	20
Mataiva, Tuamotus	sub-lagoonal	53 130 210 >300
"High" islands [Nauru, Banaba, Makatea]	Highly elevated phosphate in rugged karst terrain	>300

superficial deposits associated with the Holocene transgression. Many of these dates cluster around 4 - 6 ka B.P. and appear to be correlated with a Holocene high sea-level stand (TRACEY, 1980, and see below). By comparison, it is difficult to link the sub-lagoonal phosphate deposits of Mataiva Atoll to high sea-level stands. Rather, they appear to occur near the boundaries between the low and high sea-levels associated with glacial and interglacial periods. Interestingly, this is in accord with estimates of palaeoproductivity obtained from the diatom record of Peruvian coastal shelf cores; shifts in productivity are out of phase with modal glacial and interglacial conditions (SCHRADER & SORKNES, 1991). Finally, the phosphates of the elevated islands are clearly ancient and fall beyond the range of current dating techniques. At the present time, therefore, it is difficult to use phosphate deposits on oceanic islands to construct unambiguous records of ocean palaeoproductivity. A further complication is that the normal functioning of the ocean-atmosphere system in the Pacific, which underpins the use of phosphates as indicators of environmental processes, is periodically interrupted by the suppression of equatorial upwelling during El Nino-Southern Oscillation (ENSO) events. During these episodes the easterly Trades slacken and warm waters from the western Pacific "slosh back" along the equator, flattening and deepening the normally shallow thermocline of the eastern Pacific and thus preventing nutrient-rich waters from reaching the surface (PHILANDER, 1990). Changes in ocean productivity, and in the animal populations which depend upon this productivity, are rapid and catastrophic (e.g. BARBER & CHAVEZ, 1986). Recovery of slow-breeding seabird populations after severe ENSO events means that such events must be be reflected in rates of phosphate accumulation.

HOLOCENE ENVIRONMENTAL CHANGE

The dynamic record of environmental change documented above on Tertiary and Pleistocene timescales has continued into the post-glacial period; once again there are no simple ocean-scale patterns of environmental change but regional differences in response which need to be recognised and categorised.

Part of the explanation of Holocene sea level changes on the Pacific plate lies in the interaction between the volume, melting history and location of sources of meltwater from the decay of ice sheets and the 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 & CLARK, 1976). These effects do characterise large areas but their interactions have given rise to a range of Holocene sea level curves within the Pacific basin and at its margins (e.g. NAKIBOGLU *et al.*, 1983; LAMBECK & NAKADA, 1985). The early Holocene record is dominated by meltwater effects. The full pattern of post-glacial sea-level rise has been recovered by drilling operations into

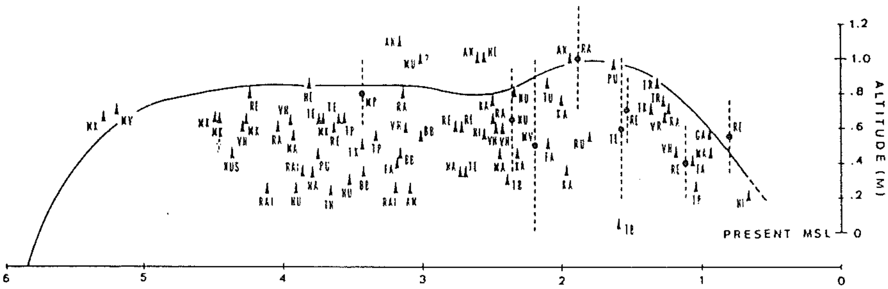

Fig. 18 : Elevation versus time plot of radiocarbon-dated sea-level archive of materials in growth position from French Polynesia. Arows = minimum position of mean sea level; closed circles and dashed lines = estimated mean sea level position and uncertainty limits. Solid line is the best-fit mean sea level curve (after PIRAZZOLI & MONTAGGIONI, 1988a).

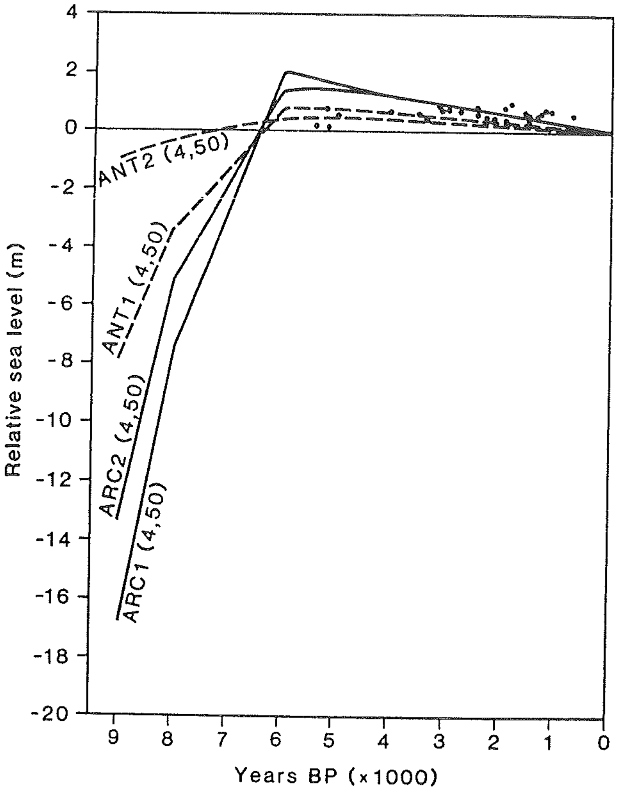

Fig. 19 : Relative sea-level models as a function of ice load, mantle properties and lithospheric thickness (after NAKADA & LAMBECK, 1987). Observed sea levels (closed circles) for NW Tuamotu archipelago (from PIRAZZOLI & MONTAGGIONI, 1986).

Holocene reefs at Barbados (FAIRBANKS, 1989) and the Huon Peninsula, New Guinea (CHAPPELL & POLACH, 1991). In New Guinea reef growth kept pace with sea-level rise from at least 11.1 ka BP, showing mean and maximum reef growth rates of 10 m ka^{-1} and 13 m ka^{-1} respectively (CHAPPELL & POLACH, 1991). Thereafter, there is widespread evidence through the south-central Pacific for a higher-than-present Holocene sea level of ~ ± 1.0 m between ~ 6,000 years BP and ~ 2,000 years BP (Fig. 18; PIRAZZOLI & MONTAGGIONI, 1988a, 1988b). Geophysical models, 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, suggest that the Pacific high stand at ~ 6,000 years BP was largely due to the control of mantle viscosity, with the potential contribution of Antarctic ice controlling the "peakedness" of the event (LAMBECK & NAKADA, 1985; LAMBECK & NAKIBOGLU, 1986; NAKADA & LAMBECK, 1987). 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 (Fig. 19). These differences may be partly due to on-going tectonic factors.

PIRAZZOLI and MONTAGGIONI (1985) have argued that the sealevel curve from the N.W. Tuamotu archipelago, which indicates mean sea level at + 0.8 m between ca. 5 and 1.2 ka B.P., represents the true eustatic level for this period and that regional deviations from this curve reflect vertical movements associated with island subsidence (see above; in addition, a subsidence rate of 0.14 - 0.15 mm yr^{-1} over 5,000 years has been suggested for Moorea-Tahiti) or lithospheric flexure. At Mangaia, Late Holocene sea-levels were considerably higher than +0.8 m, reaching +1.3 m at 5 ka BP and 1.7m at 3.4 BP (WOODROFFE *et al.*, 1990). On Mitiaro, emergent reef flats reaching + 1.23 m (STODDART *et al.*, 1990) have been dated to 5.6 - 3.1 ka BP. Similar deposits are found on Mauke and erosional notches and benches on Atiu may also be contemporaneous with this high sealevel stand (WOODROFFE *et al.*, 1990). Elsewhere in the Cook Islands, outside the area affected by flexure effects, a sea-level of +0.5 m between 4.4 and 2.4 ka BP has been recorded (SCOFFIN *et al.*, 1985). Like the longer-term record, flexure effects may also have affected the Holocene sea-level record in linear island chains, as has been argued by PIRAZZOLI (1983) for the Society Islands. Finally, patterns of differential uplift may result from plate thermal rejuvenation. Thus in the Austral Islands higher-than-expected sealevel features at Rurutu can be compared to sealevel indicators on the neighbouring island of Tubuai (Fig. 20); rates of uplift at Rurutu have been calculated at 17 cm ka^{-1} for the Late Holocene (PIRAZZOLI & SALVAT, 1992).

All the above examples show that there is no single Late Holocene sea-level in the South Pacific Ocean that neo-tectonic contributions need to be assessed at individual localities. This assessment is likely to be complicated, however, by the imprint of ocean dynamical effects. Thus, for example, the average sea surface in the western Pacific is ~ 40 cm higher than the corresponding surface in the eastern Pacific because of the wind stress of the easterly Tradewinds. During ENSO events, however, this differential is lost; this change can be seen in tide gauge records and is recorded in reef flat corals on equatorial atolls (WOODROFFE & Mc LEAN, 1990). ENSO events provide many signals in the environmental record; although a full discussion is beyond the scope of this review, a few examples can be given. Strong ENSO events lead directly

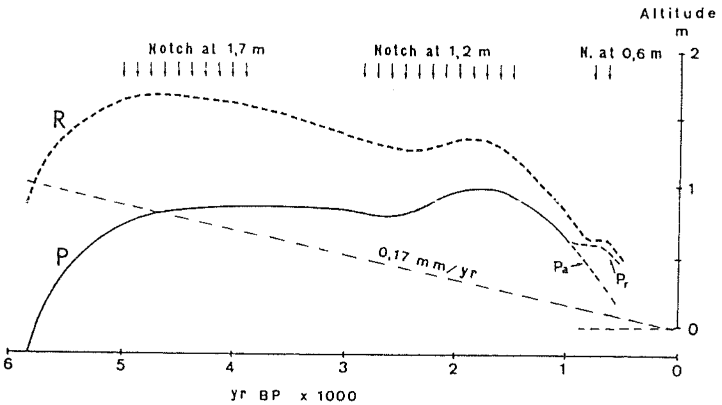

Fig. 20 : Late Holocene relative sea-level changes at Rurutu (R) and Tubuai (T), Austral Islands. Curve R is derived from P by the addition of an uplift rate of 0.17 mm a^{-1}. (after PIRAZZOLI & SALVAT, 1992).

to thermal stress in east Pacific corals, leading to coral bleaching, mass coral mortality (GLYNN & D'CROZ, 1990) and indirectly, through increased storm activity over a warmer eastern - central Pacific, to increased storm impacts on reef islands (e.g. HARMELIN-VIVIEN & LABOUTE, 1986); both types of impact suppress coral reef growth. The use of corals as recorders of the changes in rainfall associated with these circulation system reversals is a growing area of research (e.g. LEA *et al.*, 1989; COLE & FAIRBANKS, 1990; SHEN *et al.*, 1992). ENSO events also result in periodic disturbance to vegetation structure, on both low coral atolls (STODDART & WALSH, 1992) and high volcanic islands (PARKES *et al.*, 1992); here the challenge is to isolate these effects from the record of human disturbance (e.g. Easter Island : FLENLEY & KING, 1984).

CONCLUSIONS

This review has aimed to show that the ocean basins cannot be seen as stable and passive recorders of environmental change. The difficulties of disentangling eustatic, isostatic and tectonic processes in explaining sea-level variations which are well-known from other, more intensively-studied parts of the globe apply equally well to the islands of the Pacific Ocean. Furthermore, an ocean basin of the size and extent of the Pacific introduces further ocean dynamic effects which must also be considered (and which are only touched upon in this paper).

Understanding all these processes, and accounting for their effects, is important because the interactions between ocean and atmosphere in this vast region, and their variability over historical timescales, have ramifications far beyond the Pacific basin itself. This is well seen in the "teleconnections" between ENSO events in the Pacific basin and climatic variability at higher latitudes. On longer timescales the Ocean plays an important long-term role in modulating world climate : the Pacific is a major heat sink and a major store of carbon dioxide which, in a key zone near the equator, is one of the few places on the globe where the deep store of CO_2 reaches the surface and interacts with the atmosphere. Models which invoke changes in nutrient supply or ocean mixing in equatorial zones, thereby altering ocean productivity, have proved to be particularly attractive in providing an explanation for the rapid fluctuations in atmospheric CO_2 at glacial/interglacial boundaries (e.g. SARMIENTO & TOGGWEILER, 1984; SIEGENTHALER & WENK, 1984). Furthermore, shifts in the locus of calcium carbonate deposition from deep to shallow waters, which must be correlated with sea-level changes, may also play a part in explaining the shifts in atmospheric CO_2 concentrations (OPDYKE & WALKER, 1992).

Finally, on a more immediate level, an understanding of past processes and products of environmental change is important in the context of future environmental impacts. Of particular concern is the fate of islands of unconsolidated carbonate sediments on reef platforms subject to greenhouse gas-induced sea-level rise (STODDART, 1990). The Holocene record provides an analogue for island responses to higher-than-present sea-levels in the past and shows that Pacific reef islands will be protected, at least in the initial stages of sea level change, by fossil reef flats, cemented beachrock ramparts and lithified boulder beach conglomerates.

REFERENCES

ADAMS, C.G., 1970. A reconsideration of the East Indian Letter Classification of the Tertiary. *Bull. Brit. Mus. (Nat. Hist.) Geol.*, 19, 87-137.

ALLAN, J.R., & MATTHEWS, R.K., 1982. Isotope signatures associated with early meteoric diagenesis. *Sedimentology*, 29, 797-817.

BARBER, R.T. & CHAVEZ, F.P., 1986. Ocean variability in relation to living resources during the 1982-83 El Niño. *Nature*, 319, 279-285.

BARD, E., HAMELIN, B. & FAIBANKS, R.G., 1990. U-Th ages obtained by mass spectrometry in corals from Barbados : sea level during the past 130,000 years. *Nature*, 346, 456-458.

BARRON, E.J. & PETERSON, W.H., 1990. Model simulation of the Cretaceous ocean circulation. *Science*, 244, 684-686.

BARRON, E.J. & PETERSON, W.H., 1991. Reply to FÖLLMI K.B. and DELAMETTE, M., 1991. Model simulation of mid-Cretaceous ocean circulation. *Science*, 251, 94-95.

BERGGREN, W.A., KENT, D.V., FLYNN, J.J. & VAN COUVERING, J.A., 1985. Cenozoic geochronology. *Bull. Geol. Soc. Amer.*, 96, 1407-1418.

BLOOM, A.L., BROECKER, W.S., CHAPPELL, J., MATTHEWS, R.K. & MESOLELLA K.J., 1974. Quaternary sea level fluctuation on a tectonic coast : New $^{230}Th/^{234}U$ dates from the Huon Peninsula, New Guinea. *Quat. Res.*, 4, 185-205.

BOURROUILH-LE JAN, F.G., 1985. An attempt at datation by isotopic stratigraphy of Holocene and Pleistocene notches and deposits of some so-called uplifted atolls. *Proc. Fifth Int. Coral Reef Congr., Tahiti*, 3, 125-130.

BRAITHWAITE, C.J.R., 1980. The petrology of oolitic phosphorites from Esprit (Aldabra), western Indian Ocean. *Phil. Trans., Roy. Soc. Lond.*, 288B, 511-543.

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.

BURKE, W.H., DENISON, R.E., HETHERINGTON, E.A., KOEPNICK, R.B., NELSON, H.F. & OTTO, J.B., 1982. Variation of seawater $^{87}Sr/^{86}Sr$ throughout Phanerozoic time. *Geology*, 10, 516-519.

CALMANT, S. & CAZENAVE, A., 1986. The effective elastic lithosphere under the Cook-Austral and Society Islands. *Earth Planet. Sci. Lett.*, 77, 187-202.

CALMANT, S & CAZENAVE, A., 1987. Anamolous elastic thickness of the oceanic lithosphere in the south-central Pacific. *Nature*, 328, 236-238.

CAMPBELL, J.F., 1984. Rapid subsidence of Kohala volcano and its effect on coral reef growth. *Geo Marine Lett.*, 4, 31-36.

CANE, M.A., 1986. El Niño. *Ann. Rev. Earth Planet. Sci.*, 14, 43-70.

CAPO, R.C. & DE PAOLO, D.J., 1990. Seawater strontium isotopic variations 2.5 Ma to the Present. *Science*, 249, 51-55.

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, 553-570.

CHAPPELL, J., 1983. A revised sea level record for the last 350,000 years from Papua New Guinea, *Search*, 4, 99-101.

CHAPPELL, J. & POLACH, H., 1991. Post-glacial sea-level rise from a coral record at Huon Peninsula, Papua New Guinea, *Nature*, 349, 147-149.

CHAPPELL, J. & SHACKLETON, N.J., 1986. Oxygen isotopes and sea level. *Nature*, 324, 137-140.

CHAPPELL, J. & VEEH, H.H., 1978. Late Quaternary tectonic movements and sea level changes at Timor and Atuaro Island. *Bull. Geol. Soc. Amer.*, 89, 356-368.

CLAGUE, D.A. & DALRYMPLE, G.B., 1987. The Hawaiian - Emperor volcanic chain. *U.S.G.S. Prof. Pap.*, 1350, 5-49.

COLE, J.E. & FAIRBANKS, R.G., 1990. The Southern Oscillation recorded in the $\delta^{18}O$ of corals from Tarawa Atoll. *Paleoceanography*, 5, 669-683.

COUDRAY, J. & MONTAGGIONI, L.F., 1983. Coraux et récif coralliens de la province indopacifique : répartition géographique et altitudinale en relation avec la tectonique globale. *Bull. Soc. Geol. Fr.*, 24, 981-993.

CROUGH, S.T., 1978. Thermal origin of mid-plate hot-spot swells. *Geophys. J. Roy. Astron. Soc.*, 55, 451-469.

DANA, J.D., 1890. *Characteristics of volcanoes*. University Press, John Wilson, Cambridge, 399 p.

DARWIN, C.D., 1842. *The Structure and Distribution of Coral Reefs*. Smith Elder, London.

DAVIS, W.M., 1928. The coral reef problem. *Amer. Geogr. Soc. Spec. Publ.*,9, 1-596.

DETRICK, R.S., & CROUGH, S.T., 1978. Island subsidence, hot spots, and lithospheric thinning. *J. Geophys. Res.*, 83, 1236-1244.

DUBOIS, J., LAUNAY, J. & RECY, J., 1975. Some new evidence on lithospheric bulges close to island arcs. *Tectonophys.*, 26, 189-196.

DUNCAN, R.A. & CLAGUE, D.A., 1985. Pacific plate motion recorded by linear volcanic chains, In : Nairn AEM, F.G. STEHLI and S. UYEDA (Eds.), *The Ocean Basins and their margins : The Pacific Ocean* Vol. 7A, Plenum, New York, pp. 89-121.

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

EMERY, K.O., TRACEY, J.I. & LADD, H.S., 1954. Geology of Bikini and nearby atolls *U.S.G.S. Prof. Pap.*, 260-A, 1-265.

FAIRBANKS, R.G., 1989. A 17,000 year glacio-eustatic sea-level record influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature*, 342, 637-642.

FARRELL, W.E., & CLARK, J.A., 1976. On postglacial sea level. *Geophys. J.R. Astron Soc.*, 46, 647-667.

FISCHER, K.M., McNUTT, M.K. & SHURE, L., 1986. Thermal and mechanical constraints on the lithosphere beneath the Marquesas swell. *Nature*, 322, 733-736.

FLENLEY, J.R. & KING, S.M., 1984. Late Quaternary pollen records from Easter Island. *Nature*, 307, 47-50.

FÖLLMI, K.B. & DELAMETTE, M., 1991. Model circulation of mid-Cretaceous ocean circulation. *Science*, 251-94.

FOSBERG, F.R., 1957. Description and occurrence of atoll phosphate rock in Micronesia. *Amer. J. Sci.*, 225, 584-592.

GLYNN, P.W. & D'CROZ, L., 1990. Experimental evidence for high temperature stress as the cause of El Nino-coincident coral mortality. *Coral Reefs*, 8, 181-191.

GRAY, S.C., 1991. Comment on "Post-Miocene diagenetic and eustatic history of Enewetak Atoll : Model and data comparison". *Geology*, 19, 956-957.

GRIGG, R.W., 1982. Darwin Point : a threshold for atoll formation. *Coral Reefs*, 1, 29-34.

GRIGG, R.W., 1988. Paleoceanography of coral reefs in the Hawaiian-Emperor chain. *Science*, 240, 1737-1743.

GRIGG, R.W. & EPP, D., 1989. Critical depth for the survival of coral islands : Effects on the Hawaiian archipelago. *Science*, 243, 638-641.

GRIGG, R.W. & HEY, R., 1992. Paleoceanography of the tropical eastern Pacific Ocean. *Science*, 255, 172-178.

HAMILTON, W., 1979. Tectonics of the Indonesian Region. *U.S.G.S., Prof. Pap.*, 1078, 1-345.

HAQ, B.U., HARDENBOL, J. & VAIL, P.R., 1987. Chronology of fluctuating sea levels since the Triassic. *Science*, 235, 1156-1167.

HAQ, B.U., HARDENBOL, J. & VAIL, P.R., 1988. Mesozoic and Cenozoic chronostratigraphy and eustatic cycles. In : C.K., WILGUS, B.S.,HASTINGS, H., POSAMENTIER, J. VAN WAGONER, C.A. ROSS & C.G.S.C. KENDALL (Eds.) *Sea-level changes : an integrated approach.* Soc. Econ. Pal. Miner., Spec. Publ., Tulsa, Okla., 42, 71-108.

HARMELIN-VIVIEN, M.L. & LABOUTE, P., 1986. Catastrophic impact of hurricanes on atoll outer reef slopes in the Tuamotu (French Polynesia). *Coral Reefs*, 5, 55-62.

HODELL, D.A., MUELLER, P.A. & MEAD, J.R., 1991. Variations in strontium isotopic composition of seawater during Neogene. *Geology*, 19, 24-27.

HUTCHINSON, J.E., 1950. Survey of existing knowledge of vertebrate biochemistry. 3. The biochemistry of vertebrate excretion. *Bull. Amer. Mus. Hist.*, 96, 1-554.

INGLE, J.C., 1981. Origin of neogene diatomites around the North Pacific rim. In : R.E. GARRISON & co-editors, *The Monterey formation and related siliceous rocks of California*. Soc. Econ. Palaeont. Mineral., Spec. Publ., 159-180.

JACKSON, E.D., SILVER, E.A. & DALRYMPLE, G.B., 1972. Hawaiian - Emperor chain and its relation to Cenozoic circumpacific tectonics. *Bull. Geol. Soc. Amer.* 83, 601-618.

JANECEK, T.R. & 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 biostatigraphic evidence from nearby Pacific Ocean and Caribbean cores. *Geology*, 6, 630-634.

KEIGWIN, L.D. & 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 p.

KENNETT, J.P., KELLER, G. & SRINIVASAN, M.S., 1985. Miocene planktonic foraminiferal biogeography and palaeoceanography development of the Indo-Pacific region. In : J.P. KENNETT (Ed.), *The Miocene Ocean : palaeoceanography and Biogeography*. Geol. Soc. Amer. Mem., 163, pp. 197-136.

KOEPNICK, R.B., BURKE, W.H., DENISON, R.E., HETHERINGTON, E.A., NELSON, H.F., OTTO, J.B. & WAITE, L.E., 1985. Construction of the seawater $^{87}Sr/^{86}Sr$ curve for the Cenozoic and Cretaceous : supporting data. *Chem. Geol.*, 58, 55-81.

KOEPNICK, R.B., DENISON, R.E. & DAHL, D.A., 1988. The Cenozoic sea-water Sr-87/Sr-86 curve : data review and implications for correlation of marine strata. *Paleoceanography*, 3, 743-756.

LADD, H.S. & SCHLANGER, S.O., 1960. Drilling operations at Eniwetok Atoll. *U.S.G.S. Prof. Pap.*, 260-Y, 863-903.

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

LAMBECK, K. & 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. & NAKIBOLU, S.M., 1986. Recent sea level changes in the South-West Pacific In : W. REILLY & B.E. HARTFORD (Eds.), Recent Crustal Movements of the Pacific Region. *Bull. Roy. Soc. N.Z.*, 24, pp. 485-491.

LE PICHON, S., FRANCHETEAU, J. & BONIN, J., 1973. *Plate tectonics*. Elsevier : New York.

LEA, D.W., SHEN, G.T. & BOYLE, E.A., 1989. Coralline barium records temporal variability in Equatorial Pacific upwelling. *Nature*, 340, 373-376.

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

LINCOLN, J.M. & SCHLANGER, S.O., 1987. Miocene sea-level falls related to the geologic history of Midway Atoll. *Geology*, 15, 454-457.

LINCOLN, J.M. & SCHLANGER, S.O., 1991. Atoll stratigraphy as a record of sea level changes : Problems and prospects. *J. Geophys. Res.*, 96, 6727-6752.

LINCOLN, J.M., SCHLANGER, S.O., BERGERSON, D., COOGAN, A.H. & PERMOLI SILVA, L., 1989. Cretaceous foundations and multiple periods of uplift in the northern Marshall Islands, west central Pacific. *Geol. Soc. Amer. Abstr. with Programs*, 21, A195.

LIPMAN, P.W., NORMARK, W.R., MOORE, J.G., WILSON, J.B. & GUTMACHER, C.E., 1988. The giant submarine Alika debris slide, Mauna Loa, Hawaii. *J. Geophys. Res.*, 93, 4279-4299.

LUDWIG, K.R., HALLEY, R.B., SIMMONS, K.R. & PETERMAN, Z.E., 1988. Strontium isotope stratigraphy of Enewetak Atoll. *Geology*, 16, 173-177.

LUDWIG, K.R., MUHS, D.R. & SIMMONS, K.R., 1992. Sr-isotope record of Quaternary marine terraces on the California coast and off Hawaii. *Quat. Res.* 37, 267-280.

LUDWIG, K.R., SZABO, B.J., MOORE, J.G. & SIMMONS, K.R., 1991. Crustal subsidence rate off Hawaii determined from $^{234}U/^{238}U$ ages of drowned coral reefs. *Geology*, 19, 171-174.

LYLE, M.W., PRAHL, F.G. & SPARROW, M.A., 1992. Upwelling and productivity changes inferred from a temperature record in the central equatorial Pacific. *Nature*, 355, 812-815.

McNUTT, M. & MENARD, H.W., 1978. Lithospheric flexure and uplifted atoll. *J. Geophys. Res.*, 83, 1206-1212.

MAJOR, R.P. & MATTHEWS, R.K., 1983. Isotopic composition of bank margin carbonates on Midway Atoll : amplitude constraints on post-early Miocene eustasy. *Geology*, 11, 335-338.

MARK, R.K. & MOORE, J.G., 1987. Slopes of the Hawaiian Ridge. *U.S.G.D. Prof. Pap.*, 1350, 101-107.

MENARD, H.W., 1982. Influence of rainfall upon the morphology and distribution of atolls, In : R.A. SCRUTTON & H. TALWANI (Eds.), *The Ocean Floor*, Wiley, New York, pp. 305-311.

MOLINA-CRUZ, A., 1977. The relation of the southern tradewinds to upwelling processes during the last 75,000 years. *Quat. Res.*, 8, 324-338.

MILLER, K.G., WRIGHT, J.D. & FAIRBANKS, R.F., 1991. Unlocking the ice-house : Oligocene-Miocene oxygen isotopes, eustasy and margin erosion. *J. Geophys. Res.*, 96, 6829-6848.

MONTAGGIONI, L.F., 1988. Holocene reef growth history in mid-plate high volcanic islands. *Proc. Sixth Int. Coral Reef. Symp. Townsville*, 3, 455-460.

MONTAGGIONI, L.F., 1989. Le soulèvement polyphasé d'origine volcano-isostasique : clef de l'évolution post-oligocène des atolls du Nord-Ouest des Tuamotus (Pacifique central). *C.R. Acad. Sci. Paris*, 309 (II), 1591-1598.

MONTAGGIONI, L.F. *et al.*, 1985. Les récifs coralliens de l'île de Makatea, archipel des Tuamotu, Pacifique Central : géomorphologie et répartition des peuplements. *Ann. Inst. Oceanogr., Paris*, 61, 1-15.

MOORE, C.H., 1989. *Carbonate diagenesis and porositty*, Elsevier, New York.

MOORE, J.G., 1970. Relationship between subsidence and volcanic load, Hawaii. *Bull. Volcanologie*, 34, 562-576.

MOORE, J.G., 1987. Subsidence of the Hawaiian Ridge. *U.S.G.S., Prof. Pap.*, 1350, 85-100.

MOORE, J.G. & CAMPBELL, J.F., 1987. Age of tilted reefs, Hawaii. *J. Geophys. Res.*, 92, 2641-2646.

MOORE, J.G. & CLAGUE, D.A., 1992. Volcano growth and evolution of the island of Hawaii. *Bull. Geol. Soc. Amer.*, 104, 1417-1484.

MOORE, J.G. & FONARI, D.J., 1984. Drowned reefs as indicators of the rate of subsidence of the island of Hawaii. *J. Geol.*, 92, 752-759.

MOORE, J.G. & MOORE, G.W., 1984. Deposit from a giant wave on the island of Lanai. *Science*, 226, 1312-1315.

MOORE, J.G., CLAGUE, D.A., HOLCOMB, R.T., LIPMAN, P.W., NORMARK, W.R. & TORRESAN, M.E., 1989. Prodigious submarine landslides on the Hawaiian Ridge. *J. Geophys. Res.*, 94, 17465-17484.

MOORE, J.G., CLAGUE, D.A., LUDWIG K.R. & MARK, R.K., 1990. Subsidence and vulcanism of the Haleakala Ridge, Hawaii. *J. Vulcanol. Geotherm. Res.*, 42, 1-11.

MOORE, T.C. et al., 1981. The biological record of the ice-age ocean. *Palaeogeog. Palaeoclimatol. Palaeoecol.*, 35, 357-370.

MORHANGE, C., 1990. Mise en évidence d'une déformation de la lithosphère à partir de l'étude quantitative de l'évolution géomorphologique des îles Australes et Cook du sud en Polynésie. *Rev. Geomorph. Dynam.*, 39, 1-13.

MORHANGE, C., 1992. Essai de quantification de l'évolution géomorphologique d'un archipel volcanique tropical né d'un point chaud : le cas des îles de la Société en Polynésie française. *Zeitschr. für Geomorph., NF.*, 36, 307-324.

NAKADA, M. & 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. & AHARON, P., 1983. Postglacial sea level in the Pacific : implications with respect to deglaciation regime and local tectonics. *Tectonophys*. 91, 335-338.

NUNN, P., 1986. Implications of migrating geoid anomalies for the interpretation of high-level fossil coral reefs. *Bull. Geol. Soc. Amer.*, 97, 946-952.

OHDE, S. & ELDERFIELD, H., 1992. Strontium isotope stratigraphy of Kita-daito-jima Atoll, North Philippine Sea : implications for Neogene sea-level change and tectonic history. *Earth Planet Sci. Lett.*, 113, 473-486.

OPDYKE, B.N. & WALKER, J.C.G., 1992. Return of the coral reef hypothesis : Basin to shelf partitioning of $CaCO_3$ and its effect on atmospheric CO_2. *Geology*, 20, 733-736.

OTA, Y., 1986. Marine terraces as reference surfaces in late Quaternary tectonic studies : examples from the Pacific rim. *Bull. Roy. Soc. N.Z.*, 24, 257-375.

OTA, Y. & KAIZUKA, s., 1991. Tectonic geomorphology at active plate boundaries. Examples from the Pacific rim. *Zeitschr für Geomorph., Suppl. Bd.,* 82, 119-146.

PARKES, A., TELLER, J.T. & FLENLEY, J.R., 1992. Environmental history of the Lake Vaihiria drainage basin, Tahiti, French Polynesia, *J. Biogeog.*, 19, 431-447.

PARSONS, B. & SCLATER, J.G., 1977. An analysis of the variation of ocean floor bathymetry and heat flow with age. *J. Geophys. Res.*, 82, 803-827.

PEDERSEN, T.F., 1983. Increased productivity in the eastern equatorial Pacific during the last glacial maximum (19,000 to 14,000 yr B.P.). *Geology*, 11, 16-19.

PHILANDER, S.G.H., 1990. *El Niño, La Niña and the Southern Oscillation.* Academic Press, San Diego, 1-293.

PIRAZZOLI, P.A., 1983. Mise en évidence d'une flexure active de la lithosphère dans l'archipel de la Société (Polynésie française) d'après la position des rivages de la fin de l'Holocène. *C.R. somm. Séances. Acad. Sci. Paris,* 296, 695-698.

PIRAZZOLI, P.A. et al., 1993. A one million-year-long sequence of marine terraces on Sumba Island, Indonesia. *Mar. Geol.*, 109, 221-236.

PIRAZZOLI, P.A. & 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. & 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., 1988a. Holocene sea-level changes in French Polynesia. *Palaeogeog. Palaeoclimatol. Palaeoecol.*, 68, 153-175.

PIRAZZOLI, P.A. & MONTAGGIONI, L.F., 1988b. The 7000 year sealevel curve in French Polynesia : geodynamic implications for mid-plate volcanic islands. *Proc. Sixth Int. Coral Reef. Symp., Townsville*, 3, 467-472.

PIRAZZOLI, P.A. & SALVAT, B., 1992. Ancient shorelines and quaternary vertical movements on Rurutu and Tubuai (Austral Isles, French Polynesia). *Zeitschr. für Geomorph., NF,* 36, 431-451.

PIRAZZOLI, P.A. & VEEH, H.H., 1987. Age $^{230}Th/^{234}U$ d'une encoche émergée et vitesses de soulèvement quaternaire à Rurutu, îles Australes. *C.R. Acad. Sci. Paris*, 305 (II), 919-923.

PORTER, S.C., 1979. Hawaiian glacial ages. *Quat. Res.*, 12, 161-187.

PORTER, S.C., 1987. Pleistocene subglacial eruption on Mauna Kea. *U.S.G.S. Prof. Pap.* 1350, 587-598.

QUINN, T.M., 1991a. Reply to Gray, "Post-Miocene diagenetic and eustatic history of Enewetak Atoll : Model and data comparison", *Geology*, 19, 957-958.

QUINN, T.M., 1991b. The history of post-Miocene sea level change : Inferences from stratigraphic modeling of Enewetak Atoll. *J. Geophys. Res.* 96, 6713-6725.

QUINN, T.M. & MATTHEWS, R.K., 1990. Post-Miocene diagenetic and eustatic history of Enewetak Atoll : Model and data comparison. *Geology*, 18, 942-945.

QUINN, T.M., LOHMANN, K.C. & HALLIDAY, A.N., 1991. Sr isotopic variation in shallow water carbonate sequences : stratigraphic, chronostratigraphic, and eustatic implications of the record at Enewetak Atoll. *Paleoceanography*, 6, 371-385.

REA, D.K. & 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. & JANECEK, T.R., 1982. Late Cenozoic changes in atmospheric circulation deducted from North Pacific eolian sediments. *Marine Geol.*, 49, 149-167.

ROE, K.K. & BURNETT, W.C., 1985. 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. *Mar. Micropalaeontol.*, 7, 163-187.

ROMINE, K., & MOORE, T.C., 1981. Radiolarian assemblage distribution and palaeoceanography of the eastern equatorial Pacific Ocean during the last 127,000 years. *Palaeogeog., Palaeoclimatol. Palaeoecol.*, 35, 281-314.

ROSEN, B.M., 1982. Darwin, coral reefs and global geology. *Bioscience*, 32, 519-525.

SALLER, A.H. & KOEPNICK, R.B., 1990. Eocene to Early Miocene growth of Enewetak Atoll : Insights from strontium isotope data. *Bull. Geol. Soc. Amer.*, 102, 381-390.

SARMIENTO, J.L. & TOGGEWEILER, R., 1984. A new model for the role of the oceans in determining atmospheric CO_2. *Nature*, 308, 621-624.

SAVIN, S.M. et al., 1981. Miocene benthic foraminiferal isotope records : A synthesis. *Mar. Micropalaeontol.,* 6, 423-450.

SCHLANGER, S.O., 1963. Subsurface geology of Eniwetok Atoll. *U.S.G.S. Prof. Pap.* 260-BB, 991-1066.

SCHLANGER, S.O. & PREMOLI SILVA, S., 1986. Oligocene sea-level falls recorded in mid-Pacific atoll and archipelagic apron setting. *Geology,* 14 ; 392-395.

SCHRADER, H. & SORKNES, R., 1991. Peruvian coastal upwelling : Late Quaternary productivity change revealed by diatoms. *Mar. Geol.*, 97, 233-249.

SCOFFIN, T.P. & DIXON, J.E., 1983. The distribution and structure of coral reefs : one hundred years since Darwin. *Biol. J. Linnean Soc.*, 20, 11-38.

SCOFFIN, T.P., STODDART, D.R., TUDHOPE, A.W. & WOODROFFE, C.D., 1985. Exposed limestones of Suwarrow Atoll. *Proc. Fifth Int. Coral Reef Symp., Tahiti*, 3, 137-140.

SHACKLETON, N.J., 1984. Oxygen isotope evidence for Cenozoic climatic change. In : P.J. BRENCHLEY (Ed.), *Fossils and climate*, Wiley, Chichester, pp. 27-34.

SHACKLETON, N.J. & 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 : J.P. KENNETT *et al., Initial Reports of the Deep Sea Drilling Project*, *29*, U.S. Govt. Printing Office : Washington D.C., pp. 743-755.

SHACKLETON, N.J. & OPDYKE, N.D., 1973. Oxygen isotope and palaeomagnetic stratigraphy of equatorial Pacific core V28-238 : Oxygen isotope temperatures and ice volumes on a 106 year scale. *Quat. Res.*, 3, 39-55.

SHEN, G.T., *et al.*, 1992. Surface ocean variability at Galapagos from 1936-1982 : Calibration of geochemical tracers in corals. *Paleoceanography*, 7, 563-588.

SIEGENTHALER, U. & WENK, T., 1984. Rapid atmospheric CO_2 variations and ocean circulation. *Nature*, 308, 624-626.

SPENCER, T., 1983. Limestone erosion rates and micro-topography : Grand Cayman Island, British West Indies. Unpublished PhD thesis, Univ. of Cambridge.

SPENCER, T., 1985. Rates of karst processes on raised reef limestones and their implications for coral reef histories. *Proc. Fifth Int., Coral Reef Congr., Tahiti*, 6, 629-634.

SPENCER, T., 1989. Tectonic and environmental histories in the Pitcairn Group, Palaeogene to Present : Reconstructions and speculations. *Atoll Research Bulletin*, 322, 1-41.

SPENCER, T., STODDART, D.R. & WOODROFFE, C.D., 1987. Island uplift and lithospheric flexure; observations and cautions from the South Pacific. *Zeitsch. für Geomorph. NF., Suppl. Bd.*, 63, 87-102.

SPENCER, T., STODDART, D.R., WOODROFFE, C.D. & HARMON, R.S., 1988. Lithospheric flexure and raised limestones, S Cook-Austral Islands. *Proc. Sixth Int. Coral Reef Symp., Townsville*, 3, 485-489.

STEARNS, H.T. 1978. Quaternary shorelines in the Hawaiian Islands. *Bull. Bernice P. Bishop Mus.*, 237, 1-57.

STODDART, D.R., 1976. Continuity and crisis in the reef community. *Micronesica*, 12, 1-9.

STODDART, D.R., 1990. Coral reefs and islands and predicted sea-level rise. *Prog. Phys. Geog.*, 14, 521-536.

STODDART, D.R., 1992. Biogeography of the Tropical Pacific. *Pacific Science*, 46, 276-293.

STODDART, D.R. & SCOFFIN, T.P., 1983. Phosphate rock on coral reef islands. In : A.S. GOUDIE & K. PYE (Eds.) *Chemical Sediments and Geomorphology*, Academic Press, London, pp. 369-400.

STODDART, D.R. & SPENCER, T., 1987. Rurutu reconsidered : the development of *makatea* topography in the Austral Islands. *Atoll Research Bulletin*, 297, 1-19.

STODDART, D.R. & WALSH, R.P.D., 1992. Environmental variability and environmental extremes as factors in the island ecosystem. *Atoll Research Bulletin*, 356, 1-71.

STODDART, D.R., SCOFFIN, T.P. & SPENCER, T., 1985. Reef growth and karst erosion of Mangaia, Cook Islands : a reinterpretation. *Zeitsch. für Geomorph. Suppl. Bd.*, 57, 121-140.

STODDART, D.R., WOODROFFE, C.D. & SPENCER, T., 1990. Mauke, Mitiaro and Atiu : Geomorphology of Makatea Islands in the Southern Cooks. *Atoll Research Bulletin*, 341, 1-66.

SZABO, B.J. & MOORE, J.G., 1986. Age of - 360 m reef terrace, Hawaii and the rate of late Pleistocene subsidence of the island. Geology, 14, 967-968.

SZABO, B.J., TRACEY, J.I. & GOTER, E.R., 1985. Age of stratigraphic intervals in the Quaternary of Enewetak Atoll, Marshall Islands. *Quat. Res.*, 23, 54-61.

TAYLOR, F.W. & BLOOM A.L., 1977. Coral reefs on tectonic blocks, Tonga Island arc. *Proc. Third Int. Coral Reef Symp.*, 2, 275-281.

TEN BRINK, U.S. & WATTS, A.B., 1985. Seismic stratigraphy of the flexural moat flanking the Hawaiian Islands. *Nature*, 317, 421-424.

TRACEY, J.I., 1980. Quaternary episodes of insular phosphatization in the central Pacific. In : R.S. SHELDON and W.C. BURNETT, (Eds.), *Fertilizer mineral potential in Asia and the Pacific*, E-W. Center, Honolulu, pp. 197-205.

TRUDGILL, S.T., 1976. The subaerial and subsoil erosion of limestones on Aldabra Atoll, Indian Ocean. *Zeitsch für Geomorph., Suppl. bd.*, 26, 201-210.

TURNER, D.L. & 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.H., HEATH, G.R. & MOORE, T.C., 1975. Cenozoic history and palaeoceanography of the Central Equatorial Pacific Ocean. *Mem. Geol. Soc. Amer.*, 143, 1-134.

VEEH, H.H., 1966. $^{230}Th/^{238}U$ and $^{234}U/^{238}U$ ages of Pleistocene high sea level stand. *J. Geophys. Res.*, 71, 3379-3386.

VEEH, H.H., 1985. Uranium-series dating applied to phosphate deposits on coral reef islands. *Proc. Fifth Int. Coral Reef Congr., Tahiti*, 3, 463-469.

WARDLAW, B.R. & QUINN, T.M., 1991. The record of Pliocene sea-level change at Enewetak Atoll. *Quat. Sci. Revs.*, 10, 247-258.

WATTS, A.B. & TEN BRINK, U.S., 1989. Crustal structure, flexure, and subsidence history of the Hawaiian Islands. *J. Geophys. Res.*, 94, 10473-10500.

WATTS, A.B., BODINE, J.H. & RIBE, N.M., 1980. Observations of the flexure and the geological evolution of the Pacific Ocean basin. *Nature*, 283, 532-537.

WATTS, A.B., TEN BRINK, U.S., BUHL, P. & BROCHER, T.M., 1985. A multichannel seismic study of lithospheric flexure across the Hawaiian - Emperor seamount chain. *Nature*, 315, 105-111.

WENTWORTH, C.K., 1925. The geology of Lanai. *Bull. Bernice P. Bishop Mus.*, 24, 1-72.

WENTWORTH, C.K., 1927. Estimates of marine and fluvial erosion in Hawaii. *J. Geol.* 35, 117-133.

WHEELER, C.W. & AHARON, P., 1991. Mid-oceanic carbonate platforms as oceanic dipsticks : examples from the Pacific. *Coral Reefs*, 10, 101-114.

WILLIAMS, C.A., 1986. An oceanwide view of Palaeogene plate tectonic events. *Palaeogeog., Palaeoclimatol., Palaeoecol.*, 57, 3-25.

WOODROFFE, C.D., & Mc LEAN, R.F., 1990. Microatolls and recent sea level change on coral atolls. *Nature*, 344, 531-534.

WOODROFFE, C.D., SHORT, S., STODDART, D.R., SPENCER, T. & HARMON, R.S., 1991. Stratigraphy and chronology of late Pleistocene reefs in the southern Cook Islands, South Pacific. *Quat. Res.*, 35, 246-263.

WOODROFFE, C.D., STODDART, D.R., SPENCER, T., SCOFFIN, T.P. & TUDHOPE, A.W., 1990. Holocene emergence in the Cook Islands. *Coral Reefs*, 9, 31-39.

WOODRUFF, F., SAVIN, S.M. & DOUGLAS, R.G., 1981. Miocene stable isotope record : a detailed Pacific Ocean study and its palaeoclimatic implications. *Science*, 212, 665-668.