GEOPHYSICS, STRUCTURE AND EVOLUTION
OF SELECTED SOUTH PACIFIC SEAMOUNTS
AND SUBMARINE PLATEAUS

by

Peter J. Hill

A thesis submitted for the degree of Master of Science
of the Australian National University.

The bulk of this thesis represents my own work; all other sources used have been acknowledged. Five of the ten sections in the thesis were prepared in collaboration with research colleagues. My contribution to these joint sections (Sections 1, 2, 3, 7 and 8) is estimated to be as follows: - Section 1 90%, Section 2 50%, Section 3 40%, Section 7 70% and Section 8 35%.

Section 1 is essentially all my own work - I planned and executed the shipboard program related to the investigation of geoid anomalies, analysed the data and wrote the text. I proposed the shipboard program that lead to the discovery of the Manihiki mud volcano complex, which is the subject of Section 2. I interpreted the geophysical data - seismic reflection, gravity and magnetics - and had major input into the bathymetric analysis and discussion on the origin of the mud volcano field. I played a leading role in the planning and execution of the SeaMARC II and geophysics investigation of Machias Seamount - Section 2, and contributed significantly to the interpretation of the data.

The gravity and magnetic fieldwork conducted on Nauru, which forms the basis of Section 7, was carried out by me. I was responsible for the reduction and interpretation of these data, as well as the bathymetric compilation, and also wrote the sections on tectonics, palaeomagnetism, evolution and discussion on subsidence.
and uplift. My input to Section 8 was mainly in the execution, interpretation and reporting of the geoelectrical sounding and gamma-ray spectrometer surveys conducted on Nauru as part of the hydrogeological investigation.

Signed -

Peter J. Hill
Eight of the ten sections in this thesis are based on the geophysical and geological data acquired during two recent geoscience research cruises in the Southwest Pacific, on which I participated as shipboard scientist. These cruises were by, (i) the New Zealand naval oceanographic ship HMNZS 'Tui' in April-May 1986 (Rarotonga - Nuku'alofa) and (ii) the Hawaii Institute of Geophysics (HIG) research ship R/V 'Moana Wave' in February-March 1987 (Pago Pago - Honolulu). Data collected in October 1987 during a geophysics and hydrogeological investigation of Nauru Island form the basis of the other two sections of the thesis.

The 'Tui' and 'Moana Wave' cruises were part of the Tripartite II Program undertaken by the governments of Australia, New Zealand and the United States of America, in co-operation with the Committee for Co-ordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas (CCOP/SOPAC) to carry out joint marine geoscientific research and mineral resource studies in the South Pacific region. The CCOP/SOPAC secretariat has its headquarters in Suva, Fiji; members of CCOP/SOPAC include the Southwest Pacific island nations of Cook Islands, Fiji, Kiribati, Papua New Guinea, Solomon Islands, Tonga, Tuvalu, Vanuatu and Western Samoa.

The 'Tui' and 'Moana Wave' research programs were aimed at two principal objectives -:

(i) to investigate potential marine mineral resources in the SW Pacific region
(ii) to extend our knowledge of the region's geological framework and tectonic development.

Because of the importance of the resource-assessment component in the programs, much of the survey work was focused on areas of elevated seafloor. Of particular interest was the distribution and environment of deposition of Co-rich ferromanganese-oxide crusts on the flanks of seamounts and Manihiki Plateau. Attempts were also made to establish whether or not the Manihiki Plateau Cretaceous volcaniclastic beds intersected by drilling at DSDP Site 317 and found to be Cu-bearing, were also metalliferous over a more extensive area. This was done by conducting geophysical investigations of selected parts of the plateau margins to identify suitable exposures of stratigraphically equivalent units and then sampling such targets by follow-up dredging. Data of considerable relevance to the regional geological framework and tectonic studies were acquired during the course of the specific resource-oriented investigations. Such data of more general geological significance were augmented by operating the geophysical systems during transit lines between major survey sites, and selecting these routes to optimize the usefulness of the data collected.

Section 1 of this thesis deals with detection of seamounts in the SW Pacific from geoid anomalies using the GEOS-3 and SEASAT satellite radar altimeter data sets, and outlines the discovery during the 'Tui' and 'Moana Wave' cruises of four previously uncharted seamounts having up to 4000 m relief.

During the 1986 'Tui' cruise an unusual piercement structure was recorded by seismic in the sediment pile of the NE Manihiki High. This feature and the adjacent area was investigated in more detail in the following year during the 'Moana Wave' expedition using amongst other techniques, air-gun seismic and the swath-mapping
SeaMARC II system, which produces both bathymetric and side-scan images of the seafloor. The feature was identified as a mud volcano about 200 m high, one in a field of about 100 such cones found scattered about a central composite edifice with relief of about 1900 m. Other known occurrences of mud volcanoes are generally restricted to the world's mobile belts, while the size of the structures is usually no more than tens of metres high - with the largest being about 400 m in the Caspian Sea area. The Manihiki mud volcano complex is unique in its sheer size and the fact that it is located in a tectonically stable mid-ocean region. Details of its morphology, geophysical expression, sampling results and discussion on possible modes of origin are presented in Section 2.

Aspects of the geophysics, geology, structure and evolution of four large SW Pacific seamounts - Machias, Capricorn, Niue and Nauru are the subject of Sections 3, 4, 5, 6, 7 and 8. These seamounts possess similar morphology, being 50-100 km in basal diameter and of 4300-5000 m relief, and consist of isolated basaltic volcanoes topped by carbonate platforms of about 500 m interpreted thickness. Machias and Capricorn seamounts are located on the seaward side of the Tonga Trench and are starting to be subducted as they ride the Pacific plate into the trench. Nauru, Niue and Capricorn seamounts probably all evolved by mantle hot-spot volcanism. Section 3 outlines the results of the 'Moana Wave' SeaMarc II survey of Machias Seamount, describing its morphology and neo-tectonic sedimentological and structural patterns imposed on the edifice as it descends into the Tonga Trench. A broader geophysical and geological examination of Machias Seamount is presented in Section 4, invoking seismic, bathymetric, gravity and magnetic data from the 'Tui' and 'Moana Wave' cruises, seafloor sampling results from the 'Tui' cruise, plus related data sets from (vi)
various previous investigations. The results of geophysical surveys and dredging investigations in the vicinity of Capricorn Seamount and Niue are presented in Sections 5 and 6, respectively. The section on Niue includes an examination of nearby seamounts Endeavour and Lachlan, and also draws on the results of a detailed geophysical survey I conducted on Niue Island in 1979. Sections 7 and 8 were produced following a request by the Commission of Inquiry into Rehabilitation of the Worked-out Phosphate Lands in Nauru to conduct an investigation of the hydrogeology of Nauru. The fieldwork was conducted on Nauru over a two week period in October 1987, and included gravity and magnetic surveys of the island to establish its structural framework; as well as geoelectrical depth probes and a gamma-ray spectrometer survey to investigate the shallow geology and hydrogeology. Section 7 deals with the structure and evolution of Nauru, while Section 8 reports on the hydrogeology and near-surface geology of the island.

The final stage of the 'Tui' cruise program involved a detailed geophysical survey of Tofua Trough, a sediment filled depression on the Tonga Ridge between the chain of active andesitic volcanoes and the Tonga platform to the east. The results of this investigation are presented in Section 9. The last part of this thesis, Section 10, examines the structurally complex eastern margin of the Manihiki High plateau in the light of geophysical and geological data collected by both the 'Tui' and 'Moana Wave' expeditions. Detailed coverage of a key area over the central section of the margin by SeaMARC II mapping provides new insight into its morphology and structure.

All of the sections in the thesis were prepared with the intention of submitting them for publication as research papers in
various journals and volumes, as presented here or in slightly modified form. A number of the sections have already been submitted for review, while some sections will be expanded by contributions from co-authors from the 'Tui' and 'Moana Wave' cruises. Some minor variations in style and layout of the sections reflect different publication requirements.
ACKNOWLEDGEMENTS

Preparation of this thesis was made possible through a grant provided by the Australian International Development Assistance Bureau (AIDAB) to enable me to complete one year's full time study and research at the Australian National University, mainly working up the results of the two Tripartite II cruises on which I participated.

I thank my principal supervisor Dr Keith Crook, supervisor Dr John Tipper (both of the Geology Department, ANU) and adviser Mr Howard Stagg (Bureau of Mineral Resources, Geology and Geophysics, Canberra) for their assistance. I am also grateful to Dr David Falvey (Chief Scientist) and Dr Neville Exon (Principal Research Scientist) of the Division of Marine Geosciences and Petroleum Geology, BMR, for supporting my involvement in this research project.

The scientific successes of the HMNZS 'Tui' and R/V 'Moana Wave' cruises are a tribute to the dedication, professionalism and skills of the ship's officers and crews, shipboard scientists and technicians. All fellow participants are heartily thanked for their contribution in the acquisition of the high quality geophysical and geological data sets which form the basis of much of this thesis. The cordiality and spirit of co-operation shown onboard ship helped to make the cruises extremely rewarding experiences both from a scientific and personal aspect.

The Government of Nauru has kindly allowed publication of the scientific results of the geophysical and hydrogeological investigations on Nauru, conducted on behalf of the Commission of Inquiry into Rehabilitation of the Worked-out Phosphate Lands in
Nauru. The co-operation and logistic support provided by the Nauru Phosphate Corporation during this investigation were much appreciated.
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SECTION 1

MARINE GEOPHYSICAL VERIFICATION OF NEW SW PACIFIC SEAMOUNTS REVEALED BY SATELLITE ALTIMETRY

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Abstract

GEOS-3/SEASAT radar altimeter images were used to detect major uncharted seamounts in the SW Pacific. Verification of the predicted seamounts was undertaken during research cruises of HMNZS 'Tui' and RV 'Moana Wave' during 1986/87. Four large seamounts were located and mapped by a variety of shipboard recording systems including 12 kHz and 3.5 kHz echo sounders, seismic, gravity, magnetics and SeaMARC II.

Of the three seamounts located in the southern Line Islands area, the largest has a height of 4030 m and basal diameter of 40 km. The fourth seamount, in the Samoan Basin, has a height of 2825 m and basal diameter of 35 km. The summits of the seamounts are located at $14^\circ\ 5.4'S\ 156^\circ\ 10.0'W$, $1^\circ\ 55.5'S\ 156^\circ\ 17.3'W$, $3^\circ\ 22.3'S\ 155^\circ\ 41.0'W$ and $12^\circ\ 25.2'S\ 166^\circ\ 27.9'W$. Ferromanganese-oxide crusts on these seamounts constitute a potential mineral resource for Co, Ni and Cu. Crust up to 5 cm thick was dredged from the largest seamount.

Despite limited success by some marine expeditions in locating uncharted seamounts from satellite altimetry, our results indicate that it is a reliable and cost-effective technique for delineating major submarine topographic features, particularly in remote and poorly-explored regions such as the SW Pacific.

Introduction

The SW Pacific is a region not well mapped by ship bathymetric surveys. Coverage is generally sparse and patchy, with large data gaps still present. In contrast, the altimeter sub-satellite tracks
are fairly uniformly and systematically distributed (see Figure 1 for example). The considerable new information available from analysis of the altimeter observations facilitates detection of previously unknown bathymetric features (Lambeck & Coleman (1982), Lazarewicz & Schwank (1982), and Baudry & others (1987)).

During Tripartite II* research cruises by HMNZS 'Tui' in 1986 and RV 'Moana Wave' in 1987 (Coulbourn, Hill & others, 1987), geoid anomalies corresponding to possible uncharted seamounts were investigated. The study was undertaken because seamounts in the central and south-west Pacific are economically significant since they host ferromanganese-oxide crusts potentially rich in Co, Ni and Cu (Halbach & others, 1982; Cronan, 1984). Four uncharted seamounts were located and mapped. The seamounts have been designated MW1, MW2, MW3 and Tl for ease of reference. The first three of these are situated in the southern Line Islands area, while the fourth (Tl) lies in the Samoan Basin (Figure 1).

* Co-operative Australian, New Zealand and USA marine geoscience research program co-ordinated by the Committee for Co-ordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas (CCOP/SOPAC), Suva.
Satellite radar altimeter missions GEOS-3 (Stanley, 1979) and SEASAT (Born & others, 1979) generated a vast data set of sea surface height observations covering most of the world's oceans. GEOS-3 and SEASAT currently provide the best estimates of geoid height over oceans. Sea surface heights as recovered from the satellite altimeter measurements are a close approximation to the marine geoid. Deviations from the geoid are produced by oceanographic effects such as currents, eddies and storms. Such perturbations generally amount to less than 0.5 m, and are typically only in the order of centimetres in the open oceans. The other contribution to local variations in the marine geoid is the gravimetric effect of lateral density variations in the crust and upper mantle. Because of the large, relatively shallow density contrast at the water/seafloor interface, seafloor topography has a major influence - hence the observed strong correlation between short-wavelength geoid anomaly and bathymetry. Seamounts, in particular, produce distinct and characteristic geoid signatures with amplitudes typically 0.5 - 3.0 m and diameters 25-100 km.

**GEOS-3/SEASAT and Seamount Detectability**

Editing, correction and adjustment procedures for recovery of sea surface heights have been documented for GEOS-3 (Rapp, 1979) and SEASAT (Rapp, 1982). Our work is based on a combined GEOS-3/SEASAT data set (Liang, 1983). This data set is constructed from smoothed altimeter signal sampled at approximately 1 second intervals. This sampling rate applies to both GEOS-3 and SEASAT and represents a 7 km data spacing on the sea surface.

The noise level for the SEASAT sea surface heights is about 10 cm and 20-50 cm for GEOS-3. Power spectra analyses indicate that the
geoid signal starts to emerge from a high-frequency noise 'floor' at wavelengths of 30 km for SEASAT and 60 km for GEOS-3 (Marks & Sailor, 1986). Because of the spectral overlap of noise and geoid signature for typical seamounts (of diameter 20-60 km), little can be done in the way of further processing to substantially improve the signal/noise ratio.

The maximum geoid anomaly to be expected from a seamount depends on its shape, height and density, as well as the mechanism of isostatic compensation of the volcanic load. This mechanism governs the reconfiguration of the underlying crustal density structure. In modelling the lithosphere as a thin elastic plate overlying a weak fluid substratum, Watts & Ribe (1984) have shown that the age of the lithosphere at the time of loading is an important factor in determining geoid anomaly. As the lithosphere ages, it cools and becomes more rigid (effective elastic thickness increases) in response to surface loading. Thus seamounts emplaced close to ridge crests show significantly lower geoid anomalies than those emplaced off-ridge on the flanks. The difference is greatest for very large seamounts (factor of about 300%) but diminishes with seamount size to a factor of about 40% for more typical seamounts (30 km diameter).

Lambeck & Coleman (1982) have examined detectability of seamounts from altimeter measurements by modelling geoid anomaly over conical seamounts resting on lithosphere of moderate effective rigidity. The sides of their nominal seamounts slope at 8.5° and rise from a seafloor at 4.5 km depth. These parameters are typical for the Cook-Austral region and probably apply for much of the Pacific. The results indicate that for a seamount approaching the surface (ie height equals 4.5 km) the maximum geoid anomaly is 2.7 m
directly over the centre. For a measurement noise level of 0.5 m, a
perceptible geoid signal would still be obtained from sub-satellite
tracks passing within 30 km of the load centre. Taking the case of
smaller seamounts - a 2.5 km high seamount would be detectable
provided that the track passes within 10 km of its centre.

The quality of altimeter data varied with sea-state and other
factors so the spatial resolution of seamounts from altimetry is
difficult to assess. Various studies indicate an along-track
uncertainty in seamount location from 1.1 to 15 km, while cross-track
the uncertainty is 25 to more than 40 km. The accuracy of cross-track
location is significantly improved for seamounts with multi-track
coverage, particularly if this coverage includes non-parallel tracks
(seamount MW3, discussed later, is a good example).

We conducted a simple test to evaluate altimeter resolution of
seamounts in our study region. Well-mapped isolated seamounts, atolls
and islands with good altimeter coverage were selected, and the
distance between the centre of the bathymetric feature and the
visually estimated centre of the associated geoid anomaly high was
measured. Thirty-two such measurements yielded a value of 4.0 +/- 3.3
(SD) km. It seems, therefore, that a value of 15 km for the
along-track uncertainty is conservative.

Seamount size cannot be determined with any precision directly
from geoid or altimeter-derived gravity images (Sandwell, 1984). By
modelling seamount response on two or more adjacent tracks, Baudry &
others (1987) were able to (i) improve the accuracy of the cross-track
location of seamounts to within 15 km, and (ii) obtain estimates of
seamount size and shape. Their method was applied to the
Southern-Cook and Austral area, and successfully tested during two recent oceanographic cruises (Baudry and Diament, in press). Previous field investigations aimed at locating uncharted seamounts from altimeter data in the Atlantic and other parts of the Pacific had only limited success. Uncharted seamounts were detected in only about 50% of the cases tested (Keating & others, 1984).

Selection of Sites for Investigation

Altimeter data used during the cruises comprised, (i) black & white and colour images of SEASAT-derived gravity (provided by W. Haxby of Lamont-Doherty Geological Observatory) - both cruises, and (ii) colour images of along-track filtered residual sea surface heights (Baudry, 1986) derived from the combined GEOS-3/SEASAT data sets (Liang, 1983) - RV 'Moana Wave' cruise. Possible uncharted seamounts were located by comparing the altimeter images with available bathymetric charts including Mammerickx & others (1973), Kroenke & others (1983), NZ Navy Hydrographic Office 1:1M plotting sheets, plots of digital data from the US National Geophysical Data Center, and British Admiralty navigation charts.

Navigation

Positioning aboard RV 'Moana Wave' was by satellite navigation using the U.S. Navy operated Transit and Global Positioning Systems. Though HMNZS 'Tui' was equipped with a Transit satellite receiver, a failure of the system meant that the initial crossing of seamount T1 was by conventional navigation. The positioning is considered reliable, however, because of a fix on Nassau Island not long before the crossing.
Investigation Results

The three areas investigated are shown in detail in Figures 2-5. The four seamounts (MW1, MW2, MW3 & T1) are mapped with a 500 m contour interval. GEOS-3/SEASAT sub-satellite tracks and profiles of sea surface height derived from the Liang (1983) data set are illustrated for comparison with the bathymetry. The sea surface heights are referred to the GRS80 ellipsoid (a=6378137 m, f = 1/298.257222). Profiles of bathymetry and free-air gravity along RV 'Moana Wave' tracks across the seamounts are presented in Figure 5.

Table 1 provides a summary of information on the seamounts, including exact position, size, morphology and geological/geophysical characteristics.
<table>
<thead>
<tr>
<th>Seamount</th>
<th>MW1</th>
<th>MW2</th>
<th>MW3</th>
<th>T1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Area</td>
<td>Southern Line Islands</td>
<td>Southern Line Islands</td>
<td>Southern Line Islands</td>
<td>Samoan Basin</td>
</tr>
<tr>
<td>Summit position</td>
<td>1°45.4'S 156°10.0'W (approx.)</td>
<td>1°55.5'S 156°17.3'W</td>
<td>3°22.3'S 155°41.0'W</td>
<td>12°25.2'S 166°27.9'W</td>
</tr>
<tr>
<td>Depth (m)</td>
<td>2200</td>
<td>&lt;2700</td>
<td>1170</td>
<td>2775</td>
</tr>
<tr>
<td>Depth of surrounding ocean (m)</td>
<td>5000</td>
<td>5000</td>
<td>5200</td>
<td>5600</td>
</tr>
<tr>
<td>Seamount height (m)</td>
<td>2800</td>
<td>&gt;2300</td>
<td>4030</td>
<td>2825</td>
</tr>
<tr>
<td>Max. recorded FA gravity anomaly (microm.s⁻²)</td>
<td>710</td>
<td>720</td>
<td>1470</td>
<td>940</td>
</tr>
<tr>
<td>Max. recorded magnetic anomaly (nT)</td>
<td>220</td>
<td>400</td>
<td>700</td>
<td>540</td>
</tr>
<tr>
<td>Morphology</td>
<td>Cone-shaped with steep sides</td>
<td>Steep, gullied flanks</td>
<td>Gaussian shape with relatively flat summit area</td>
<td>Steep-sided with jagged peaks near the top - possible subsidiary cone on eastern flank</td>
</tr>
<tr>
<td>Approximate diameter near base (km)</td>
<td>25</td>
<td>30</td>
<td>40</td>
<td>35</td>
</tr>
<tr>
<td>Mean slope of mid-flanks (deg.)</td>
<td>15</td>
<td>13</td>
<td>14</td>
<td>14</td>
</tr>
<tr>
<td>Seamount</td>
<td>MW1</td>
<td>MW2</td>
<td>MW3</td>
<td>T1</td>
</tr>
<tr>
<td>----------</td>
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</tr>
<tr>
<td>Area</td>
<td>Southern Line Islands</td>
<td>Southern Line Islands</td>
<td>Southern Line Islands</td>
<td>Samoan Basin</td>
</tr>
<tr>
<td>Sediment cover</td>
<td>300 m thick horizontally-layered sediment ponds in basement depressions adjacent to seamount.</td>
<td>As for MW1</td>
<td>Capped by veneer of pelagic sediment (from-fossil ooze/chalk) 0-30 m thick.</td>
<td>No obvious sediment cover on the seamount, (red-clay) 20-30 m of pelagic sediment covers the basal slope and extends out onto the abyssal plain.</td>
</tr>
<tr>
<td>Dredged Samples</td>
<td>-</td>
<td>-</td>
<td>FeMn-oxide encrusted hyaloclastite cobbles. Crusts up to 5 cm thick.</td>
<td>-</td>
</tr>
</tbody>
</table>

The Seamount Area is characterized by thick sediment layers, particularly in the Southern Line Islands. MW1 shows a 300 m thick horizontally-layered sediment cover with ponds in basement depressions adjacent to the seamount. MW2 has a similar sediment pattern as MW1. MW3 is capped by a veneer of pelagic sediment (from-fossil ooze/chalk) 0-30 m thick. The Samoan Basin has no obvious sediment cover on the seamount, with a (red-clay) 20-30 m of pelagic sediment covering the basal slope and extending out onto the abyssal plain. Dredged samples from MW3 show FeMn-oxide encrusted hyaloclastite cobbles with crusts up to 5 cm thick.
Seamounts MW1 & MW2

The area shown in Figure 2 is well covered by a number of altimeter sub-satellite tracks. The sea surface heights indicate a somewhat confused pattern though there is a general geoid high of about 1.0 m over the centre of the area. Baudry (1986) predicted a seamount at \(1^\circ 54^\prime S \ 156^\circ 12^\prime W\).

RV 'Moana Wave' crossed the area from south to north with the seafloor swath-mapping system SeaMARC II (Blackinton & others, 1983) deployed. The ship crossed the upper flank of a large seamount MW2 and then almost directly over the top of a slightly smaller, cone-shaped seamount MW1 to the NNE. The geoid high thus represents a complex of at least two topographic highs. Mammerickx & others (1973) show a relatively small high of 600 m (above surrounding seafloor) in the position of MW2 indicating that a survey may have crossed over the lower slopes of MW2 previously.

Seamount MW3

A number of criss-crossing altimeter sub-satellite tracks provide good coverage of the area in Figure 3. A well-defined geoid high of 1.3 m occurs at the centre of the area clearly indicating the presence of a large seamount. A location of \(3^\circ 24^\prime S \ 155^\circ 42^\prime W\) was predicted by Baudry (1986).

The RV 'Moana Wave' was set on a course to intersect the centre of the geoid high, with the result that the ship passed directly over the summit of the seamount on its first pass. MW3 was mapped by seismic and also sampled (Table 1), with thick ferromanganese oxide
crust being recovered.

Seamount Tl

As can be seen from Figure 4, seamount Tl is actually located in a 60 km gap between altimeter sub-satellite tracks. Though data along the nearest tracks to the SW indicate a geoid anomaly of only about 0.7 m, they show a trend of increasing anomaly height towards the seamount. The possibility of a seamount was first recognised from W. Haxby's SEASAT-derived gravity images in which data gaps are filled by interpolation from adjacent tracks (Haxby & others, 1983).

Tl was initially mapped by HMNZS 'Tui' with a single crossing of the summit. A subsequent crossing was made in 1987 by RV 'Moana Wave' in order to improve the bathymetric and geophysical control on the feature.

Conclusions

We have demonstrated the effectiveness of using satellite altimeter images and displays to detect and locate uncharted seafloor topographic features in the SW Pacific. Our predictions have been confirmed by follow-up shipboard verification surveys. Where medium-large uncharted seamounts are indicated on two or more adjacent sub-satellite tracks, the location accuracy appears to be better than 15 km.

Time constraints allowed the testing of only a few major geoid anomalies during the research cruises. Examination of altimeter data covering the SW Pacific indicates a considerable number of other geoid
anomalies which could be interpreted as major uncharted seamounts. These await confirmation by other ship surveys. Each new find has important seafloor resource implications for the South Pacific island nations, particularly in relation to deposits of metalliferous crusts.

Acknowledgements

P.J. Hill thanks fellow members of the 1986 HMNZS 'Tui' and 1987 RV 'Moana Wave' expeditions for their support and dedicated data acquisition efforts during the seamount searches; he is particularly grateful for the support of co-chief scientists Geoff Glasby, Maury Meylan and Bill Coulbourn. Brian Pashley drafted the figures and Joan Brushett typed the manuscript. William Haxby of Lamont-Doherty Geological Observatory provided the spectacular SEASAT-derived gravity images which first revealed the significant potential of the satellite altimeter data and stimulated this study. Thanks also go to Richard Rapp of the Ohio State University for generously supplying the processed combined GEOS-3/SEASAT digital data sets.

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Figure Captions

Figure 1. Location map and regional bathymetry (after 'Geographic map of the circum-Pacific region', Circum-Pacific Council for Energy and Mineral Resources, 1978). Stippled areas indicate water depths shallower than 4000 m. The lower map depicts the same area at reduced scale and illustrates SEASAT/GEOS-3 data coverage.

Figure 2. Area of seamounts MW1 and MW2, showing (a) research ship's track, (b) observed bathymetry (500 m contour interval), (c) GEOS-3/SEASAT sub-satellite tracks, data points and along-track profiles of sea-surface heights. Sea-surface height increases to the left of the baselines (tracks) and positive values, relative to a base of 15.50 m, are highlighted by shading.

Figure 3. Area of seamount MW3. Data presentation is as in Figure 2, except that (i) some sea-surface height profiles have been omitted for clarity, and (ii) the baselines represent 14.00 m.
Figure 4. Area of seamount T1. Data presentation is as in Figure 2, except that the sea-surface baselines represent 17.00 m.

Figure 5. Ship bathymetry and free-air gravity profiles over seamounts MW1 & MW2, MW3 and T1. Track locations A-B, C-D, and E-F are shown in Figures 2, 3 and 4, respectively.
FIGURE 1 (S1).
MW1 & MW2

Relative sea-surface height (m)

-3000 — Bathymetry (m)

FIGURE 2 (S1).
**MW3**

**Relative sea-surface height (m)**

-3000 - Bathymetry (m)

**FIGURE 3 (S1).**
FIGURE 4 (S1).
Bathymetry & Gravity Profiles

FIGURE 5 (S1).
SECTION 2

A COMPLEX OF MUD VOLCANOES ON AN OCEANIC PLATEAU

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Our expedition to the Cook Islands was funded through the Committee for Co-ordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas (CCOP/SOPAC). We thank Haru Matsumoto and Karen Mansfield of the HIG SeaMARC II team for their advice and assistance during the course of our cruise. We also thank Hseuh-Wen Yeh for the isotopic analyses and our colleagues at HIG for their interest in and discussion of this find. Hawaii Institute of Geophysics contribution No. 0000.
ABSTRACT

A complex of mud volcanoes was discovered on the northeastern edge of Manihiki Plateau (10° 18.5'S, 161° 27.5'W) during a February 1987 cruise of the RV MOANA WAVE in the territorial waters of the Cook Islands. The edifice is 1900 m high and is the central feature in a field of about 100 mud volcanoes. Individual cones appear as acoustically transparent piercements on single-channel seismic reflection profiles collected during this and earlier expeditions to the area. SeaMARC II side-scan images of these piercement structures typically reveal nearly conical, steep-sided peaks of about 1 km diameter rising several hundred metres above the smooth and featureless plateau surface. Many of the cones are formed by multiple flow units. A few have summit depressions, and moats partially surround some of the cones.

Distributed like holes on a dart board, individual cones become more densely spaced about a bulls-eye located near the northeastern edge of the Manihiki Plateau, 50 km SW of Rakahanga Island. At that location about 40 vents coalesce to form an edifice about 25 km in diameter, rising from a plateau depth of 3200 m to a summit at 1310 m depth.

Patterns in the side-scan images suggest fluid sediment flow radiating from the center of this feature. Incised channels trend away from the edifice toward the northeast. This irregular sea floor truncates tilted, NE-dipping subbottom reflectors, but to the south, the sea floor and subbottom reflectors are smooth and dip gently to the SW.
Recent foraminiferal ooze was cored from a satellite cone. Middle Eocene planktonic foraminifera are embedded in burrowed limestone dredged from the summit of the composite edifice. Some of the limestone cobbles are manganese encrusted. Coralline fragments were also recovered in the same dredge.

INTRODUCTION

Shale Diapirs and Mud Volcanoes

For most geologists, sedimentary volcanism and mud volcanoes are geological curiosities or items of interest, but probably not viewed as of great importance to the geological record, particularly not to the marine sedimentological record of oceanic plateaus. The existence of a field of these features on an ocean plateau is surprising, indicates a dynamism not previously expected, and raises the prospect of at least a limited resource potential.

Subaerially, some of the most spectacular active examples of mud volcanoes are along the exposures of trench fore-arcs, such as the arid Makran of Pakistan (Harms et al., 1984), on Timor (Barber et al. 1986) and on the Cedros Peninsula, Trinidad (Higgins and Saunders, 1967). O'Brien (1968), Freeman (1968) and Hedberg (1974) provide a more complete list of locations, examples, and source literature. At those locations mud, dry methane, and methane-charged fluids are episodically spewed and discharged from central vents and vent complexes. Gases mix wall rock and sediment from various subbottom depths and buoy the viscous slurry to the surface in violent explosions or in quiet emplacement in cones. The cones are usually steep-sided and symmetrical, some have relatively flat summits and
most are strewn with heterogeneous boulders of disparate ages and are rapidly eroded once extrusion ceases. Denudation of the mud tends to leave lithified blocks behind as a more resistant lag deposit.

A loss of reflections creates a vertical swath or wipeout on seismic profiles and is an indication of the presence of submarine sedimentary piercement structures. The loss of seismic energy is associated with gas-charged pore waters and sediment (Behrens, 1988). The surface expressions of piercements is critical to resolving the cause of diapirism and the amount of acoustic structure attributable to seismic artifact (Collette and Rutten, 1970). Lancelot and Embley (1977) compiled a global distribution of deep sea piercements and discussed probable modes of origin, relying on a combination of seismic and 3.5 kHz echo-sounder profiles in their interpretation. Wipeouts are commonly associated with salt diapirism (Walker and Ensminger, 1970; Stanley et al., 1974; Behrens, 1988).

Features of limited lateral extent, such as cones, are difficult to image with seismic profiling systems unless the ship track passes directly across the summit. With the advent of such side-scan imaging systems as GLORIA and SeaMARC II, however, steep protrusions resembling mud volcanoes have been imaged on the submerged portions of the Barbados and Mariana fore-arcs (Biju-Duval et al., 1982; Stride et al., 1982; Westbrook and Smith, 1983; Hussong and Fryer, 1985). These findings depend on the side-scan system capabilities, primarily the ability to map swaths of sea floor at high levels of resolution. This report extends earlier inventories of diapiric structures and sites through our discovery, with side-scan imaging of a field of mud volcanoes on a midocean plateau.
Marine Expeditions to the Manihiki Plateau

In late 1986 and early 1987, the Committee for Coordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas (CCOP/SOPAC) sponsored three cruises of the RV MOANA WAVE to assess the mineral resource potential of several Pacific islands nations. During the last of these three voyages, SeaMARC II was used to map a portion of the Manihiki Plateau margin in the vicinity of Rakahanga Island, Cook Islands (Figs 1 & la). The area was selected for detailed survey because seismic reflectors correlative with Lower Cretaceous, volcaniclastic, copper-bearing beds drilled at DSDP 317 (Schlanger, Jackson et al., 1976) crop out along an escarpment to the west of Rakahanga. The native copper occurs as flecks disseminated in the 150 m of volcaniclastic sediment immediately overlying basalt basement at Site 317 (Jenkyns, 1976). Correlative volcaniclastic sediments were dredged during a 1984 expedition of the RV SONNE to a segment of the escarpment NW of this study area (Beiersdorf and Erzinger, 1986), but the sought-after native copper was not recovered in their dredge hauls.

Additionally, the area was selected for study because earlier seismic reflection profiles across the region revealed pronounced wipeouts. For example, one of these features is shown in Figure 4a of the DSDP Site 317 chapter, but no mention is made of it in the accompanying text (Schlanger, Jackson et al., 1976). Wipeouts on records from the Manihiki plateau are displayed as steep-limbed parabolic diffractions perched above a section devoid of reflectors. Strong, subhorizontal reflectors are abruptly truncated against the sides of these acoustic wipeouts. Viewed as curiosities prior to our survey, with the first side-scan track across the region these seismic
features became the principal focus of our investigation in the Cook Islands.

The Manihiki Plateau is a broad, midocean plateau generally at depths between 2000 and 4000 m (Fig. 1). It rises gradually from the Samoan Basin to the west and drops abruptly to the Penrhyn Basin to the east along a steep escarpment and a deepening series of ridges and troughs. The northeastern border with the central basin of the Pacific is complex, formed by two major faults and a 300 km wide band of rough sea floor (Winterer et al., 1974). Danger Island and Suwarrow Troughs slice through the central portion of the plateau and separate the western plateau from the high plateau to the east. Rakahanga and Manihiki Islands are located on the NE extremity of the high plateau.

The Manihiki Plateau was the subject of several geophysical expeditions and considerable speculation regarding its origin (Hussong, et al., 1976 and references therein). Plate-tectonic reconstructions suggest that its anomalously shallow depth results from voluminous outpourings of lava at a former triple junction (Heezen et al., 1966; Winterer et al., 1974). The sedimentary section at the centre of the high plateau is one of the thickest of any in the South Pacific and is well known from Leg 33 of the Deep Sea Drilling Project. The three holes drilled at DSDP Site 317 penetrated 944 m, cored 766.5 m and recovered 490.4 m of sample (Schlanger, Jackson et al., 1976). At Site 317 the sediments are 910 m thick, predominantly carbonates in the uppermost 600 m, becoming rich in volcanic sandstones and siltstones in the lowermost 250 m (Fig. 2). Micro- and macrofossils indicate a progressive deepening of the site since the Cretaceous. Reconstruction of the paleoenvironment of the
region suggests widespread volcanism (Rea and Vallier, 1983) and a very shallow water environment in the Cretaceous, perhaps with islands, and certainly with carbonate banks in the vicinity. Sapropelic Early Cretaceous sandstones indicate stagnation of bottom waters, probably in an environment of closed and semi-closed basins (Jenkyns, 1976).

METHODS

In addition to SeaMARC II capabilities, the MOANA Wave is equipped with GPS and TRANSIT satellite navigation and gathers the standard suite of underway geophysical data: gravity, magnetics, 3.5-kHz bathymetry, and seismic reflection profiles.

Technical details of the SeaMARC II imaging system are described in Blackinton et al. (1983) and in Hussong and Fryer (1983). High amplitude returns are printed dark in our data display so that rough bottom, as well as hills and ridges facing the ship tracks appear as dark areas; but smooth sea floor as well as depressions, flow fronts, and escarpments dropping away from the direction of view are imaged as white areas.

The side-scan images and bathymetric swaths acquired during our 24 hour survey of Manihiki Plateau were cut from the recorder and mounted on a plot of the ship tracks to monitor survey progress and to serve as a guide for later sea floor sampling. The mosaic was hand contoured to a 100-m interval. The side-scan mosaic was interpreted by enhancing gray-tone contrasts with ink lines while at the same time referring to the contoured bathymetry and 3.5 kHz records.
Seven parallel ship tracks, oriented SW-NE, are the basis of our survey. Survey time in this area was very limited, because of prior commitments of this cruise to additional objectives in Kiribati territorial waters.

The smooth surface of the Manihiki Plateau lies at about 3000 m depth in the SW corner of the study area (Fig. 3). In contrast to the eastern edge of the plateau, which falls away to the deep Penrhyn Basin in a series of ridges and troughs, contours along the NE edge of the plateau are complicated by an irregular distribution of small hills and cones about a hummocky seamount rising to 1310 m depth. Just to the north of this seamount and along a SW-NE trend, depths decrease to more than 3800 m. Channels originate at the hills and cones and trend away toward the northeast. This pattern and the slope of the sea floor is locally reversed near Rakahanga atoll.

The origin of the complicated bathymetry is explained in the bizarre and detailed features revealed in the SeaMARC II side-scan images. The mosaic of these swaths and interpretation of line drawings reveal an array of isolated cones clustering about and coalescing to form a central edifice at 10°18.5'S, 161°27.5'W (Figs. 4 and 5).

Satellite cones are steep-sided (slopes as steep as 30°) and rise as much as 500 to 600 m above the smooth surface of the plateau; some are formed by multiple flows and flank effusions (Figs. 6 and 7). Some have small summit depressions and most meet the flat-lying plateau surface in sharp contacts. Those satellite cones to the north
and northeast of the main edifice are particularly spectacular for their relief and for the presence of moats and channels that lead away toward the deeper waters to the northeast (Fig. 7). The density of satellite cones increases toward the central edifice.

The mud volcano complex, the focus of this survey, is about 25 km across and is the aggregate of about 40 individual cones and complicated flow units. Nondescript, low hills and discontinuous flows form a summit flanked on all sides by pronounced cones. The ensemble produces the blossom-shaped configuration displayed in the bathymetry (Figs. 4, 5 and 6). Contrast in gray tones around these cones produces a three-dimensional image that matches the relief indicated in the SeaMARC II bathymetry. Similar contrasts mark the rough sea floor to the northeast of the seamount where erosional channels trend down-slope across the edge of the plateau.

Seismic reflection profiles display the topography of the area in greatly exaggerated vertical relief that enhances perception of the setting of this seamount on the edge of the plateau (Figs. 8 and 9). Seismic reflector sequences are truncated by acoustic diapirs representing the location of cones (Fig. 8). Where ship tracks pass directly over cones, an umbrella-shaped, parabolic diffraction pattern caps each of these diapirs. In each of our profiles, sub-bottom reflectors dip away from the main edifice and are erosionally truncated as depths increase to the NE. Normal faulting is evident in the NE portions of these profiles. Some profiles display slumps of surficial sediment. Because reflector sequences are only slightly disturbed over a few broad basement ridges on the high plateau, it is relatively easy to trace horizons from DSDP 317 to the study area (Fig. 9). Bottom-simulating reflectors (BSRs) are either absent, or
indistinguishable from the reflections off the strata that make up the sedimentary section.

The free-air gravity-anomaly pattern mainly reflects the sea floor topography with gravity highs centered over Rakahanga and Manihiki atolls and a 61 mgal high centred over the mud volcano (Fig. 10b). Three-dimensional gravity modelling of the feature indicates an apparent average density of 2.4 g/cc. This value is higher than expected for unconsolidated sediment, and suggests that the "mud" volcano may be constructed over a high density body - probably uplifted basement, or an igneous intrusion.

The magnetic field over the area is moderately anomalous (Fig. 10c). Both Manihiki and Rakahanga atolls have strong magnetic signatures associated with their volcanic pedestals. Variations trending northward along the plateau margin are in the order of 300 nT, and are believed to result from structural relief on the basaltic basement and possible infrabasement magnetic heterogeneities. Though the magnetic field is disturbed in the vicinity of the mud volcano, the observed field cannot be directly attributed to the morphology of the edifice. The seismic data indicate that the feature is located over a fault zone, so part or all of the anomalous field may be due to faulted basement at the plateau edge. The possibility of an underlying magnetic intrusion cannot be ruled out, however. The magnetic data along the SW-NE line crossing the mud volcano just to the north of its summit were processed by Werner deconvolution (Jain, 1976; Hsu and Tilbury, 1977). The clustering of magnetic source estimates at depths of 2.5 to 3.5 km below sea level, directly beneath the feature is indicative of an underlying magnetic body or bodies, for example, uplifted basement or an igneous intrusion. Although the
gravity and magnetic signatures of this feature are suggestive of igneous volcanism, its seismic signature and surficial rocks indicate a different origin.

Sampling efforts were severely limited because of the large number of preassigned objectives for this cruise. Attempts were made to drop four free-fall corers and two free-fall grab samplers on two isolated satellite cones to the northwest of the composite seamount (Figs. 4 and 11). No samples were recovered from drops on the large cone shown in Figure 7 (one corer was lost, a second corer and a grab sampler returned empty). Two floats returned with cores. Shipboard inspection of these core tops and bottoms revealed that both contained assemblages of Recent planktonic foraminifera, but these corers may have fallen to the side of this small cone. Disappointed at the lack of information and lack of more exotic sample in the cores, an attempt was made to sample the composite seamount. Dragging a pipe dredge backed with burlap across the summit of this feature retrieved a haul of manganese-encrusted, stained limestones mixed with manganese-encrusted, shallow-water macrofossils and Recent foraminiferal sands.

The foraminiferal assemblage in the limestone is of middle Eocene age and is typical of warm-water, open-marine conditions (Fig. 12). The middle and upper Eocene are present at about 350 m subbottom at DSDP Site 317 (Fig. 2), and we assume similar source depths for this sample. Stable isotopic analysis on two samples from this limestone cobble produced carbon isotopic values that match those expected for middle Eocene planktonic foraminifera ($^{\delta^{13}}C = 1.93$ and $2.00$) and oxygen values ($^{\delta^{18}}O = 0.73$ and $0.73$) that are only slightly higher than expected (after Figures 11 and 7 of Savin and Yeh, 1981). Secondary carbonates precipitated or altered in the
presence of microbially generated CO\textsuperscript{2} have strongly negative del 13\textsuperscript{C} values (Behrens, 1988). The positive values obtained from our sample indicates a lack of diagenetic alteration.

**INTERPRETATION**

**Comparison with Other Deep-Sea and Subaerial Cones**

The cones dotting and clustering along the northeast edge of the Manihiki Plateau are yet another entry in the broad spectrum of diapiric features described two decades ago (O'Brien, 1968). They are remarkable for their conical shapes, steep, smooth, sediment-covered slopes, abrupt contacts with flat-lying sediments of the plateau, curious seismic signature, and location on an ocean plateau. The composite edifice that they form is spectacular in terms of its 25 km diameter, 1,900 m of vertical relief, and intricacy of flows and bedforms between its multiple cones and mounds (Fig. 13).

Lancelot and Embley's (1977) catalogue of deep-sea piercement structures contains no examples closely resembling the diapirs displayed in our seismic profiles. The Manihiki diapirs produce truncation and bending of adjacent reflectors, apparent transparency of the piercements, and high surface relief (e.g. Fig. 8), a form similar to the diapirs of the Magdelena delta, Columbia (Figure 3 of Shepard, 1973).

The presence of Eocene planktonic foraminifera in limestone dredged from the summit of the composite edifice and the sediment cores retrieved from the satellite cones are evidence of a
non-volcanic origin for these cones, and a comparison of satellite cones with small submarine volcanic cones and domes reveals nothing comparable in the published record of side-scan and swath-mapped sea floor. Domes were swath-mapped with SEABEAM on the the flank of the East Pacific Rise (Lonsdale, 1983), but those are about 2 to 5 km in diameter and only a couple of hundred metres high, broader than most of the Manihiki satellite cones. Those features show none of the complications and only a fraction of the 1,900 m of vertical relief of the composite seamount of this study area. Cones of the Manihiki field are smaller than the back-arc igneous volcanoes surveyed in the Mariana island arc; those seamounts are 15 to 20 km in diameter and 2,000 m high (Hussong and Fryer, 1983). As important, the surfaces of the Manihiki cones exhibit little of the sharply contrasting gray tones displayed in the SeaMARC II image of the back-arc volcanoes. During one of our surveys in the Line Islands, Kiribati, we crossed a steep-sided (volcanic ?) cone of several hundred metres relief and about 1 km diameter on the summit of Chapman guyot. Although steep-sided, that feature displays none of the seismic transparancy so characteristic of the Manihiki cones. The diapiric signature seen on seismic reflection profiles (Fig. 8) is, therefore, only partly a result of scattering of the acoustic energy from the steep sides of these cones.

The diapirs, mud volcanoes and mud ridges seen in GLORIA images of the Barbados fore-arc, and in SeaMARC II images of the westernmost portion of the Timor Trough, north Panama thrust belt, and the Mariana fore-arc differ in morphology from the Manihiki Plateau examples. Although about the same diameter, the Barbados examples are of much lower relief (about 60 m) with planar tops (Langseth et al., 1988), and the Timor Trough mud ridges are elongated - probably in response
to fault control (Breen et al., 1986). Mud volcanoes topping the crests of anticlines along the north Panama thrust belt (Breen et al., 1988) are the size of the smallest cones depicted in the side-scan records from the Manihiki plateau. The Mariana seamounts are about 20 to 30 km in basal diameter with 1500 m relief and one is extruding serpentine (Hussong and Fryer, 1985). They are larger than individual Manihiki cones but are about the same size as the composite seamount. Much of the irregular surface in the Manihiki images resembles the "mud lumps" of the Mariana fore-arc.

The form of the satellite cones closely resembles subaerial mud volcanoes of the Makran (Harms et al., 1984) and Caspian Sea area (Zakubov et al., 1971), and the mosaic of the composite feature displays the kind of complicated outlines and array of cones mapped for Palau Semau, Indonesia (Barber et al., 1986). In Texas, clastic dikes and dike swarms occur along faults - cones are positioned at the intersection of fault traces (Freeman, 1968). A similar structural control may be responsible for the location of the Manihiki cones.

Mode of Origin of Manihiki Plateau Cones

The mechanism producing and maintaining this field of mud volcanoes remains unknown because of a lack of direct measurements. Some clues are available in the moating and erosive channels developed around satellite cones and along the northern border of the field. Peripheral moats could form as a result of local subsidence due to the subsurface withdrawal (Shepard, 1973), lateral migration, and ultimate expulsion of the sediment that forms the cones (for example, off
Barbados, Langseth et al., 1988). Alternatively, and in view of the extensive network of channels leading NE from the field (Fig. 3), the slurry emitted from the cones may form erosive density currents. A third possibility is that pore-water sapping is creating the submarine analog of the morphology seen in coastal New Zealand (Schumm and Phillips, 1986). These processes are not mutually exclusive and each may be important to the forms displayed in the mosaic.

In a review of the role of methane in shale diapirism, Hedberg (1974) established just how commonplace methane overpressuring is in marine sediments and how ubiquitous methane discharges are in subaerial shale diapirs and mud volcanoes. Methane is, therefore, a likely agent for the formation of the Manihiki Plateau cones, especially because a small portion of the Cretaceous section drilled at DSDP 317 is sapropelic (Jenkyns, 1976). Organic carbon was measured at 28.7% in a thin interval in Core 317A-16 (Fig. 2 and Cameron, 1976), and that interval lies within volcaniclastic sandstones of relatively high porosity, overlain by cherty chalks of lesser porosity (Fig. 2) and permeability (Gieskes, 1976), a barrier to diffusion and an ideal setting for overpressuring. Other attributes of the pore fluids display strong gradients between the ooze, chalk and limestone sequence of Units 1 and 2 and the volcaniclastic sediments of Unit 3 (Fig. 2).

The composite seamount is a large-scale structure that could conceivably only be created by a vigorous, localized geodynamic process. A magmatic heat source at depth may be a driving force. Such a heat source could accelerate hydrocarbon generation in the sapropelic Cretaceous beds and could mobilize pore fluids by convection. Magma emplacement may have produced local uplift and
fracturing of overlying formations, facilitating upward movement of fluids and fluidized sediment. Although explosive emission of magmatic water vapor and carbon dioxide may contribute to mud volcano formation (Chivas et al., 1987), the lack of diagenetic alteration indicated in the carbon and oxygen isotopic values of foraminiferal limestone dredged from the summit argues against the involvement of high temperature fluids in the emplacement of this rock. Lithospheric flexure due to loading by nearby Rakahanga and Manihiki atolls may have localized faults to account for the presence and size of this composite seamount.

Although the importance of this type of feature to the total marine geological record is difficult to assess, this newly-discovered field is not unique, though it may be one of the largest. Wipeouts are seen in seismic reflection profiles at several other crossings of the Manihiki Plateau and also on the Ontong-Java Plateau (Furomoto et al, 1976; Colwell and Tiffin, 1986). The economic importance of this field to the Cook Islands is also difficult to assess at this time, but the find itself provides a basis for cautious optimism regarding the resource potential of the area.
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LIST OF FIGURES

Figure 1. Location map for Manihiki Plateau mud volcano field, Cook Islands. Bathymetry after Kroenke et al. (1983). Shaded area locates the high Manihiki Plateau.

Figure 1a. R/V Moana Wave survey lines over the area of mud volcanoes. Lines 1-5 are the same as shown on Figure 4. Half-hourly positions on the tracks are shown as small squares.

Figure 2. Lithologic section, porosity and selected attributes of the interstitial waters at DSDP Site 317 (after Schlanger, Jackson, et al., 1976; Gieskes, 1976).

(a) Lithologic summary (depth scale in hundreds of meters):
Unit 1 -- nanno-foram and foram-nanno oozes and chucks
Unit 2 -- foram-nanno ooze, firm ooze, and chalk with black chert (triangles)
Unit 3 -- limestone and volcanic siltstones and sandstones with scattered specks of native copper and abundant zeolites. Black line indicates approximate subbottom depth of layer rich in organic carbon.
Unit 4 -- basalt.

(b) Porosity determined by GRAPE as a part of the routine physical properties analyses aboard Glomar Challenger.
(c) Ammonia and sulphate concentrations at Site 317.

(d) Alkalinity and salinity at Site 317.

(e) Alkaline earth and dissolved silica concentrations at Site 317.

Figure 3. Bathymetry based on SeaMARC II computer-generated mosaic. Contour interval = 100 m. Depths are calculated assuming a sound speed of 1500 m/s.

Figure 4. Portion of a mosaic of seven SeaMARC II side-scan sonar tracks across the NE edge of the Manihiki Plateau. Each track line appears as a stripe centred in a 10 km-wide swath of coverage. High amplitude echos are printed dark in this display. Numbers locate NE-SW trending ship tracks and locate seismic reflection profiles. Boxes outline areas shown in detail in Figures 6 and 7. Broken line locates profile displayed in Figure 11 with cross-ticks marking free fall corer (FFC) and grab (FFG) sampling attempts. Circle to the SE of track 3 locates Pipe Dredge 1 (DP1).

Figure 5. Line-drawing interpretation of SeaMARC II mosaic shown in Figure 4. Rough sea floor and SE-facing slopes are shown as dark areas.

Figure 6. Half of a SeaMARC II side-scan swath showing a three-lobed satellite cone with central spire. As for the mosaic shown in Figure 4, high amplitude returns
are printed dark, so that smooth sea floor and slopes falling away from the ship track appear as white. Image is located in Figure 4.

Figure 7. Detailed SeaMARC II images of selected satellite cones displayed with 3.5-kHz bathymetric profiles. Profiles are referred to the track of the ship, which is shown as the center line in the set of three parallel lines. Images are located in Figure 4. Each scan images three of the same cones, but from opposite look directions. Vertical relief of these three cones is minimized in the profiles because in each instance they are located several kilometres to the side of the ship track. Erosional channel and moat surrounding cone are well displayed in 7b.

Figure 8. 40-200 Hz seismic reflection profile showing the disruption of reflector sequences by acoustic diapirs. Equidistant vertical lines are half-hourly time marks; each interval representing about 6 km of ships track. Number locates profile in Figures 4 and 9.

Figure 9. Line drawings of selected, NE-SW oriented, 40-200 Hz seismic reflection profiles within the Rakahanga survey area. Vertical scale is two-way travel time in seconds. Horizontal divisions represent half-hourly intervals (approximately 6 km). Letters identify reflectors correlated to the lithologic section drilled at DSDP 317. Numbers locate profiles in Figure 4.
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<thead>
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<th>LITHOLOGY</th>
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<td>58</td>
<td>Quaternary to Pliocene calcareous ooze.</td>
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<tr>
<td>C</td>
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<td>Early Eocene cherty chalk.</td>
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<td>B</td>
<td>576</td>
<td>Paleocene (?) - section not recovered at DSDP 317.</td>
</tr>
<tr>
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</tr>
<tr>
<td></td>
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<td>Basalt.</td>
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**Figure 10.** Regional geophysical setting of the Manihiki Plateau mud volcano field.

(a) Regional bathymetry showing the relationship of the mud volcano fields to Rakahanga and Manihiki atolls. Contour interval is 200 m.
(b) Free-air gravity anomaly pattern over the mud volcano field and adjacent Rakahanga and Manihiki atolls. Contour interval is 10 mgal. Dot pattern locates high and dash pattern locates low values.

(c) Magnetic anomaly pattern over the mud volcano field and adjacent Rakahanga and Manihiki atolls. Contour interval is 50 nT. Dot pattern locates high, positive values and dash pattern locates values less than -100 nT.

Figure 11. Seismic profiles showing location of sampling attempts on satellite cones. FFC denotes free-fall corer drop and FFG free-fall grab sampler drop. Profile and drop-sites are located on Figure 4. Equidistant vertical lines are half-hourly marks (representing approximately 6 km distance.

(a) 3.5 kHz profile displaying irregular terrain as acoustically transparent mounds and complete absence of reflectors over the summit of a satellite cone.

(b) 40 - 200 Hz seismic reflection profile showing wipeouts that mark the location of satellite cones. C/C denotes course change.

Figure 12. Planktonic foraminifera dredged from the summit of the composite mud volcano. Specimens drawn protrude from the surface of a limestone cobble and are illustrated
according to their orientation in the rock. Even the best preserved specimens are friable and are stained red-brown. They protrude from bore holes near manganese encrustations, which are about 2 cm thick. Some bore holes are partially filled with Recent planktonic foraminifera. The specimens of Hantkenina, Morozovella, and Acaranina drawn here indicate a middle Eocene age for this limestone. Scale bar equals 1 mm.

Figure 13. Three-dimensional conceptual diagram displaying the general configuration of cones on the northeast edge of the Manihiki Plateau, Cook Islands.
MOANA WAVE TRACK MAP
MANIHIKI PLATEAU

Summit of composite mud volcano

Rakahanga Atoll

Manihiki Atoll

FIGURE 1a (S2).
FIGURE 2 (S2).

DSDP Site 317

POROSITY

NH₄⁺  SO₄²⁻  ALK  Ca⁺⁺  Mg⁺⁺  SiO₂

mmoles/1  100  200  300  500  35  40  400  800

mmoles/1  0  20  3  3  3  3  3
FIGURE 3 (S2).

Bathymetry
FIGURE 4 (S2).

See also next page
FIGURE 4 (S2).
(cont.)
Side-scan Interpretation

FIGURE 5 (S2).
Side-scan of Cone

FIGURE 6 (S2).
Side-scan & 3.5 kHz Bathymetry of Cones
Disruption of Seismic Sequences

FIGURE 8 (S2).
Seismic Profile Interpretation

FIGURE 9 (S2).
FIGURE 10 (S2).
Free-fall Sampling Locations

FIGURE 11 (S2).

SECONDS

TWO-WAY TRAVEL TIME

FFG1
FFC13 & 14
FFG2
FFC15 & 16
Foraminifera from Summit

FIGURE 12 (S2)
MACHIAS SEAMOUNT, WESTERN SAMOA:
TECTONIC DISMEMBERMENT AND SUBDUCTION OF A GUYOT

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ABSTRACT

Machias Seamount is a guyot poised on the seaward flank of the Tonga Trench. A SeaMARC II bathymetric and side-scan sonar survey shows that the trench-facing half of the guyot is dissected by faults aligned parallel to the local strike of the Tonga Trench (WNW-ESE). A lack of faults and the presence of radially distributed sediment flow characterize the NE-facing flank of the guyot. Sediment flow is pervasive on the trench-facing slope, but the pattern is not radial because the neo-tectonic fabric controls patterns of resedimentation. The features displayed in the side-scan image are ample evidence of the remobilization and mixing of sediment and rock prior to subduction of the seamount.
INTRODUCTION

The convergence of lithospheric plates along active margins produces a suite of related morphotectonic features such as volcanic arcs, fore-arc basins, deep-sea trenches and outer swells and associated normal faults. The focus of this report is the effect of plate convergence on seamounts and on sediment accumulation patterns.

Three cruises of the RV MOANA WAVE, sponsored by the Committee for Co-ordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas (CCOP/SOPAC), were devoted to mineral assessment of several Pacific island nations. During the last of these voyages, SeaMARC II was used to produce a bathymetric map and side-scan image of Machias Seamount, located south of Savaii, Western Samoa (Fig. 1). Although not promising in terms of its potential for mineral resources, this guyot is of scientific interest because of its tectonic situation - it is poised on the seaward flank of the Tonga Trench and is destined for subduction within the next 0.5 Ma. Within the study area, the strike of the trench is roughly NW-SE, part of the large arcuate bend of the northern end of this subduction zone. Of particular interest is the tectonic dismemberment of half of Machias guyot by normal faulting.

Escarpsments occur seaward of the Sunda, Japan, Mariana-Bonin, Manila, Solomon, Middle America, Peru-Chile, Puerto Rico and Pliny-Strabo Trenches [1-9] and along the Tonga-Kermadec Trench [10]. Visual observations from submersible dives along the seaward slope of the Japan Trench [11, 12] and dives and drilling off Guatemala reveal escarpments interpreted as normal faults [13, 14]. Where dredged, for example, on the north slope of the Puerto Rico Trench, scarps have
yielded serpentinized peridotite, basalt, and Cretaceous limestone [8, 15]. Most escarpments in this setting are considered a product of normal faults resulting from extension of the upper surface of flexed, elastic lithospheric plates [16-22].

The neo-tectonic fabric produced by plate flexure is oriented parallel to the strike of the adjacent trench and is often not coincident with the strike of the primary tectonic fabric produced at the spreading center along which the lithospheric plate originated. It appears that the primary fabric is reactivated as the secondary fabric is created [7]. This combination of structural orientation occurs along the seaward flank of the Japan Trench [23], of the Middle America Trench [24, 14, 25], and along the Arica Bight segment of the Peru-Chile Trench [26].

The extent to which the structural fabric of a subducting plate is transferred to the landward flank of an active margin is unknown, but the subject of several seamount-related investigations. Bathymetric surveys of the Mariana-Bonin Trench and Michelson Ridge guyots to the east [27, 4], a SeaMARC II survey of the Mariana Trench and Dutton Ridge [28] and surveys and submersible dives on the Daiichi Kashima and Erimo Seamounts in the Japan Trench [29-32] indicate that seamounts both large and small (less than 40 km diameter) are subducted, but that the smaller seamounts are especially prone to break up by fracturing of the subducting plate prior to reaching the trench axis.

The SeaMARC II survey of Machias Seamount reveals the details of a seamount dismemberment by demonstrating that the neo-tectonic fabric in this setting is related to lithospheric flexure, that the
fabric pervades bathymetric irregularities on the subducting plate, and that it has the capacity to thoroughly reorganize structure and sediment accumulation patterns prior to subduction.

METHOD

The MOANA WAVE is equipped with GPS and TRANSIT satellite navigation systems and gathers the standard suite of underway geophysical data: gravity, magnetics, bathymetry (3.5 kHz sonar), and seismic reflection profiles simultaneously with the acquisition of SeaMARC II images. Technical details of the SeaMARC II imaging system are described elsewhere [33]. Structures and fabric oriented parallel to the ship track are enhanced in the side-scan image relative to features striking perpendicular to track. Multiple look directions add to the portrayal of sea-floor features. Rough bottom and hills and ridges facing the ship tracks appear as dark areas; smooth sea-floor and depressions, flow fronts and escarpments dropping away from the look direction are imaged as white areas. Bottom detection errors occur when the instrument temporarily loses track of the sea-floor. The result is a scalloping effect seen along the border of some of the ship tracks.

To retain some of the detail lost in reduction of the mosaic of side-scan images to page size, a line drawing interpretation was created to enhance gray-tone contrasts and to outline key features.
RESULTS

SeaMARC II Bathymetry

The SeaMARC II bathymetry of Machias Seamount reveals a guyot standing about 7000 m in relief to the northeast of the Tonga Trench (Fig. 2a). The summit area is flat, smooth and blanketed with sediment, probably carbonate ooze. It is elongated along a NW-SE axis, a reflection of the structural fabric disrupting this mountain. The NE flank of the guyot descends at a uniform gradient to the abyssal sea-floor at about 4500 m, and the SW flank falls away by steps to the axis of the Tonga Trench at 7800 m depth. The contours along the SW flank are unevenly spaced and depict a series of terraces separated by steeper slopes and scarps. Inspection of the pattern of the contours shows an arcuate WNW to ESE trend for these scarps. The flanks of this guyot are lobed on all sides, the result of debris flows and submarine erosion, but the style of modification varies from radial on the NE slope to complex on the SW flank. Adding to the contrast in morphology across the summit are several NE-SW oriented canyon systems on the trench-facing slope. One large canyon extends from about the 4600-m contour down to the trench axis. Canyons are absent from the NE slope.

The trench axis is poorly defined in the mosaics because at the time of this cruise the SeaMARC II system was not configured for water depths greater than 7000 m. Our interpretation of the noisy signal is that depths reach a maximum of 7800 m in a closed depression just to the west of a small fan associated with the large canyon, centred on the NE slope of the trench (Fig. 2a). The sides of the lower trench slopes are uneven. On the NE side, a large reentrant cuts into the
base of the guyot and on the SW side the boundary with the Tonga
fore-arc is indented where a closed depression borders the trench.
Isolated hills, irregularly distributed ridges and depressions on the
lowermost slope of the SW flank of the trench (fore-arc) are also seen
in side-scan images of other fore-arcs [34-36].

SeaMARC II Side-scan Image

The most noticeable feature on the side-scan image is the
sequence of subparallel, WNW-ESE (about 115 degrees) trending dark
stripes that form an arcuate trace across the SW flank and summit area
of Machias Seamount but are absent from the NE side of the guyot (Fig.
2b). White patches represent smooth, sedimented sea-floor on the
guyot summit and shadow zones on the slopes. Second order features
such as large debris slides and gullies are difficult to discern,
because some detail is lost in the reduction of the original large
mosaic to page-size format.

SIDE-SCAN INTERPRETATION

A line drawing interpretation of the SeaMARC II Mosaic focuses
attention on the detail contained in the side-scan images (compare
Figures 2b and 3b). The drawing clearly displays the WNW-ESE fabric
cutting across, and mostly confined to the SW flank and summit area of
this guyot. These features are probably normal faults, a style of
tectonic fabric typical for this setting. Hinge faults [37] - faults
with near vertical slip planes - may develop at the extremes of the
subduction zone as the subducting and non-subducting portions of the
same plate separate. Shallow seismic events characterize the area
immediately to the west of Machias Seamount (Fig. 1). Focal
mechanisms for events just west of the trench indicate that hinge faulting is a dominant tectonic process [37, 38]. Faulting continues to the trench axis.

The normal faults on the SW flank of Machias Seamount deflect the downslope transport of sediment and locally channel flows into grabens. As a consequence, flows on the SW flank are short and follow arcuate paths. Scarps disappear locally where buried by sediment flow. A ridge at the 6600 m contour is probably a slump covering a section of fault scarp (Fig. 3a). The effect of these offsets on patterns of sedimentation is clearly portrayed in a comparison of sediment flows from opposite flanks of the guyot.

Faults are absent on the NE flank of this guyot and large sediment flows extend radially from the summit to the base of the guyot (Figs 2b and 3b). The roughness depicted in the image suggests that these flows contain large boulders. These large, localized and radial gravity flows are absent from the SW flank of the guyot.

The smoothness and low reflectivity of the summit are suggestive of undisturbed sediment cover. At less than 700 m depth, foraminiferal oozes probably blanket this guyot. The thickness of the sedimentary cap is not clearly discernible on our single-channel seismic coverage. Thickness of sedimentary overburden moving under the influence of gravity is also difficult to assess. The thinnest flow are probably the sheet flows displacing sediment from the NE flank of the summit and thickest flows are probably those on the SW flank. Deeper terraces to the SW of the summit may represent parts of the original flat-lying top of the guyot.
DISCUSSION

Only limited consideration has been given to sedimentary processes on the seaward flank of trenches, probably because many of the classic cross-sections of the setting show a smooth, or mildly offset, sediment-draped sea-floor descendingly beneath the landward slope. Bathymetry along the seaward flank of many trenches, however, displays spectacular relief. Sediment is probably reworked on the seaward slope of the Middle America Trench [39] and bottom currents are strong enough to suspend sediment on the northern slope of the Puerto Rico trench [40], exposing Cretaceous limestone at about 6100 m depth [15]. The chance recovery of Miocene and Eocene foraminiferal ooze in core tops from the seaward flank of the Peru-Chile trench demonstrates the potential for resedimentation [41]; and horsts and grabens on the subducting plates may play a critical role in the eventual subduction of normal and reworked sequences of pelagic sediment [5].

The side-scan image of Machias seamount displays a disruption of structure and sediment cover before subduction, the kind of process suggested in the bathymetric coverage of the Michelson Ridge guyots [27, 4]. This reorganization can mix basement rock with surficial pelagic carbonates, and create a melange prior to any involvement of the sedimentary section and oceanic basement with tectonic processes in the fore-arc.

ACKNOWLEDGMENTS

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subduction line in the Middle America trench. Science, 196:423-426.


[39] Ross DA (1971) Sediments of the northern Middle America


Figure 1. Location map, regional tectonics and seismicity for Machias Seamount. Base map after CCOP/SOPAC Atlas [42]. Computer plot of epicenters south of 14°S latitude and east of 174°W longitude provided by A.S. Murray (BMR). Note the position of Machias Seamount at the northern end of the Tonga Trench where the structural trend is roughly NW-SE.

Figure 2. SeaMARC II coverage of Machias Seamount.
(a) Bathymetric contours are designated in kilometers and contour interval is 100 m.
(b) Mosaic of SeaMARC II side-scan images of Machias Seamount. This survey required about 15 hrs of ship time. Each track line appears as a stripe centered within 10 km of side-scan coverage. High amplitude echos are printed as dark areas in this display. Tie line across summit not shown.

Figure 3. Interpretation of SeaMARC II coverage of Machias Seamount.
(a) Bathymetric contours at 1000 m interval and interpretation of geological structure.
(b) Line drawing interpretation of SeaMARC II mosaic shown in Figure 2b. The interpretation is constructed with shadows cast toward the south and in a stylized fashion highlights the interplay of tectonics and sedimentation.
Dot-dash pattern delineates extent of the trench floor and sub-horizontal sea-floor of low reflectivity to the NE of this guyot. The area appears as flat, gray tones on the side-scan image, despite several hundred metres of relief shown in the bathymetric data over the trench axis.
Location Map
Tectonics; Seismicity

FIGURE 1 (S3).
FIGURE 2 (S3).

See also next page
FIGURE 3 (S3).

**Interpretation**

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SECTION 4

GEOPHYSICS AND GEOLOGY OF MACHIAS SEAMOUNT

Introduction and Previous Work

Machias Seamount (Fig. 1) was first surveyed in 1979 by R.V. 'Machias' during a search for offshore minerals around Samoa (Gauss, 1980) and named during the preparation of the CCOP/SOPAC bathymetric map of the SW Pacific (Kroenke et al., 1983). Two successful dredge hauls were taken near the crest of the seamount, during the 'Machias' cruise. In 1982, R.V. 'Kallisto' surveyed the seamount (named Uo-Matae) and recorded a minimum depth of 760 m (Pushchin et al., 1983). The 'Kallisto' expedition sampled the seamount and adjacent seafloor, ran seismic and magnetic lines, and recorded earthquake activity from ocean-bottom seismometers.

A detailed bathymetric survey at 1:250 000 scale was completed by the RAN Hydrographic Survey in 1985 using M.V. 'Cape Pillar'. The results of this survey were kindly made available to us. The most recent survey of Machias Seamount was that of R.V. 'Moana Wave' (Hawaii Institute of Geophysics) in early 1987, as part of the CCOP/SOPAC Tripartite II program. During this cruise (Coulbourn, Hill et al., 1987), the bathymetry of the entire seamount was surveyed with the SeaMARC II swath-mapping/side-scan system (Blackington et al., 1983). Seismic, gravity and magnetic data were recorded concurrently with the SeaMARC II.
The 1986 'Tui' geological/geophysical program at Machias Seamount included the surveying of two long geophysical (magnetics, gravity and bathymetry) lines across the feature (Fig. 2), and the dredging of a number of sites on the flanks and summit of the seamount to recover a representative suite of FeMn-oxide crusts and rock samples.

**Tectonic Setting**

Machias Seamount lies on the very edge of the Pacific plate, at the sharp westward bend of the Tonga Trench just south of the Samoan Islands. With the Indo-Australian and Pacific plates converging at about 9 cm/year in an E-W direction (AAPG, 1984), Machias is in a highly unstable tectonic position. The seamount is precariously perched on the outer wall of the trench, with its summit only 30 km from the 7.5 km deep trench floor to the SW. The lower SW flank of the seamount has already reached the trench axis and is on the verge of being subducted. Because of the bend in the trench, subduction in the vicinity of Machias is oblique. Assuming continued plate convergence at the current rate, the summit of Machias Seamount is expected to reach the trench axis within about 450,000 years. The seamount may not subduct in its entirety, but partially accrete on the inner (insular) wall of the trench.

The area immediately to the west of Machias is seismically very active, with hypocentres almost exclusively at shallow depth. Farther west, earthquakes show a trend of increasing depth in conformity with the west dipping Benioff zone. Nevertheless, some shallow seismicity persists between Samoa and Fiji, and is attributed to active back-arc spreading in the Lau Basin and movement along a postulated E-W
trending transform fault which offsets the Pacific/Indo Australian plate convergence from the northern end of the Tonga Trench to the New Hebrides Trench.

The shallow seismicity in the Machias Seamount area is believed to be due to three types of earthquake mechanism -

(i) one associated with extensional deformation of the upper surface of the Pacific lithosphere as it flexes prior to subduction (Stauder, 1968; Hanks, 1971; Jones et al., 1979).

(ii) one associated with hinge faulting - a phenomenon localized at the trench bend and produced as the Pacific plate 'tears' along a vertical plane to accommodate subduction of the southern part of the plate beneath the Tonga arc and continued horizontal westward drift of the northern part (Isacks et al., 1969; Johnson & Molnar, 1972).

(iii) one associated with underthrusting of the Pacific plate in a westerly direction along shallow-dipping fault planes (Johnson & Molnar, 1972).

SeaMARC II bathymetry and side-scan images of Machias Seamount reveal extensive fracturing of the seamount by normal faults striking sub-parallel with the local trench axis, and considerable remobilization of the sediment cover produced by the tectonism (Coulbourn et al, in press). Much of the observed structure is thought to be due to the type (i) mechanism, i.e. tensional fracturing of the convexly flexed Pacific ocean crust.
High seismic activity was recorded in the vicinity of Machias Seamount by ocean-bottom seismometers deployed during the 'Kallisto' cruise. Over a recording period of 135.5 hours, 684 seismic events were registered by one instrument stationed just NNW of the seamount (L.N. Smirnov & A.G. Bugaevsky, in Pushchin (1983)). Analysis of S-P times indicates that most of the shocks originated from within 70 km of the seamount summit.

The Samoan Islands, only 120 km north of Machias, were constructed by Pliocene-Recent basaltic volcanism. Volcanism has been active on Savaii and Manua Islands during historic times (Friedlander, 1910; Sterns, 1944). Though such activity at opposite ends of the island chain appears incongruous, existing palaeomagnetic data is broadly supportive of a hot spot origin for the Samoan Islands (Keating, 1985). It was earlier proposed by Kear & Wood (1959) that horizontal stresses generated in the Pacific plate by the differential plate motion at the northern extremity of the Tonga Trench was responsible for the volcanism. Following a similar argument, Natland (1980) suggested that plate deformation caused by subduction at the nearby Tonga Trench influenced or even produced the extensive recent (post-erosional) volcanic activity in Samoa. The evolution of the island chain probably involved both hot-spot volcanism and volcanism induced by plate deformation. The proximity of Machias to both Tonga Trench and Samoan chain suggests a possible link between origin of Samoan volcanism and volcanism at Machias since mid-Pliocene.
Bathymetric Compilation

A comprehensive bathymetric compilation of available data has been prepared for the area of Machias Seamount and adjacent Tonga Trench to the south-west. This bathymetric map at 200 m contour interval (Fig. 3) is based on the 1986 HMNZS 'Tui' coverage (Fig. 2), plus the following data sets:

(a) CCOP/SOPAC 1987 SeaMARC II swath-mapping survey by R/V 'Moana Wave' of the Hawaii Institute of Geophysics (Coulbourn, Hill et al., 1987; Coulbourn et al., in press) - see Figure 2 for ship's tracks.

(b) 1985 hydrographic survey by the Royal Australian Navy Hydrographic Office Detached Survey Unit aboard M.V. 'Cape Pillar' (Ocean Bathymetric Survey - Western Samoa 1:250,000, sheets 4 & 6).

(c) bathymetric data acquired during research cruises -

(i) 1962 R/V 'Vema' (Lamont-Doherty Geological Observatory)

(ii) CCOP/SOPAC 1979 R/V 'Machias' (Hawaii Institute of Geophysics)

(iii) CCOP/SOPAC 1980 R/V 'Machias' (Hawaii Institute of Geophysics)

(iv) 1982 R/V 'Kallisto' (Pushchin et al., 1983)
Gravity and Magnetic Contour Maps

The free-air gravity anomaly map (Fig. 5) and magnetic anomaly map (Fig. 6) of the Machias Seamount/Tonga Trench area were produced from a combination of the 1986 'Tui' and 1987 'Moana Wave' data. The free-air anomaly contours are at 25 mgal interval, while the magnetic anomaly data have been contoured at 50 nT contour interval.

The 'Tui' free-air anomalies are based on the old Potsdam value of absolute gravity (981274 mgal), and were calculated from the 1930 International Gravity Formula. The 'Moana Wave' free-air anomalies are based on IGSN71 (i.e. new Potsdam value of 981260 mgal), and were calculated from the 1967 IGA formula for normal gravity. For low latitudes, as around Machias Seamount, the two reference systems produce a difference of only about 2 mgal in calculated free-air anomaly. During the compilation +2 mgal was added to the 'Tui' gravity values, so that the contours of Figure 5 represent free-air gravity anomaly based on the more recent gravity system (IGSN71/1967 IGA formula).

Both 'Tui' and 'Moana Wave' magnetic data were reduced to IGRF80 magnetic anomalies (Peddie, 1982) prior to compilation of Figure 6.
Bathymetric and Seismic Results

The summit area of the seamount undulates slightly but is roughly flat-topped at a depth of about 750 m. This area is elongate in shape, with its long axis aligned sub-parallel to the local trend of the Tonga Trench axis, and is about 8 km in length and 2.5 km wide (Fig. 3). It was mapped in detail during the 1985 RAN 'Cape Pillar' survey. The shallowest point is a local high located close to the centre of the summit area at a depth of 636 m.

The sides of the seamount are generally steep (10°-35° slopes) and of rugged terrain. Gullies and canyons are deeply incised on the flanks, particularly those facing the trench. The larger canyons are 400-600 m deep and 6-10 km long. Curvilinear escarpments 100-200 m in height intersect the flanks of the seamount along a NW-SE trend. These structural features are most common on the trench-facing side of the seamount, and are interpreted as fault scarps produced by steeply dipping normal faults (Coulbourn et al., in press). The faults are generally down-thrown to the SW.

The slopes of the upper part of the seamount are broken at various levels by a series of prominent terraces. The most notable terrace development occurs on the southern flank of Machias (Figs. 3 & 4), where the slope flattens out onto large platforms 5-25 km² in area. Fault scarps usually bound the terraces on the up-slope side. The terraces are believed to be sections of a flat-topped summit area that was once several times larger than at present. Substantial portions of the summit area were apparently down-faulted as the southern part of the seamount deformed in descending the trench. The terraces on the SW flank tilt roughly 2-4° towards the
trench, whereas the summit area is close to horizontal suggesting that this part of the guyot has, as yet, been little influenced by the descent and concomitant fracturing.

The trench floor SW of Machias lies at an average depth of about 7.5 km. A maximum depth of 7.8 km is attained in a local depression directly down-slope from the seamount. The terrain on the lowermost slopes of the trench is uneven. On the seamount side, one large canyon and a number of smaller ones cut into the base of the seamount; sediment fans associated with the canyon systems extend out onto the trench floor. The Tonga fore-arc side of the trench is characterized by hummocky terrain with local ridges and troughs. The Tonga Trench appears to have diverted to the seaward (NE) side of a large mounded area right at the base of the fore-arc slope.

The 'Moana Wave' seismic profiling of the Machias Seamount area did not reveal any thick sediment accumulations. Results were similar to those obtained during the 1982 'Kallisto' cruise. The seismic data indicate an extensive blanket of well-stratified sediment several hundred metres thick to the north of the seamount. These sediments lie in water depths of about 4500 m and probably comprise hemipelagic silts/clays bearing radiolaria and tuffaceous material from volcanic ash falls (see sampling data). Apart from up to several hundred metres of sediment ponded in local depressions on the trench floor and lower insular slope, no clear evidence of further substantial deposits was found.

The seismic data over the summit area of Machias were closely examined to try to establish the thickness of sedimentary section present. Bubble pulse reverberation has tended to mask much of the
stratigraphic detail, nevertheless, it has been possible to
tentatively identify a reflection event 0.5 sec below seafloor (Fig.
4) as the base of the sedimentary capping. Assuming a sediment
velocity of 2.0 km/sec, this implies a 500 m thick sediment platform
on Machias. This thickness happens to correspond to the thickness of
capping clearly defined by seismic on Capricorn Seamount (see Section
5 of this thesis).

Seafloor Sampling Results

Results of earlier investigations

Exon (1982) has reviewed seafloor sampling activities conducted
in the Samoan region prior to 1982. The Machias Seamount area is
covered by this study, with data from 5 successful stations documented
for the area of Figure 2. Carbonate rocks were recovered from two
dredge station over the summit of the seamount (see below). A
considerable number of sites on, and in the vicinity of Machias
Seamount, were also sampled during the 1982 R/V 'Kallisto' expedition
(Pushchin et al., 1983).

Locations of previous successful seafloor sampling sites are
indicated on the bathymetry map of Figure 3. The following table
provides station data and a brief description of material recovered.
<table>
<thead>
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<th>Equipment</th>
<th>Water depth(m)</th>
<th>Sample description</th>
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</thead>
<tbody>
<tr>
<td>'Penguin' 331</td>
<td>Grab</td>
<td>4631</td>
<td>Volcanic mud with radiolaria</td>
</tr>
<tr>
<td>'Machias' WS-79(1)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Rock dredge</td>
<td>670</td>
<td>Fe/Mn stained cemented coral/algal calcarenite.</td>
</tr>
<tr>
<td>4</td>
<td>Rock dredge</td>
<td>900-1100</td>
<td>Fe/Mn stained coral rock, calcarenite, calcrudite</td>
</tr>
<tr>
<td>7</td>
<td>Free-fall grab</td>
<td>5400</td>
<td>Angular volcanic pebbles to 8 mm diam.</td>
</tr>
<tr>
<td>17</td>
<td>Free-fall grab</td>
<td>5580</td>
<td>Light brown mud; mud pellets</td>
</tr>
<tr>
<td>'Kallisto'</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>16-75</td>
<td>Grab bathometer</td>
<td>7480</td>
<td>Brown clay</td>
</tr>
<tr>
<td>16-77</td>
<td>Gravity corer</td>
<td>7400</td>
<td>Laminated silt/clay</td>
</tr>
<tr>
<td>16-80</td>
<td>Grab</td>
<td>5080</td>
<td>Tuffaceous silt and clay</td>
</tr>
<tr>
<td>16-83</td>
<td>Grab</td>
<td>760</td>
<td>Reef limestone fragments</td>
</tr>
<tr>
<td>16-85</td>
<td>Grab</td>
<td>2000</td>
<td>Basalt rubble</td>
</tr>
<tr>
<td>16-87</td>
<td>Grab</td>
<td>3120</td>
<td>Foraminiferal sand, rare pebbles</td>
</tr>
<tr>
<td>16-88</td>
<td>Grab</td>
<td>3600</td>
<td>Carbonate/silt mixtite, tuffaceous silt</td>
</tr>
<tr>
<td>16-89</td>
<td>Gravity corer</td>
<td>4200</td>
<td>Carbonate/silt mixtite</td>
</tr>
<tr>
<td>16-90</td>
<td>Grab</td>
<td>2080</td>
<td>Carbonate/silt mixtite</td>
</tr>
<tr>
<td>16-94</td>
<td>Dredge</td>
<td>6000-5800</td>
<td>Basalt, basalt tuffs,</td>
</tr>
<tr>
<td>Dredge</td>
<td>Depth</td>
<td>Description</td>
<td></td>
</tr>
<tr>
<td>-----------</td>
<td>-------</td>
<td>----------------------------------</td>
<td></td>
</tr>
<tr>
<td>16-95</td>
<td>5400-5200</td>
<td>tuffaceous sandstones</td>
<td></td>
</tr>
<tr>
<td>16-96</td>
<td>4800-4500</td>
<td>Basalt (black, porous)</td>
<td></td>
</tr>
<tr>
<td>16-97</td>
<td>3600-3400</td>
<td>Basalt (black, porous), in-situ</td>
<td></td>
</tr>
<tr>
<td>16-98</td>
<td>2400-2200</td>
<td>Reef limestones and corals</td>
<td></td>
</tr>
<tr>
<td>16-99</td>
<td>5800-5600</td>
<td>Basalt (porous), in-situ</td>
<td></td>
</tr>
<tr>
<td>16-100</td>
<td>4000-3800</td>
<td>Basalt (porous)</td>
<td></td>
</tr>
</tbody>
</table>
1986 'Tui' sampling results

Sample was recovered from six of the seven rock dredge stations on Machias Seamount. The locations of the successful stations are indicated on Figure 3, and a summary description of material recovered is provided below.

<table>
<thead>
<tr>
<th>Station</th>
<th>Depth</th>
<th>Sample description</th>
</tr>
</thead>
<tbody>
<tr>
<td>U341b</td>
<td>2495-2500</td>
<td>Altered tuff; coral fragments</td>
</tr>
<tr>
<td>U342</td>
<td>1350(?2100)</td>
<td>Carbonate rubble.</td>
</tr>
<tr>
<td>U343</td>
<td>730-715</td>
<td>Limestone, some Mn stained.</td>
</tr>
<tr>
<td>U344</td>
<td>675-670</td>
<td>Carbonate sand; 1 pebble coral; 1 pebble pumice.</td>
</tr>
<tr>
<td>U345</td>
<td>1972-2166</td>
<td>Coral rubble; basalt cobble with dunite inclusion, fresh basalt pebbles.</td>
</tr>
<tr>
<td>U346</td>
<td>4077</td>
<td>Fresh basalt cobbles and pebbles, dunite inclusions; some coral rubble.</td>
</tr>
</tbody>
</table>

Age determinations

Limestone samples dredged from station 16-98 during the R/V 'Kallisto' cruise were palaeontologically examined for dateable fauna. O. Tkalich (in Pushchin et al., 1983) reports the presence of foraminifera Globigerinoides trilobus (Reuss), G. ruber (d'Orb), Globigerinite glutinata (Egger) and Globohorotalia sp. These species occur over the broad stratigraphic interval of Miocene - Holocene. Corals identified (E. Krasnov in Pushchin et al., 1983) include recent
shallow-water (20-50 m), reef building *Stylophora ex gr. mordax* Dana and *Fungia* sp., as well as Miocene-Holocene *Coniocorella* sp., and *Agaricia rugosa* Lam. Of special significance was the identification of Miocene *Progyrosmilia torulosa* Gerth. The results date the carbonate capping on Machias Seamount as Miocene-Holocene, and imply that subsidence of Machias Seamount from near sea-level to its present depth of about 750 m occurred during the last 2 m.y.

K-Ar radiometric dating of fresh basalt from station U346 (NW of Machias Seamount, water depth 4077 m) by C. Adams of the Institute of Nuclear Sciences, NZ DSIR (Lower Hutt) gives an age of 1.42+/-0.08 m.y.

The Gravity and Magnetic Fields

Main features

Free-air gravity anomalies (Fig. 5) correlate well with the seamount and trench topography. A gravity high of +125 mgal is located over the summit of Machias Seamount, while over the seamount flanks the closed gravity contours largely replicate the pattern of the bathymetry contours. The field decreases rapidly toward the trench axis and bottoms out at -225 mgal in a broad low just SW of the trench axis. As expected, the contours over the lower trench trend parallel to it.

The magnetic anomaly field is of relatively low amplitude for a seamount the size and relatively shallow depth of Machias. Over the seamount, magnetic anomaly ranges from about -100 nT to +150 nT. The short wavelength anomalies over the seamount compare with the broad,
long wavelength anomaly pattern over the deeper parts of the trench, and reflect the differences in depth to magnetic basement. The field over the seamount is more complex than a simple dipole field, though the general low (-50 to -100 nT) just north of the summit area and the general high (100 to 150 nT) just south of the summit area suggest that the overall character of the field is dipolar. The observed pattern is consistent with a volcanic pedestal to Machias Seamount that has a primary reverse magnetization.

Three-dimensional modelling

Gravity and magnetic modelling of Machias Seamount was undertaken to estimate basic density and magnetization parameters of the edifice, and to identify any major internal structural or lithological boundaries. The methods and computer programs developed by Plouff (1976) for calculating gravity and magnetic fields of polygonal prisms were applied to our study.

The topography of Machias Seamount was approximated by a stack of nine polygonal slabs as indicated in Figure 7. The upper and lower surface of each slab is horizontal, and the sides are vertical. The vertices, which define the polygonal outlines of the slabs, were scaled from the bathymetric contours of Figure 3. Together, the nine slabs represent a composite body of depth extent 0.8-5.0 km.

Assuming regional isostatic compensation of the seamount load (implicit in the geometry of the model adopted) and uniform density, a least-squares estimate of seamount density was made. For this calculation the observed data set comprised 126 gravity values interpolated from the contours of Figure 4 over a 2 km x 2 km grid
within the rectangular area WXYZ centred on the summit of Machias Seamount as indicated in Figure 7. The observed data grid was not extended farther south to minimize biasing effects of the high gravity gradient south of the seamount. The calculations produced a value of $2.48+/-0.04 \text{ t/m}^3$ for the seamount density (assuming seawater density of $1.03 \text{ t/m}^3$), with correlation coefficient 0.96 and gravity datum of -51 mgal. Theoretical contours corresponding to this best-fit gravity model are displayed within the large square area on Figure 7. The match between these contours and those of the observed field (Fig. 5) is generally very good, as expected. The significant mis-match obvious in the SW corner is due to the gravitational effects of the trench not accounted for in the seamount model adopted. The difference between observed and theoretical (model) gravity is best seen in the residual gravity map (Fig. 8). The trench effect shows up as the pronounced SW-NE gravity gradient. The low residual gravity over the seamount and the absence of correlation with seamount topography attest to the suitability of the gravity model. The residual gravity map shows no evidence of major anomalous masses within the seamount so it appears that Machias Seamount is of fairly uniform density. Lateral variations in density, in particular, appear to be minimal.

The magnetic modelling was done in an analogous manner to that of the gravity modelling. Uniform magnetization of the model representing the volcanic pedestal of the seamount was assumed. The observed magnetic anomaly data for the least-squares estimation of magnetic parameters were interpolated from the contours of Figure 6 over the rectangular area centred on the seamount summit as shown in Figure 9. An area smaller than that used for the gravity analysis was selected because of the poorer magnetic data distribution over the
western flank. As in the gravity case, values were scaled over a 2 km grid within the selected area. This yielded 104 magnetic data points for the least-squares analysis. A number of models were tested to simulate an increase in thickness of non-magnetic capping on the seamount. This was achieved by removing the upper-most polygonal slab (A, B, C etc in Figure 7) from successive models tested. The best solution was obtained with the top of the magnetic body at 1.4 km depth (i.e. slabs C-I). For this solution the multiple correlation coefficient was 0.75 and magnetic datum level 13 nT. The best-fit intensity of total magnetization was 1.2 A/m, declination 185° and inclination 5°. The calculated field for this model is depicted in Figure 9.

As mentioned, it appears that the volcanic edifice of Machias does possess a degree of magnetic heterogeneity. As a result, the fit of modelled data to observed data is only fair. In addition, the model solutions are not highly sensitive to variation in depth of the magnetic body, at least for the realistic range of depths trialled. Thus 1.4 km is considered to be only an approximate estimate for depth to magnetic basement.

Because some magnetic heterogeneity is present in the volcanic edifice, the usefulness of the calculated magnetization direction is limited for making palaeomagnetic deductions. The declination of 185° is almost meridional, suggesting that the edifice has suffered little, if any rotation since it was constructed. The inclination of 5° is lower than expected. The field of a geocentric axial dipole would have an inclination of 28.1° at Machias, in its present position. The relatively low calculated declination is probably best explained by anomalous magnetization within Machias rather than a southerly drift.
of the Pacific plate on which it rests.

**Magnetic source depths from Werner deconvolution**

The automatic interpretation technique of Werner deconvolution (Jain, 1976; Hsu & Tilbury, 1977) was applied to the magnetic profile of the long NE-SW 'Tui' line which crosses the summit area of Machias Seamount (Fig. 2). Magnetic source depths (as well as magnetic intensity and dip parameters) were calculated for both interface (fault, contact) and thin sheet (dyke) models. Plots of estimated magnetic sources, together with observed magnetic profile and bathymetry, are presented in Figure 10. The symbol size is proportional to the magnetic intensity of the estimated source. Clusters of estimates are indicative of a reliable solution, while isolated estimates are generally not highly significant.

The low intensity estimates within the water-column are due to minor fluctuations in the magnetic profile produced by normal magnetic noise levels. Other estimates, also appearing in the water-column but located close to the seafloor, may be due to elevated topography adjacent to the ship's track. In general, the magnetic estimates correlate well with the surface and interior of the seamount. The most important parameter that the Werner deconvolution can provide is the depth to the magnetic volcanics beneath the summit of Machias Seamount. Though the estimates do not clearly define the non-magnetic /magnetic interface, it appears to lie in the depth range of 0.9-1.4 km.
Discussion of Results

**Thickness of carbonate capping**

Basalt was recovered in dredge hauls from as high on the seamount as 2000 m (station 16-85) and 1972-2166 m (station U345). These results imply a maximum thickness of about 1250 m for the capping, assuming that the basalt had not intruded the capping. The magnetic modelling/Werner deconvolution indicates a thickness in the range 150-650 m. Thus a value of 500 m inferred from the seismic data appears to be a reasonable and consistent estimate.

**Lithological composition/physical properties**

Basically, the data portray Machias as a large (4 km high) volcanic edifice capped by a relatively thin (500 m) carbonate platform. The platform is composed of partially recrystallized, reefal limestones mantled by a veneer of unconsolidated foraminiferal sand and coral/carbonate debris. Samples dredged from the flanks show that the volcanic pedestal is composed mainly of basalt. Both vesicular and massive types are present (in about 1:1 ratio); some contain ultramafic inclusions. Tuffs are represented in the dredge samples and so appear to be a minor lithological component of the volcanic edifice.

The modelling results indicated a mean density of 2.48 t/m$^3$ for the entire seamount and magnetization of 1.2 A/m for the volcanic pedestal. Laboratory measurements of density made on basalt samples dredged from Machias during the 'Kallisto' cruise gave an average value of about 2.45 t/m$^3$. Basalts within the interior of the
edifice are likely to be less vesicular and less weathered than the outcrop/talus dredged and so have a slightly higher density. Such higher density is countered, however, by the lower density carbonate capping and tuff component of the seamount. Thus the laboratory determined 2.45 t/m$^3$ is probably a good estimate for the mean density of the whole seamount, and agrees closely with the 2.48 t/m$^3$ estimated by modelling. The calculated pedestal magnetization is significantly lower than for typical Pacific seamounts. This is attributed to mixed magnetic polarities of basalt flows, dykes and other internal igneous structures emplaced during different intervals in the earth's normal/reversed polarity cycle.

The basalt recovered at station U346 and dated as 1.42+/-0.08 m.y. would have been emplaced during the 0.95-1.62 m.y. reversed polarity period (Labreque et al., 1977). The scale of this relatively young volcanism is not known, but it may have made a significant contribution to the general reversed magnetization of the seamount. Igneous intrusions associated with this episode may have reset some of the pre-existing thermoremanent magnetization by the heating of wall rocks. The presence of dunite inclusions within fresh basalt recovered in the same dredge haul (and also at station U345) suggests that this late volcanic activity is partly sourced directly from the mantle. This is in keeping with the major lithospheric deformation taking place close to Machias. The 'hinge' faulting, in particular, cuts right through the entire lithosphere potentially allowing access of mantle magmas to fractures and other conduits at higher levels in the crust.
Absence of Mn crusts

Thin FeMn-oxide coatings or stains were observed on some samples from nearly every dredge station. Such coatings were most prevalent within recesses on coral or limestone fragments. No manganese crusts were recovered from Machias Seamount, however.

The absence of crusts is attributed to the relatively young surfaces on the seamount. Crusts have not had time to develop fully. SeaMARC II images of the seamount (Coulbourn et al., in press) show extensive faulting, erosional features such as canyons and gullies, as well as sediment flows and fans. The high mobility of the debris flows is indicated by the recovery of mixtures of very well rounded to angular pebbles and cobbles in dredge hauls. Mass wasting is highly active on Machias as a result of the tectonic deformation. The NE flank is least affected and may show some crust development. It was not dredged during our cruise.

Quaternary subsidence

Examination of the bathymetric contours on the upper trench slope adjacent to Machias Seamount indicates that 750 m of subsidence is readily explained by descent of the seamount down the trench slope. Machias has moved westward about 15 km since going over the trench edge. Such motion of the Pacific plate corresponds to a time interval of approximately 170,000 years. The mean subsidence rate would have been about 4.4 mm/year. The presence of dead modern reef-building corals on the seamount summit indicates that upward growth of the reef could not match a rapid submergence. This rapid submergence which 'drowned' the reef would have occurred by a
combination of seamount subsidence (4.4 mm/year) coupled with a high (but unknown) rate of eustatic sea-level rise. Coral reef accretion rates in the tropical Pacific during the Holocene have been 3-10 mm/year (Adey, 1978), so it would only take a modest rate of eustatic sea-level rise to allow submergence to exceed accretion.

Comparison with Capricorn Seamount and Niue

Capricorn Seamount and Niue (see Sections 5 & 6 in this thesis) are located about 500 km to the south. Machias, Capricorn and Niue all lie on the Pacific plate, are about the same size and shape, and are capped by coralline limestone platforms. Palaeontological evidence indicates that these platforms are all of Miocene age, while geophysical and other data indicate a similar thickness of roughly 500 m.

These similarities suggest that the pre-Pleistocene evolutionary history of all three seamounts has followed a similar path, particularly in relation to subsidence. Epeirogenic movement of the Pacific plate appears to have been much the same between Miocene-Pleistocene at all three seamounts. Such movement need not have occurred simultaneously at the three locations, but may have involved time lags of up to a few million years. The analogy between Machias and Capricorn can be extended through the Quaternary because of their almost equivalent tectonic situations on the outer wall of the Tonga Trench.
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List of Figures

Figure 1. Location map and regional tectonic setting. Bathymetry base map after Kroenke et al. (1983); contour interval is 500 m.

Figure 2. HMNZS 'Tui' and R/V 'Moana Wave' geophysical survey lines across Machias Seamount. Ten minute positions are indicated as box symbols along the 'Tui' tracks and as cross symbols along the 'Moana Wave' tracks, respectively. Hourly positions are annotated with the survey number (61 for the 'Tui' cruise and 87 for the 'Moana Wave' cruise), Julian day and GMT.

Figure 3. Bathymetry of the Machias Seamount area at 200 m contour interval. Seafloor sites sampled during the HMNZS 'Tui', R/V 'Kallisto', R/V 'Machias' and R/V 'Penguin' expeditions are indicated. Sampling stations are annotated with station number, together with consolidated/semi-consolidated lithotypes recovered according to the broad classification - basalt (B); limestone (L); tuff (T).

Figure 4. Seismic profile from the R/V 'Moana Wave cruise across Machias Seamount. The profile represents a SW-NE oriented section comprising segments S4-S3 and S1-S2 (see Figure 3 for locations). The seismic section for segment S4-S1 (over the Tonga Trench) is depicted as a line drawing since it was shot in a direction opposite to that of S1-S2.
Figure 5. Free-air gravity anomaly at 25 mgal contour interval.

Figure 6. Total magnetic intensity anomaly at 50 nT contour interval, based on the IGRF80 reference field.

Figure 7. Gravity anomaly (at 25 mgal contour interval) for seamount density of 2.48 t/m$^3$. The seafloor topography is represented by polygonal bodies (A-I) with depth extents (km) as follows - A 0.8-1.0, B 1.0-1.4, C 1.4-1.8, D 1.8-2.2, E 2.2-2.8, F 2.8-3.4, G 3.4-4.0, H 4.0-4.6, I 4.6-5.0. The polygonal outlines are shown as dashed lines and the vertices as dots. A regional gravity correction of -51 mgal has been added.

Figure 8. Residual gravity at 10 mgal contour interval.

Figure 9. Magnetic anomaly (at 25 nT contour interval) for a model represented by polygonal slabs C-I (depth extent 1.4-5.0 km). Magnetization parameters: intensity of magnetization 1.2 A/m, declination 185°, inclination 5°. A regional magnetic correction of 13 nT has been added.

Figure 10. Werner deconvolution magnetic source estimates for the SW-NE magnetic profile of HMNZS 'Tui' cruise across Machias Seamount (see Figure 2 for location). The magnetic profile and bathymetry along the line are indicated.
FIGURE 1 (S4).
MACHIAS SM

TUI & MOANA WAVE TRACKS

FIGURE 2 (S4)
FIGURE 3 (S4)

Depths in Km
200 m contour interval

SAMPLING STATIONS

* TUI
* KALLISTO
* MACHIAS
* PENGUIN

Bathymetry
FIGURE 5 (S4).
Magnetic Anomaly

FIGURE 6 (S4).
Gravity anomaly for seamount density 2.48 t/m³

FIGURE 7 (S4).
Residual Gravity

FIGURE 8 (S4).
Magnetic Anomaly for Model

FIGURE 9 (S4).
FIGURE 10 (S4).
SECTION 5

CAPRICORN SEAMOUNT - GEOLOGY AND GEOPHYSICS
OF A SUBDUCTING GUYOT

Introduction

Capricorn Seamount (Fig. 1) was first surveyed by R.V. 'Horizon' in 1953 during the Scripps Institution of Oceanography Capricorn expedition and named after the expedition (Raitt et al., 1955). The seamount was shown to rise more than 8,500 m above the base of the Tonga Trench and 5000 m above the level of the seafloor to the east. It was seen to be approximately flat topped with the surface sloping gently (45° - 1°) to the west. A minimum depth of 395 m was recorded. Three dredge hauls were attempted but all were unsuccessful. Menard (1964) subsequently referred to this feature as Capricorn Guyot.

In 1958, RNZFA 'Tui' carried out further sounding and dredged seafloor sample from a depth of 843 - 880 m (Kustanowich, 1962; Brodie, 1965). A schematic bathymetry was drawn (Brodie, 1965). The main flat-topped area of the seamount was shown to be in the depth range of 800 - 1,000 m and have a diameter of about 18 km. The minimum depth (395 m) was seen to correspond to a flat-topped knoll on the eastern margin of the seamount's summit area. Brodie (1965) recognized the 450 m and 900 m surfaces of the seamount as being wave cut and supported this conclusion with the observation of smooth limestone pebbles in the dredge haul. Manganese oxides completely

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covered the limestone pebbles. Brodie also attempted to interpret the geological history of the seamount.

In 1980, Cullen and Burnett (1986) dredged two stations on Capricorn Seamount in a search for phosphorite nodules but recovered only pumice.

The 1986 HMNZS 'Tui' geophysical coverage is depicted on the track map of Figure 3. All lines shown were surveyed with gravity, magnetics and 12 kHz PDR bathymetry. In addition, single-channel seismic profiling was completed along the E-W line over the summit area of Capricorn Seamount and across the Tonga Trench (X - X'), and also along the S-N line over the summit area (Y-Y').

The seafloor sampling program at Capricorn Seamount was aimed at recovering representative seamount material including FeMn-oxide crusts. Seven dredge hauls were made at various locations on the seamount and each produced sample return.

**Tectonic Setting and Seismicity**

The seafloor morphology (Fig. 1) and seismicity (Fig. 2) of the area in the vicinity of Capricorn define the major tectonic elements and reflect their tectonic interaction. The Pacific plate is subducting beneath the Tonga Ridge in an approximately E-W direction at a rate of 9+ cm/year (AAPG, 1984). This convergence is the product of back-arc spreading in the Lau Basin and large-scale relative motion of the Indo-Australian and Pacific plates.

Capricorn Seamount is situated right on the edge of the
The seismicity map (Fig. 2) shows the very high earthquake activity west of the Tonga Trench axis associated with underthrusting of the Pacific plate beneath the Tonga Ridge (Isacks et al., 1969). Earthquakes are shallow (0-70 km) along the trench, but deepen to the west in conformity with a Wadati-Benioff zone that dips to the west at about 45° for focii in the shallow-intermediate depth range (Hamburger & Isacks, 1987). Though less intense than to the west, seismic activity east of the trench axis is still moderately high within about 150 km of the axis (Fig. 2). Capricorn Seamount is located within this zone. Earthquakes to the east of the trench axis are all shallow (less than 70 km deep) and result from internal deformation of the Pacific plate. Focal mechanism studies (Johnson & Molnar, 1972; Chen & Forsyth, 1978) indicate two types of focal-mechanism solution, (i) normal faulting on fault planes striking approximately N-S, and (ii) thrust faulting, also on northerly striking planes, but dipping steeply to the west. Bending of the Pacific plate as it enters the trench and longitudinal compressive stress within the plate are believed responsible for these styles of structural deformation.

Bathymetric rises and gravity highs on the seaward side of deep-sea trenches have been modelled as upward flexure of the oceanic lithosphere behaving as a thin elastic sheet overlying a fluid substratum (Hanks, 1971; Watts & Talwani, 1974). An outer rise and
gravity high are present east of the trench in the Capricorn Seamount area (see Chase et al., 1982 and Haxby, 1987), but are not clearly defined because of local variations in both bathymetry and gravity field, particularly to the north of the seamount. From a bathymetric profile south of Capricorn, Dubois et al. (1975) estimate the amplitude of the bathymetric bulge as 240 m. This suggests that Capricorn was uplifted by a similar amount on its approach to the trench.

**Bathymetry Compilation**

A new bathymetric map of the Capricorn Seamount area, including adjacent Tonga Trench, has been produced (Fig. 4). The map is based on the 1986 'Tui' data recorded along tracks shown in Figure 3. The compilation includes extra detail of the Capricorn Seamount summit area provided by soundings from the 1953 R/V 'Horizon' survey (Rait et al., 1955), together with infill of remaining bathymetry data gaps utilizing the 200 m contours shown on the map of Chase et al. (1982).

**Gravity and Magnetic Contour Maps**

Contour maps of free-air anomaly (Fig. 7) and magnetic anomaly (Fig. 8) for Capricorn Seamount and adjacent Tonga Trench were prepared solely from the 1986 'Tui' data. This was because very few other research cruises had recorded magnetic and gravity data over the area, and the limited data that were available appeared to be of poor quality or to have navigational inconsistencies.

The free-air anomalies are based on the old Potsdam datum of 981274 mgal and were calculated from the 1930 International Gravity
Formula. The magnetic anomalies are based on the IGRF80 global reference field (Peddie, 1982).

Morphology and Seismic Evidence

As the bathymetric contours (Fig. 4) and seismic sections (Fig. 5) indicate, Capricorn Seamount is a large guyot of 100 km basal diameter located on the seaward wall of the Tonga Trench. The centre of the seamount is only 45 km from the trench axis. The trench is about 9 km deep in the region, with local shallowing of the trench bottom by several hundred metres immediately west of the seamount. This shallowing is probably due to the combined effect of (i) the lower seamount flank starting to subduct, and (ii) a build-up of talus derived from the disintegrating slopes of the seamount.

The lower trench is V-shaped in cross-section adjacent to Capricorn Seamount, with both insular and seaward slopes inclined at about 9.5°. To the north and south of the seamount the insular slope remains much the same, while the seaward slope decreases markedly to about 4.5°. The upper flanks of the seamount are relatively steep with gradients generally in the range 15°-35°. The slopes become more gentle towards the base of the seamount, particularly on the western and southern sides where the lower slopes have gradients in the order of only 3°-4°. At least several local topographic highs with 100-300 m relief exist on the trench-facing flank of Capricorn (Figs. 4 & 5).

The summit area of the seamount is flat-topped at two levels. Both surfaces dip gently towards the trench. The larger surface lies at a depth of about 800-1000 m and is inclined at approximately 1.7° towards the trench (Fig. 6). The smaller (22 km²), but higher
surface lies to the east of the summit area at about 450 m depth (Brodie, 1965). Our shallowest depth recorded for Capricorn Seamount (440 m) was over a point on the southern margin of this elevated area.

As seen in the seismic enlargements and corresponding line drawings (Fig. 6), the summit area is underlain by a sedimentary section 0.5 sec thick. Assuming a sediment velocity of 2.0 km/sec, this thickness translates to 500 m. On the basis of the sampling data (see next section), it appears that the sediments comprise reefal limestones, while seismic basement corresponds to basaltic rocks of the volcanic substructure.

A system of closely-spaced, steeply-dipping faults affects both volcanic basement and sedimentary section. The faults strike predominantly in a N-S direction, parallel to the trench axis. Beneath the planar summit area fault throws are typically 20-100 m, but beyond this area (on the upper flanks of the seamount) the magnitude of throw appears to increase to several hundred metres or more. The faults are interpreted as normal faults produced by brittle fracture of the upper crust in response to tensional stress created by flexure of the Pacific plate as it enters the trench. Such fault systems have been recorded seaward of other trenches in the Pacific (Stauder, 1968; Hanks, 1971, Jones et al., 1979; Fryer & Smoot, 1985).

Seabeam and seismic mapping of an area over the southern Tonga Trench by the 1986 SEAPSO V expedition (Herzer, 1986) revealed classic horst and graben structures trending N-S which are attributed to bending of the Pacific plate at the trench. At the northern end of
the Tonga Trench, SeaMARC II images of Machias Seamount indicate considerable tectonic dismemberment by normal faulting as the guyot descends the outer trench wall (Coulbourn et al., in press).

The 1.7° westward tilt of the summit area is replicated by the bedding within the sedimentary capping and also by the basement surface. This tilt is believed to be due to movement of the seamount down the inclined outer wall of the trench. The parallelism of basement, bedding and seamount summit suggests typical atoll evolution followed by 'drowning' - involving sub-aerial planation of an oceanic volcano, subsidence and upward growth of coral reef forming a limestone platform, and finally rapid subsidence below the photic zone preventing further coral development.

The bedding within the sedimentary capping is best seen in the S-N seismic section which has not suffered major disruption by faulting. The section is well stratified, with beds sub-parallel and mainly gently undulating. Many of the reflectors can be followed almost right across the width of the summit area. The presence of onlap halfway down the sedimentary section (Fig.6) suggests a lateral facies change or unconformity at the base of this sequence; the latter possibility may signify a period of emergence. The general continuity of reflectors indicates that much of the section may comprise lagoonal sediments. Fringing reef buildups may have largely been faulted off the sides of the summit area. Strong hummocky reflectors near the base of the section may represent early reef development.

Sediment thickness at the base of the seamount to the east, north and south appears to be about 0.3-0.5 secs. Bedding character
ranges from hummockly to chaotic - no distinct reflector sequences are present. The sediments are probably fan and slump deposits derived from mass wasting and tectonic erosion of the seamount flanks. Basement is not clearly defined but appears to be irregular and faulted.

The seismic data over the inner (insular) trench wall show no clear evidence of a substantial sediment thickness, possibly because of the masking effect of the steep and rugged seafloor topography. Small sediment ponds in local depressions on the insular slope can be seen, however. These contain up to several hundred metres of recent sediment which appears to dip gently \((0.5^\circ - 1^\circ)\) away from the trench axis, suggesting possible tilting back of the 'accretionary' prism to accommodate subduction of the lower western flank of Capricorn. Such back-tilting of the inner trench wall has been observed at the southern end of the Tonga Trench where seamounts of the Louisville Ridge are being subducted (Herzer, 1986).

**Dredging Results**

A total of seven sites were dredged on Capricorn Seamount. Station locations are indicated in Figure 4. Three stations (U347-U349) were sited on the NNW flank, another three (U350, U351a & U351b) were positioned on the summit area, while the last (U352) was sited on the lower southern slope of the seamount.
The dredging results are summarized below.

<table>
<thead>
<tr>
<th>Station</th>
<th>Depth (m)</th>
<th>Sample description</th>
</tr>
</thead>
<tbody>
<tr>
<td>U347</td>
<td>3905-3864</td>
<td>Nanno-bearing abyssal clay</td>
</tr>
<tr>
<td>U348</td>
<td>2825-2640</td>
<td>Amygdaloidal basalt/welded tuff</td>
</tr>
<tr>
<td>U349</td>
<td>1873-1853</td>
<td>Brecciated basalt.</td>
</tr>
<tr>
<td>U350</td>
<td>1021-922</td>
<td>Coralline limestone block &amp; fragments, slight Mn staining.</td>
</tr>
<tr>
<td>U351a</td>
<td>996-976</td>
<td>Calcareous sand with pteropod shells &amp; pumice.</td>
</tr>
<tr>
<td>U351b</td>
<td>942-932</td>
<td>&quot;</td>
</tr>
<tr>
<td>U352</td>
<td>4684-4689</td>
<td>Coralline limestone rubble; basaltic gravel; calcareous abyssal clay.</td>
</tr>
</tbody>
</table>

Our dredging data is supplemented by that of the 1958 RNZFA 'Tui' expedition, during which a single dredge haul recovered seafloor sample from a depth of 843-880 m at the SW corner of the gently sloping summit area (Kustanowich, 1962; Brodie, 1965). The material recovered included flat limestone pebbles coated with Mn-oxide, pumice, foraminiferal sand and dead solitary corals. The limestone contained remains of molluscs, brachiopods and large foraminifera - but no corals. The brachiopods and foraminifera date the primary fragments in the limestone as Miocene. The matrix may be of
In summary, the dredging results indicate that Capricorn Seamount is a volcanic edifice of basaltic composition, capped by a Miocene limestone platform. Unconsolidated sediments of foraminiferal sand, shell & coral fragments and pumice mantles the relatively flat-lying limestone platform. The lower slopes of the seamount are covered by abyssal clays and sediment fans containing limestone and volcanic erosional products transported from higher levels on the seamount by debris flows.

The shallowest recovery of basalt was from station U349 (1837-1853 m). This implies a maximum thickness of about 1000 m for the limestone platform, and is consistent with the seismically determined thickness of approximately 500 m. The brecciated nature of the basalt from U349 may be an example of the extensive normal faulting seen in the seismic sections.

**Gravity Field and Modelling Results**

Free-air anomaly and sea-floor topography in the area of Capricorn Seamount and adjacent Tonga Trench correlate very well, as can be seen by comparing Figure 4 (bathymetry) and Figure 7 (gravity contours). A gravity high of +246 mgal is centred over the summit of Capricorn Seamount. The field decreases rapidly over the flanks of the seamount, particularly so on the trench side, and attains a minimum of about -225 mgal in a broad gravity trough directly above, and aligned with the trench axis. Over the relatively flat seafloor to the east of Capricorn (5500 m water depth), the free-air anomaly
is correspondly smooth and averages about -10 mgal.

Three-dimensional gravity modelling of the seamount was undertaken to assess mean density and to reveal any anomalous mass distributions within the edifice. For the modelling we utilized the method of Plouff (1976), which enables calculation of gravity fields for geologic bodies of complex shape by approximating them by a set of polygonal prisms. The method includes provision for inversion of observed gravity data to yield least-squares estimates of density. In modelling Capricorn we assumed uniform density and regional isostatic compensation of the seamount load.

The seamount topography was approximated by a stack of 9 polygonal slabs (A-I, see Figure 9). The outline (in plan) of each polygonal prism closely matches the bathymetric contour (Fig. 4) corresponding to the depth of the prism's upper surface. The composite body has a depth extent of 0.6-7.0 km. For the gravity inversion, the observed input data consisted of 143 gravity values interpolated from the gravity contours (Fig. 7) over a 4 km x 4 km grid within the rectangular area shown in Figure 9. The area selected was centred on the seamount summit and limited in its E-W extent to minimize undesirable biasing effects on the calculations due to the relatively steep gravity gradient over the outer trench wall.

The modelling produced a value of 2.56+/-0.02 t/m³ for the best-fit density (assuming sea-water density of 1.03 t/m³), with correlation coefficient 0.99 and gravity datum of -86 mgal. The theoretical gravity for this best-fit model is shown in Figure 9, while Figure 10 depicts contours of residual gravity (observed - theoretical).
The residual field is fairly flat, particularly over the summit region, and this serves to confirm the general validity of the gravity model. The minor decrease in residual gravity toward the eastern margin of Figure 10 is due to uncorrected effect of the non-linear gravity gradient over the trench wall. The -20 mgal residual gravity low just to the east of the summit area, and the +30 mgal high just NW of the summit area, may represent genuine mass deficiency and mass excess, respectively. Insufficient data control may partly be responsible for the +30 mgal high, since the high is located between survey lines.

The calculated mean density of 2.56 t/m$^3$ is consistent with a seamount composed largely of basaltic rocks, including dense flows and intrusions.

**Magnetic Field and Interpretation**

The magnetic field over Capricorn Seamount (Fig. 8) is complex. The IGRF80 anomaly range is approximately -350 to +50 nT. Over the trench axis to the west, the field is seen as a broad trough -150 nT deep trending parallel to the trench. To the north and south of Capricorn the field is relatively flat with long wavelength variations only in the order of about 50 nT. East of Capricorn the field is moderately anomalous.

The anomalous field east of Capricorn is of medium-long wavelength with local peak-peak variations of up to 400 nT. In this area the seafloor lies at a depth of about 5.5 km and is fairly flat. The anomalies are therefore not sourced by seafloor topography.
However, judging from some fairly short-wavelength anomalies in the field, magnetic inhomogeneity must be present at fairly shallow depth in the oceanic crust. The magnetic sources show no significant expression in the free-air gravity profile. The anomalies may represent oceanic magnetic lineations produced by geomagnetic field reversals - lineations that have not been recognised previously. Such a notion is supported by findings of the SEAPSO V expedition (Herzer, 1986) which reported similar unexplained E-W trending magnetic anomalies on the edge of the Pacific plate about 700 km south of Capricorn - on supposed Cretaceous magnetic quiet zone crust (AAPG, 1984). The possibility that the anomalies recorded may be due to ionospheric effects has been considered but discounted, since the anomalies are of too high amplitude to be diurnal variations, and too smooth to be the result of magnetic storm activity.

Capricorn Seamount may be constructed on oceanic crust similar to that to the east. In which case, some of the complexity of the field over Capricorn may be explained by magnetic heterogeneity of the underlying crust. An attempt was made to model the seamount as a uniformly magnetized volcanic edifice topped by a non-magnetic platform of variable thickness. For this the three-dimensional magnetic modelling method of Plouff (1976) was used with similar polygonal prism representation of the volcanic pedestal as in the gravity case, but with the omission of some of the upper polygonal slabs to simulate a non-magnetic capping. No reasonably close or realistic fits to the observed data were achieved, confirming the inhomogeneous magnetic nature of the volcanic edifice or underlying crust, or both.

Despite the complexity, there does appear to be a rough
symmetry to the field about a N-S line through the centre of the seamount. The pair of dipolar anomalies over the eastern flank (-350/+50 nT) and western flank (-300/-80 nT) cannot readily be reconciled with a seamount magnetization that is radially symmetric. The anomalies may be due to isolated magnetic bodies beneath the flanks or could be part of an E-W trending magnetic lineation pattern originating from the underlying oceanic crust. Although the seamount is of composite magnetization, the general N-S alignment of bipolar anomalies suggests that the magnetization is predominantly meridional and either normal or reversed. The meridional magnetization implies little or no rotation of the seamount since construction.

The local magnetic high over the northern part of the summit area and the magnetic low over the southern part of this area signify that the seamount may possess a normally magnetized core, the upper part of which has a diameter of about 10 km. The magnetic low over the lower northern flank of the seamount and the high over the lower southern flank suggests that the core is mantled by reversely magnetized volcanics.

To estimate magnetic source depths and to obtain some idea of the geometry of the causative bodies, the automatic interpretation technique of Werner deconvolution (Jain, 1976; Hsu & Tilbury, 1977) was applied to the long E-W and S-N magnetic profiles across the summit of Capricorn. The sections of profile processed are indicated in Figure 3 as X-X' and Y-Y' and correspond to the seismic lines (Fig. 5). The results of the analysis are depicted in Fig. 11 (E-W line) and Fig. 12 (S-N line) for both thin sheet (dyke) model and interface (contact, fault) model. Corresponding magnetic and bathymetric profiles are also shown on the plots. The symbols represent the tops
of estimated magnetic sources and are proportional in size to the magnetization intensity. A clustering of depth estimates is generally indicative of a good solution - isolated estimates are not highly significant. The sprinkling of low intensity estimates plotted in the upper part of the water column is due to normal levels of non-geologic recording noise, plus 'mathematical' noise introduced in the computer processing.

The magnetic depth estimates broadly correlated with the surface and interior of Capricorn Seamount. No major magnetic sources are interpreted at or below seafloor in the area of the Tonga Trench axis and lower insular trench wall. This is not surprising since the magnetic profile over this area is smooth and relatively flat. Of particular interest is depth to magnetic basement beneath the summit area of the seamount. The tops of the more pronounced clusters of estimates lie at an average depth of about 1.3-1.4 km below sea-level. This coincides with the seismically determined base of the sedimentary capping. Other magnetic source clusters lie at greater depth beneath the summit area, but the shapes of the causative magnetic bodies are not readily deduceable from the plots. Such magnetic sources are symptomatic of the general magnetic inhomogeneity of the whole seamount.
References


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List of Figures

Figure 1. Location map and regional tectonic setting.
Bathymetric base map (500 m contours) is after Kroenke et al. (1983).

Figure 2. Seismicity of the Capricorn Seamount area. Epicentre locations have been plotted from data provided by the National Geophysical Data Center, Colorado. The arrows indicate the direction of Indo-Australian/Pacific plate convergence.

Figure 3. HMNZS 'Tui' geophysical survey lines across and in the vicinity of Capricorn Seamount. Small box symbols along the ship's tracks indicate 10-minute positions. Hourly positions are annotated with survey number (61), Julian day and GMT.

Figure 4. Bathymetry of the Capricorn Seamount/Tonga Trench area at 200 m contour interval. 'Tui' dredge sites are shown by large dots annotated with sampling station number.

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Figure 9. Gravity anomaly (at 25 mgal contour interval) for seamount density of 2.56 t/m$^3$. The seafloor topography is represented by polygonal bodies (A-I) with depth extends (km) as follows - A 0.6-0.8, B 0.8-1.0, C 1.0-1.4, D 1.4-2.0, E 2.0-3.0, F 3.0-4.0, G 4.0-5.0, H 5.0-6.0, I 6.0-7.0. The polygonal outlines are shown as dashed lines and the vertices as dots. A regional gravity correction of -86 mgal has been added.

Figure 10. Residual gravity at 10 mgal contour interval over Capricorn Seamount.

Figure 11. Werner deconvolution magnetic source estimates for the E-W magnetic profile across Capricorn Seamount and Tonga Trench (profile location is given by Y-Y' in Figure 3). The magnetic profile and bathymetry along the line are also shown.

Figure 12. Werner deconvolution magnetic source estimates for the S-N magnetic profile across Capricorn Seamount (profile location is given by X-X' in Figure 3). The magnetic profile and bathymetry along the line are also shown.
Location Map

Figure 1 (S5).

143
Seismicity

FIGURE 2 (S5).

144
FIGURE 3 (85)
Depths in Km

Bathymetry

FIGURE 4 (S5).
Figure 5 (S5).

E-W & S-N seismic profiles across Capricorn Seamount.
FIGURE 6 (S5).
FA Gravity Anomaly

FIGURE 7 (S5).
FIGURE 8 (S5).

50 nT contour interval

Magnetic Anomaly
Gravity anomaly for seamount density 2.56 t m$^{-3}$

FIGURE 9 (S5).
Residual Gravity

FIGURE 10 (S5).
FIGURE 11 (S5).
Werner Deconvolution

FIGURE 12 (S5)
Niue is an isolated, raised coral atoll in the southwest Pacific Ocean about 270 km east of the Tonga Trench (Fig. 1). The island is roughly oval-shaped with a N-S length of 22 km and E-W width of 17 km. It rises steeply from a surrounding ocean depth of about 4750 m. Major nearby submarine features include Endeavour Seamount 40 km to the ENE and Lachlan Seamount 30 km SE of Niue.

Much of the former atoll morphology has been preserved (Schofield, 1959). The relatively flat-lying central part of the island is about 35 m above sea-level and represents the floor of the original Mutalau Lagoon. This area rises outwards to what was the ancient atoll rim (Mutalau Reef) and now forms the highest part of the island - a peripheral ridge 60-70 m above sea-level. From the ridge the land-surface drops abruptly down to sea-level in a series of terraces, the most prominent of which is the Alofi Terrace at 23 m. A fringing coral reef about 100 m wide encircles the island at sea-level.

Radiometric dating puts the age of emergence of the atoll (Mutalau Reef) at a maximum of 700,000 years (Fieldes et al., 1960). Fauna from the raised lagoon have been dated as Plio-Pleistocene.
Schofield (1959). Schofield (1959) correlates the reef and terrace levels with well-established Pleistocene sea-levels in Britain. A further correlation is made by Schofield & Nelson (1978), who remark that the coastal terrace sequence on Niue is virtually identical to that on the islands of the southern Cook Group 1000 km or so to the east. If correct, such observations imply that the terraces on Niue were produced solely by eustatic sea-level standstills without any post-Mutalau Reef uplift of the island being involved. In contrast, Dubois et al. (1975) propose that the Pleistocene emergence of Niue is due to uplift, and base their argument on the hypothesis that oceanic lithosphere behaves as an elastic plate and flexes upwards prior to subduction forming an outer bathymetric rise seaward of deep-sea trenches (Watts & Talwani, 1974). The amplitude of the rise off the Tonga Trench is roughly 240 m, which is smaller than for a number of other trench systems for which estimates lie in the range 307-513 m (McAdoo & Martin, 1984). Calculations indicate that 64 km of movement towards the Tonga Trench produced the 70 m uplift of Niue.

The geology of Niue has been described by Schofield (1959), Schofield & Nelson (1978), Jacobson & Hill (1980) and Hill (1983). The soils of Niue are thin, radioactive and thought to have been derived largely from ash fall-out produced by eruption of regional volcanoes, particularly those of the Tongan arc to the west (Wright & Van Westerndorp, 1965). Geological mapping indicates that, apart from the veneer soil, the upper part of the island platform is composed entirely of carbonate lithologies.

These comprise coral-foraminiferal-algal calcarenite and reefal limestones, dolomitized to varying degree. Groundwater and mineral exploration drilling has proved carbonates to a depth of 300 m.
Palaeontological dating of drill-core samples from most of the carbonate sequence to a depth of 170 m below sea-level yielded a middle to late Miocene age (G.C.H. Chaproniere in Jacobson & Hill, 1980). Magnetic and gravity surveys conducted on Niue (Hill, 1983) suggest that the southwest of the island is underlain by a flat-topped, dome-shaped dense volcanic core at a depth of 300-400 m below sea-level. The core is interpreted as having a lateral density contrast of 0.20 t/m and a reverse magnetization of 3.0 A/m.

The 1986 'Tui' cruise appears to be the first to focus in the geology and geophysics of the Niue seamounts. The single-channel seismic, magnetic, gravity and 12 kHz bathymetry coverage achieved is indicated on the track map of Figure 2. The resultant extension of magnetic and gravity coverage offshore from Niue significantly improves control on the modelling of the islands' volcanic substructure. In addition to the geophysical survey, five sites were dredged to the south of Niue in water depths ranging from about 350 to 3600 m, with the objective of recovering volcanic rocks from the island pedestal and manganese crusts.

**Bathymetry Compilation**

The 1986 'Tui' survey work around Niue and nearby seamounts Lachlan and Endeavour, has contributed substantially to knowledge of the bathymetric details of the area. The only previous comprehensive bathymetric compilation dates back to 1966 (Brodie, 1966). Since then a significant amount of new data, including our own, has become available. This has allowed production of an updated bathymetric map with 500 m contour interval (Fig. 3).
The new compilation is based on

1. The 1986 'Tui' results
2. British Admiralty Chart 1174 - Niue and Suwarrow Island (Niue 1:150,000), 1982
with additional control provided by:

3. the Niue Island 1:200,000 chart (Brodie 1966)
4. 1:1,000,000 Ocean Sounding Charts 355 & 356 compiled by the Hydrographic Office of the Royal New Zealand Navy
5. digital data from research cruises available through the National Geophysical Data Centre, Boulder, Colorado. Cruises with bathymetric data coverage on the map area of Figure 3, included -:

(i) 'Vema' 1962 (Lamont-Doherty Geological Observatory)
(ii) 'Hakuho Maru' 1968 (University of Tokyo)
(iii) 'Melville' 1974 (Scripps Institute of Oceanography)
(iv) 'Conrad' 1974 (Lamont-Doherty Geological Observatory).

6. the 1982 USGS compilation for the Tonga region - Chase et al., 1982 (bathymetry to west of Niue).

It should be mentioned that Niue is positioned about 2.5 km too far north on the Niue Island 1:200,000 chart (Brodie, 1966). This discrepancy was discovered on our first approach to Niue using satellite navigation. BA chart 1174 appears to show Niue correctly located and was used subsequently as a guide for navigation close in to the island.
Another significant charting error was detected during the bathymetry compilation - it concerns the moderately large seamount shown immediately to the east of Lachlan Seamount. This feature is indicated on charts of Mammerickx et al. (1973) and Kroenke et al. (1983), as well as a number of other recent bathymetry maps.

It appears to have been charted on the basis of soundings made by 'Tui' in 1963. There is evidence to suggest that the soundings were mis-plotted - an error that has been perpetuated on sheets such as Ocean Sounding Chart 356. A critical sounding of 1534 m, for example, appears to be located about 14 km too far east. Re-positioned, it falls on Lachlan Seamount. It seems, therefore, that no seamount exists immediately to the east of Lachlan Seamount. This conclusion is supported by data from a nearby French hydrographic traverse made in 1976/77, indicating water depths of about 4700 m in the area.

Gravity and Magnetic Contour Maps

Gravity and magnetic maps of the Niue seamounts area have been prepared. These depict gravity anomaly at 25 mgal contour interval (Fig. 4), and magnetic anomaly at 100 nT contour interval (Fig. 5). The maps are based largely on our cruise data combined with an extensive land-based network of gravity and magnetic observations on Niue Island (Hill, 1983). Other research cruises which were sources of data include, - 'Vema' 1962 (gravity), 'Hakuho Maru' 1968 (gravity & magnetics), 'Melville' 1974 (magnetics) and 'Conrad' 1974 (gravity & magnetics). The Geophysics Division of N.Z. DSIR also supplied marine magnetic data acquired in 1963.

The gravity data are based on the Washington D.C. (Commerce
Floor) datum of 980118.7 mgal (Woollard & Rose, 1960; Robertson, 1965), with free-air anomalies calculated using the 1930 International Gravity Formula. The gravity anomaly map has been compiled from free-air anomalies at sea and Bouguer anomalies on land (density 2.1 t/m³ for above sea-level topography). The magnetic anomaly values are relative to the global IGRF80 reference field (Peddie, 1982).

**Dredging Results**

Five dredge hauls (U354-U358) were made on the southern flank of Niue in efforts to sample volcanic core rocks and manganese crusts. The dredge locations are indicated on the bathymetry map (Fig. 3) and also annotated on the seismic sections (Fig. 6) to give an idea of their relative settings on the submarine slope off Tepa Point.

A summary of the dredging results is provided below.

<table>
<thead>
<tr>
<th>Station</th>
<th>Depth range (m)</th>
<th>Material recovered</th>
</tr>
</thead>
<tbody>
<tr>
<td>U354</td>
<td>346-467</td>
<td>6 fragments coral (16-29 mm diam.)</td>
</tr>
<tr>
<td>U355</td>
<td>586-487</td>
<td>Dredge empty.</td>
</tr>
<tr>
<td>U356</td>
<td>1873-1100</td>
<td>250 carbonate fragments (10-150 mm diam.) - many smaller pieces. Thick Fe, Mn oxide staining on some samples.</td>
</tr>
</tbody>
</table>
U357  3612-3596  14 lumps (18-45 mm diam.) altered tuff, most with Mn-crust 2-3 mm thick.  1 piece (27 mm diam) block vesicular basalt.  1 fragment coral (27 mm diam.).  1 pumice pebble.

U358  2520-2450  3 pieces (8-34 mm diam.) coralline limestone rubble with Mn-crust to 3 mm.

The paucity of volcanics and the general small quantity of material recovered in the rock dredge hauls was disappointing, especially in light of the concerted efforts made to achieve optimum recovery off Niue. It can only be concluded that the rock dredging equipment was just too light for the job. The submarine slopes off Tepa Point are quite steep (30-35°) and believed to be surfaced by high strength material such as relatively fresh volcanic core rock (?basalt) and carbonates. The rock dredge/winch combination appears to have had inadequate ripping capability to deal with such surfaces and consequently no in-situ volcanics were recovered off Niue.

The lithologies of altered tuff and vesicular basalt recovered are of particular significance to our investigation. Nevertheless, the limited quantities of these volcanic rocks recovered makes assessment of how representative they are of the Niue edifice rather speculative.

Bathymetry and Seismic Results

The geophysical lines completed are as shown in Figure 2. The ship sailed a W-E line north of Niue, turned south to pass over Endeavour
Seamount, then turned west to cross Lachlan Seamount and continued on to the area south of Niue to complete a zig-zag pattern over the southern flank of the seamount before commencing dredging operations. Seismic data were not recorded on the line out of the area towards Capricorn Seamount.

Time constraints meant that only one pass over Endeavour Seamount and Lachlan Seamount was possible. These features had not been accurately charted in the past and so our lines may not have passed directly over their highest points, though we believe they did pass over the summit areas. Shallowest depth recorded for Endeavour Seamount was 1625 m, while for Lachlan Seamount it was 920 m. Soundings plotted on Ocean Sounding Sheet 356 show slightly shallower minimum depths for these features - 1308 m and 646 m, respectively.

The seamounts rise from a surrounding ocean depth of about 4750 m - gently at first then much more steeply towards their summits. Slopes are in the order of 5° at the 3500 m isobath, but then steepen upwards to an average of 15-20°. Around Niue the steepest submarine slopes (approximately 30-35°) lie between the 2000 m isobath and the coastline.

The south-west flank of Niue is particularly steep. The isobaths show a pronounced general NW-SE trend in this area, broken only locally SW of Tepa Point by an isolated bathymetric high, and suggest the existence of a major structural zone intersecting the SW flank of Niue. In deeper water farther to the SW, the 3500 and 4000 m isobaths extend outward in a lobate pattern to outline a large mound or fan at the base of the seamount.
Though no extensive seismic coverage of the lower slopes of the seamounts was obtained, there is evidence, especially between north of Niue and Endeavour Seamount (see Figure 6), that these are underlain by several hundred metres of chaotically bedded sediment over an irregular and faulted basement. The sediment probably consists of talus and slump deposits derived from higher on the seamount flanks. The mixture of lithologies recovered at station U357 tends to support this.

Though the mid and upper slopes of the seamounts are generally too steep for seismic mapping of internal structure, a fairly strong reflector appears to be present beneath the upper SE flank of Niue (Fig. 6) at about 1.3 sec twt. A significant break in submarine slope or terrace is present where this reflector outcrops. The reflector may mark the boundary between an overlying ash/tuff sequence and underlying basalt flows (volcanic core). This is further discussed in the section on gravity/magnetic modelling.

Sections of the 12 kHz bathymetric records have been reproduced (Fig. 7) to display details of the summit topography of Endeavour and Lachlan Seamounts.

The summit area of Endeavour Seamount is of high relief, consists of a central peak flanked by numerous steep-sided local topographic highs. The jagged appearance of the summit suggests a high degree of erosional dissection, though it is possible that the individual small peaks may represent parasitic cones. There is no evidence of a sediment capping. Lachlan Seamount also has jagged summit topography, though no central peak, and the suggestion of an approximately planar surface at about 1055 m water depth. This surface may represent a
summit crater, with the 150 m high peak on the eastern side being a parasitic cone or elevated part of the crater rim. As for Endeavour Seamount, no sedimentary section is obvious in the seismic data, though some record of the summit area was lost during repair of a leaking air-gun.

Bathymetric profiles presented by Summerhayes (1967) include crossings of Endeavour and Lachlan Seamounts. These profiles show summit topographies similar to those seen in our records, and is confirmation that we did, in fact, pass over the summit areas of the seamounts so that our images of the seamounts are not expected to be unduely distorted by the effects of side-echoes.

Gravity Field over the Niue Seamounts

By comparing the bathymetry and gravity contour maps (Figs. 3 and 4, respectively) it is obvious that the gravity anomalies closely reflect the seafloor topography.

An along-track high of 115 mgal was recorded over the summit area of Endeavour Seamount, while a maximum value of 170 mgal was recorded over Lachlan Seamount. The gravity high of 247 mgal on Niue is eccentrically located on the island, and lies to the south-west over an interpreted dense volcanic core (Hill, 1983). Further evidence that the centre of mass of the Niue edifice lies toward the south-west of the island rather than directly beneath its centre, can be seen in the 50, 75 and 100 mgal contours which display a distinct outward bulge to the southwest.
Magnetic Field over the Niue Seamounts

Referring to Figure 5, it can be seen that large variations in total field anomaly occur over the Niue seamount area. The IGRF80 magnetic anomaly has a range of about -600 to +500 nT.

An extensive and deep (-600 nT) negative anomaly is located over the central north of Niue. The field increases rapidly to the south, over the southern sector of Niue, and then levels out onto a +400 nT high over the ocean about 12 km south of the island. The field over Niue has been interpreted as a reversely magnetized core located beneath the SW of the island (Hill, 1983).

In the case of Endeavour Seamount, an intense high of +500 nT is located just to the NW of the summit. This high may be related to the large high (greater than +500 nT) recorded on the 'Melville' line about 15 km to the north. The field appears to be fairly flat over the summit area of the seamount and decreases to a -400 nT low 10 km to the south. Lachlan Seamount has a +100 nT high over the northern part of its summit area. The field decreases northward to a deep -600 nT low situated about halfway between Lachlan and Endeavour Seamounts.

Large data gaps still exist around Endeavour Seamount - particularly to the east, and around Lachlan Seamount - particularly to the south. Until these seamounts have their magnetic fields surveyed in more detail, attempts to make reasonably accurate estimates of their magnetization parameters would be premature. A rough qualitative assessment can be made, however. Assuming that the seamounts possess a fairly uniform, mainly thermoremanent...
magnetization and that no rotation has taken place, it appears that 
Endeavour Seamount may be of normal polarity, while Lachlan Seamount 
is probably reversely magnetized - as is Niue's core.

The large magnetic high to the south of Niue and the large high 
recorded on the 'Melville' line to the north of Endeavour Seamount 
cannot be attributed directly to magnetization of the seamounts, and 
probably originate from magnetic heterogeneities within the underlying 
oceanic crust. Polarity reversals in the crust acquired during 
accretion may be the source of the anomalies, though no magnetic 
lineations have been mapped in the region (AAPG, 1981) and it has been 
assumed that the oceanic crust in the area was formed during the long 
interval of normal polarity in the Cretaceous.

Gravity and Magnetic Modelling

The gravity and magnetic fields over Niue and adjacent offshore 
area have now been mapped in sufficient detail to allow reasonably 
precise calculation of density and magnetization parameters, and to 
permit an attempt at modelling the internal structure of the edifice. 
Adequate gravity data are also available to enable rough calculations 
of Endeavour Seamount and Lachlan Seamount densities, though their 
magnetic fields are not known well enough to allow confident estimates 
of their magnetic parameters.

In this study the gravity and magnetic fields of the Niue 
seamounts were analysed by the 3-dimensional modelling methods of 
Plouff (1975a, 1975b, 1976). The techniques permit calculation of the 
gravity and magnetic fields of geological bodies with complex shape by 
simulating them by an assemblage of polygonal prisms. Seamounts are
readily modelled by approximating their submarine topography by a stack of polygonal prisms with vertical sides which follow the bathymetric contours. The measured anomaly data can be inverted to yield least-squares estimates of density and magnetization parameters, plus a planar regional gravity/magnetic field.

The submarine edifice of Niue was approximated by a set of nine polygonal slabs (A-I) as defined in Figure 8. The vertices of the polygonal bodies were selected so that the polygonal outlines closely matched respective bathymetric contours (Fig. 3). The composite body represents a depth extent of 0-4.5 km below sea-level. Observed gravity anomaly values were scaled over a 4 km x 4 km grid within the boxed area centred on Niue shown in Figure 8. Least squares comparison between the 100 observed and calculated values gave a best fit density of 2.32 +/- 0.02 t/m³ for the edifice, assuming a density of 1.03 t/m³ for sea-water. The calculated regional gravity anomaly datum was +28 mgal, and the standard deviation between observed and calculated values was 8 mgal. The theoretical gravity contours are shown in Figure 8, and as can be seen, closely match the observed data (Fig. 4). An important difference is obvious on Niue Island, however, with the observed high appreciably displaced to the SW. The gravity measurements made on the island are significantly more accurate (better than +/- 0.5 mgal) than those from adjacent offshore areas. In addition, the gravity field flattens out over the island so that any underlying anomalous mass distributions become more apparent in the data and are not disguised by high gradients associated with steep or rugged terrain of the island slopes.

Apparent mean densities were calculated for the edifices of Endeavour Seamount and Lachlan Seamount in a fashion similar to that
for Niue, except that representative observed gravity values were selected from or close to ship's tracks rather than from a regular grid because of the non-uniform data coverage. Endeavour Seamount was approximated by 6 polygonal bodies with depth extents 1.5-2.0, 2.0-2.5, 2.5-3.0, 3.0-3.5, 3.5-4.0 and 4.0-4.3 km. Lachlan Seamount was modelled as 7 bodies with depth extents 0.75-1.0, 1.0-1.5, 1.5-2.0, 2.0-2.5, 2.5-3.0, 3.0-3.5 and 3.5-4.3 km. The results were -

<table>
<thead>
<tr>
<th></th>
<th>Best-fit density*</th>
<th>Standard deviation</th>
<th>Regional gravity (Observed-calculated gravity)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Endeavour Seamount</td>
<td>2.54±0.11 t/m³</td>
<td>4 mgal</td>
<td>22 mgal</td>
</tr>
<tr>
<td>Lachlan Seamount</td>
<td>2.76±0.05 t/m³</td>
<td>5 mgal</td>
<td>3 mgal</td>
</tr>
</tbody>
</table>

*assuming sea-water density = 1.03 t/m³.

For magnetic modelling tests of the Niue edifice a similar spatial grid of observed data to that used for the gravity modelling was adopted. One modification, however, was the discarding of the southern-most row of data points to prevent excessive interference on solutions by the extensive magnetic high to the south of Niue which does not appear to be directly related to magnetization of the seamount. This left a grid of 90 data points at 4 km spacing within the boxed area shown in Figure 9.

In magnetic modelling of the Niue edifice, the first approach was to assume uniform magnetization and to try an experimental least-squares best fit to the observed data. The same polygonal
representation of the submarine topography as in the gravity modelling was used. Not unexpectedly, the best solution for this model yielded a poor fit to the data. Further attempts to find an acceptable solution were made by removing the upper-most polygon from successive models, in order to simulate an increasing thickness of non-magnetic top section to the edifice. Only marginal improvement was achieved, however. Obviously, the edifice is not uniformly magnetized, even allowing for a non-magnetic capping.

The next approach was to attempt to model the field by a magnetic core within the edifice, as suggested by the two-dimensional modelling of Hill (1983). Following some trial-and-error adjustments of core position and shape, a good fit to the observed data was achieved. The optimum model is shown in Figure 9 together with the calculated field based on best-fit intensity of total magnetization 2.93 A/m, declination 206° and inclination 35.5°. The multiple correlation coefficient for this fit is 0.89, and the standard deviation for observed-calculated data is 90 nT with a regional magnetic datum of -70 nT.

The modelled core is reversely magnetized and centred on Tepa Point in the SW of Niue. It is dome-shaped and flat-topped at a depth of 500 m. The dome-like symmetry of the core is flawed to the SW of Niue where it appears that a large section of its flank has been removed.

In order to highlight anomalous density distributions directly beneath the island, residual gravity over Niue was calculated and a contour map produced (Fig. 10). The residual gravity values were calculated by subtracting theoretical gravity for the best-fit island...
pedestal density of 2.32 t/m³ from the observed gravity. The residual gravity contours show a relative high of 25 mgals located over the SW of Niue - above the top of the magnetic core, which implies a relatively high density for the core.

A rough estimate of the core's density contrast can be obtained by least-squares fitting of calculated anomalies produced by the core model to corresponding residual gravity values which are based on observed data. In doing this one obtains an answer of +0.25 t/m³ for density contrast of the core. Individual core and upper edifice densities can then be readily determined by modelling of the combined structure with a few trial sets of density values. This yields a core density of 2.41 t/m³ and an upper edifice density of 2.16 t/m³.

Dense cores appear to be a common feature of Pacific atolls and volcanic islands (Robertson, 1967a, 1967b, 1970, 1987). Some seamounts have been modelled as having a core that is both dense and also highly magnetized (eg. Woodward, 1970; Davey, 1973).

Comments on Modelling Results

In the preceeding gravity analyses it has been assumed that the Niue seamounts are located on essentially planar lithosphere, implying that the seamount loads are supported by regional, rather than local isostatic compensation. This assumption appears to be valid since there is no evidence of local lithospheric depression as might be expressed by bathymetric or gravitational moating around the
seamounts. Neither are the calculated densities excessively low, as would be the case for local compensation. The calculated values are typical for common seafloor lithologies, and are in line with the range of 2.3-2.6 $t/m^3$ most frequently reported in the literature for estimates of mean seamount density. The volcanic edifices must, therefore, have been constructed well off the mid-ocean spreading axis on cold lithosphere of high flexural rigidity (Watts & Ribe, 1984).

The least-squares calculation of seamount densities involved computation of gravity base levels as a by-product of the analysis. From the gravity anomaly map (Fig. 4) it appears that the mean regional gravity anomaly is about $+20$ mgal, which approximately corresponds to the mean of base level calculated for Niue ($+28$ mgal), Endeavour Seamount ($+22$ mgal) and Lachlan Seamount ($+3$ mgal). The variations in datum are not appreciable and may well reflect factors that should genuinely contribute to the respective base level. Such factors include choice of depth extent of the polygonal model, anomalous masses in the underlying crust or a regional gravity gradient. It is possible, on the other hand, that these variations are partly the result of severe perturbations in the observed data not accounted for in the modelling assumptions - the presence of gross lateral density variations in seamount or crust, for example. The possible error in calculated densities is expected to be less than 0.10 $t/m^3$ (for datum error of 10 mgal or slightly more, depending on the amplitude of the seamount gravity anomaly).

Inclination ($i$) of the earth's magnetic geocentric dipole field at latitude $L$ is given by $\tan i = 2 \tan L$. The magnetization vector calculated for the core of Niue is inclined at $35.5^\circ$ which corresponds to a palaeolatitude of $19.6^\circ$. This compares with Niue's present
latitude of 19.1°S, and is consistent with a northly drift of Niue on the Pacific plate since the core cooled below the Curie point.

Uncertainty in the magnetization parameters, largely due to inadequate control on the core geometry, prevent confident determination of virtual geomagnetic pole and hence palaeomagnetic age for Niue.

Evolution of the Niue Seamounts

The Niue seamounts are believed to have evolved through a progression of stages (A-D) as depicted in Figure II, and as described below.

Stage A

In early-middle Miocene pillow basalts erupted onto seafloor of mid-late Cretaceous age (70-100 m.y. old at the time). The volcanism may have been initiated at a point of crustal weakness as the Pacific plate drifted over a mantle hot-spot. This conjectured hot-spot may now be located just east of Rarotonga, having created the Capricorn-Niue-Rarotonga chain of seamounts.

Continued volcanic activity produced a core of pillow-lavas and interbedded tuffs about 4 km high, with the summit reaching close to or slightly above sea-level. This was the proto-core of Niue with density of 2.41 t/m³ and high thermoremanent magnetization (2.93 A/m), imparted at a time of polarity reversal in the earth's magnetic field.

Construction of the Lachlan Seamount core may have been
contemporaneous with that of Niue because of common location on a NW
trending structural weakness in the crust and because they both appear
to possess reversed thermoremanent magnetization. The high density of
Lachlan Seamount (2.76 t/m³) suggests that the feature was built
from dense basaltic flows which probably coalesced with those of Niue
to produce a low saddle between the two edifices. Significant
invasion of the core by dykes and other intrusions may be a factor
contributing to the high density of Lachlan Seamount.

Being located east of Niue (in the younging direction of the hot
spot trace) and, apparently possessing a normal magnetic polarization,
Endeavour Seamount may have evolved a little after the other two
seamounts. A moderate density (2.54 t/m³) suggests that the core
was built mainly of pillow basalts.

Stage B

With the summit of Niue near sea-level, a change in style of
volcanism took place, with basalt flow giving way to explosive
eruptions of ash and hyaloclastite. The free expansion of magmatic
gases in the shallow marine or sub-aerial conditions, without the
confining hydrostatic pressures of the earlier deep-sea phase of
volcanism may have contributed to the change, as might possible
magmatic differentiation. Ash-falls and other pyroclastic deposits
built up over the basalt cone to form a mantle almost 3 km thick.
Consisting of loosely packed, essentially randomly oriented particles,
these volcanic sediments would have low bulk magnetization and also
low density (2.16 t/m³). The altered tuff and vesicular basalt
dredged at station U357 are probably part of this unit.
A massive explosive eruption and/or caldera collapse marked the termination of volcanic activity. This cataclysmic event resulted in the removal of a large section of the SW flank of the volcano. If an explosive event were responsible, the section may have either blown out, or became detached and moved downslope as a great submarine landslide; with the resultant debris possibly now forming the seafloor apron and mounded area to the SW of Niue.

Stage C

With the volcano extinct, sub-aerial erosion and wave action planated the upper surface to sea-level. A carbonate reef began to develop in the shallow-water tropical environment. The reef continued to grow upwards, forming an atoll as the volcanic base subsided. Evolution of atolls and subsidence mechanisms have been discussed by a number of workers (Menard, 1973; Detrick & Crough, 1978; Scott & Rotondo, 1983). Niue is considered to have developed as a typical atoll, with subsidence produced by (i) depression of the lithosphere in isostatic response to the volcanic load, and (ii) cooling and sinking of the surrounding lithosphere as the mantle hot spot was over-ridden. The coral limestone capping attained a maximum thickness of 500-600 m.

Loading by the limestone capping, combined with diagenesis, compaction and deformation of the volcanic sediments produced gravitational instability in the flanks of the edifice expressed episodically as major submarine landslides. The embayed coastal outline of Niue is believed to be due to such mass movements. Large limestone chasms oriented sub-parallel with coastline are common along Niue's coastal strip. Interpreted as solution channels developed
along tensional gashes, they are further evidence of slump activity (Schofield, 1959). Concave 'bights' in coastal outlines, as seen on Nuie, are a common feature of Pacific atolls and represent a mature stage in development of atoll morphology (Fairbridge, 1950).

Stage D

Uplift of the atoll commenced as early as 700,000 years ago (Fieldes et al., 1960), as Niue, conveyed on the Pacific plate, began to ascend the outer rise seaward of the Tonga Trench (Dubois et al., 1975). The raised atoll now stands 70 m above sea-level.

References


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**Figure 1.** Location map and regional tectonic setting. Bathymetric base map (500 m contours) is after Kroenke et al. (1983).

**Figure 2.** HMNZS 'Tui' geophysical survey lines in the Niue area. Small box symbols along the ship's tracks indicate 10-minute positions. Positions are annotated with survey number (61), Julian day and GMT.

**Figure 3.** Bathymetry of the Niue seamounts area at 500 m contour interval. Tui dredge sites (small circles) are annotated with sampling station number.
Figure 4. Gravity anomaly over the Niue seamounts area. The contours represent free-air anomaly over the ocean and Bouguer anomaly on land (Niue Island). Contours shown as solid lines are at 25 mgal contour interval, while broken contours (over Niue) are at 5 mgal interval.

Figure 5. Total magnetic intensity anomaly at 100 nT contour interval over the Niue seamounts area. The magnetic anomaly field is based on the IGRF80 reference field.

Figure 6. Seismic profiles over the Niue seamounts. The profile locations are given by A-I which denote course change positions (see Figure 2). The locations of dredge sites U354-358 on the southern flank of Niue are indicated.

Figure 7. 12 kHz bathymetric records showing morphological detail of the Endeavour and Lachlan Seamount summit areas.

Figure 8. Gravity anomaly (at 25 mgal contour interval) for Niue seamount density of 2.32 t/m$^3$. The seafloor topography is represented by polygonal bodies (A-I) with depth extents (km) as follows - A 0-0.25, B 0.25-0.5, C 0.5-1.0, D 1.0-1.5, E 1.5-2.0, F 2.0-2.5, G 2.5-3.0, H 3.0-4.0, I 4.0-4.5. The polygonal outlines are shown as dashed lines and the vertices as dots. A regional gravity correction of +28 mgal has been added.

Figure 9. Magnetic anomaly (at 100 nT contour interval) for the
volcanic core model. The core is represented by polygonal bodies with depth extents (km): J 0.5-0.7, K 0.7-1.0, L 1.0-2.0, M 2.0-3.0, N 3.0-4.0, P 4.0-5.0. The outlines (in plan) of the bodies are shown as broken lines, while their approximate polygonal representation is given by the dots corresponding to the polygon vertices. Magnetization parameters are: intensity of magnetization 2.93 A/m, declination 206°, inclination 35.5°. A regional magnetic correction of -70 nT has been added.

Figure 10. Residual gravity over Niue at 5 mgal contour interval.

Figure 11. Conceptual evolution of Niue, stages A-D.
FIGURE 3 (S6).
GRAVITY ANOMALY
NIUE SEAMOUNTS

Contours in mgal

FIGURE 4 (S6).
FIGURE 6 (S6).
FIGURE 7 (56).

Endeavour Sm

Depth (m)

1600

1650

1700

1750

1800

1850

1900

1950

2000

Lachlan Sm

Depth (m)

900

925

950

975

1000

1025

1050

1075

1100

1125

1150

Ships speed reduced from 7 to 4 knots.

Summit detail - 12 kHz bathymetry, Endeavour & Lachlan Seamounts
Gravity anomaly for seamount density 2.32 t m⁻³

FIGURE 8 (S6).
Magnetic anomaly for volcanic core model

FIGURE 9 (S6).
Residual Gravity

FIGURE 10 (S6).
Conceptual evolution of Niue

FIGURE 11 (S6).
SECTION 7

STRUCTURE AND EVOLUTION OF NAURU ISLAND.

CENTRAL PACIFIC OCEAN

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ABSTRACT

Nauru Island, in the central Pacific Ocean, is a raised atoll capping a volcanic seamount rising from an ocean floor depth of 4300 m. The land area is 22 km$^2$, and the island rises to 70 m above sea level.

Drilling has proved dolomitised limestone of upper Miocene or younger age to a depth of 55 m below sea level. Gravity and magnetic surveys show that the substructure of Nauru is approximately radially symmetrical. Bouguer anomaly increases by 18 mgal from the coast to the centre of the island. The magnetic field has a range of 830 nT with a negative anomaly located over the NNW coastline and a large E-W elongated positive anomaly over the southern sector of Nauru.

Modelling of the magnetic field suggests that the island is underlain by a reversely magnetized volcanic core with magnetization 1.5 - 1.9 A m$^{-1}$ and depth 500 m. The orientation of the magnetization vector gives the palaeomagnetic age of the core as mid-Eocene to Oligocene.

This age range for the evolution of Nauru is confirmed by an estimate based on island subsidence. Nauru may have been constructed by volcanism associated with the Samoa hot spot. Gravity modelling indicates a density of 2.5 t m$^{-3}$ for the island pedestal. The calculated values for density and magnetization are typical for Pacific seamounts, implying that basement beneath Nauru's carbonate platform is composed mainly of basaltic lavas.

Key words: Nauru, raised atoll, phosphate, central Pacific, gravity survey, magnetic survey, palaeomagnetism, volcanic core.
INTRODUCTION

Nauru Island occupies a land area of 22 km$^2$ in the central Pacific Ocean at 0° 32'S, 166° 56'E (Fig. 1). The island has been mined for its surficial phosphate deposits for about 80 years, and the current reserves indicate that the mine has approximately 9 years' life left. About 80 percent of the land area has had its vegetation and phosphate soil cover removed leaving an exposed limestone pinnacle surface, or karrenfeld.

Rehabilitation of the mined out land is under consideration and this is expected to increase the demand for water at a time when imported water will no longer be available. In this context an investigation of the hydrogeology and groundwater resources of Nauru was undertaken by the Bureau of Mineral Resources, Geology and Geophysics (BMR) in 1987 on behalf of the Commission of Inquiry into Rehabilitation of the Worked-out Phosphate Lands in Nauru. The investigation included gravity and magnetic surveys to determine the depth and shape of the island's basement; and geoelectrical soundings and drilling to determine the thickness of the freshwater layer and subsurface geology. Details of the investigation and conclusions with respect to groundwater resources have been documented (Jacobson and Hill, 1988). We report here the results of the investigation concerning the island's structure and evolution.

BATHYMETRY

Nauru is a raised coral atoll which apparently is underlain by a volcanic seamount that rises more than 4000 m from the floor of the Pacific Ocean (Fig. 2).
The only near-shore bathymetric survey for which data are available was completed in 1980. This survey mapped out to the 540 m isobath along a 1.7 km stretch of coast at the ship moorings on the western side of the island. Beyond the fringing reef, the submarine slope of the island was shown to descend steeply at about $34^\circ$. No other bathymetric information appears to be available within the 3000 m isobath surrounding Nauru. Beyond the 3000 m isobath, the data sources comprise: - Oceanic Soundings Sheet 261 (scale 1:1 000 000) compiled by the Hydrographic Office, R.A.N., Sydney (1977); and research cruise data. Relevant cruises include:

(a) 1971 R/V Dimitri Mendeleyev (Cruise 6) - USSR Academy of Sciences, Institute of Earth Physics;

(b) 1972 R/V Vityaz (Cruise 51) - USSR Academy of Sciences, Institute of Earth Physics; and

(c) 1977 R/V Vema (Cruise 34) - Lamont Doherty Geological Observatory.

The bathymetric contour map (Fig. 2) of Nauru was compiled from these data sources, and the distribution of data over the map area is indicated. The submarine slopes descend steeply to water depths of about 3000 m then level off to the surrounding abyssal plain at a depth of about 4300 m.
ISLAND GEOLOGY

The maximum elevation of the island is about 70 m. Even prior to mining, the original atoll topography had been modified by karstic erosion. An early map reproduced by Hutchison (1950) shows an irregular island rim, 30-60 m above sea level surrounding four interior depressions, three of which had a base level of about 20 m. The fourth depression contains Buada Lagoon, in the southwest of the island, which is just above sea level.

A coastal terrace extends around the island: the terrace is up to 400 m wide (Fig. 3) and is a few metres above sea level. Between the inland plateau and the coastal terrace is a narrow chain of depressions including brackish lagoons at Anabar in the northeast of the island. Peripheral to the coastal terrace is a fringing reef which extends 200 m offshore with an outer slope dipping 34° into deep water.

Radiocarbon dates have been obtained from drillcore Q1 in the coastal terrace (Fig. 3). The dates, for aragonitic coral, were 2730 +/- 60 years for a sample at 1-2 m, and 2820 +/- 60 years for a sample at 2-3 m (Table 1). The drillhole was sited 3.5 m above mean sea level on the storm ridge separating the Anabar lagoons from the sea, and the samples probably represent cemented storm deposits rather than coral reef in situ. The dates indicate that the coastal terrace is a youthful feature, and its construction is probably related to a high stage of Holocene sea level.

The phosphate capping of Nauru is, or was, several metres thick, and overlies an intensely dissected karrenfeld with karst limestone.
pinnacles up to 14 m high. Some isolated pinnacles are higher. The phosphate deposits also occupy the space between the pinnacles, and infill caves and joints in the limestone. A description of the phosphate deposits was given by Power (1910). The formation of the phosphate deposits from avian guano is believed to date back more than 300,000 years (Roe & Burnett, 1985).

Dolomitised limestone forms the bulk of the island and was drilled in this investigation to a maximum depth of 55 m below sea level. The limestone drillcores contain foraminifera which are late Miocene or younger in age (G.C.H. Chaproniere, pers. comm.). Quaternary molluscs were identified near the island's surface by Ludbrook (1964). The limestone is intensely karstified, with phosphatic cave-fillings extending well below sea level suggesting that Nauru has been above sea level for much of its history. The last emergence that contributed to karstification was probably the glacial low, 15,000 years ago, when sea level was about 100 m lower than at present (Chappell, 1983).

The arcuate eastern coastline probably represents the headward scarp of a submarine slump. Coastline concavities produced by landslides are a common feature of older atolls in the Pacific (Fairbridge, 1950). Arcuate fractures subparallel to the coastline have been observed on air photos of the southwest and northwest of the island (P.J. Barrett, pers. comm.). There is also a series of generally NW-SE linear fractures across the island, some of which have vertical displacement of up to several metres, visible in the exposed karrenfeld surface.
TECTONIC SETTING

Nauru is located at the southern end of the Nauru Basin, a 4000-5000 m deep ocean basin between the Marshall/Gilbert island chains to the east and the Ontong-Java Plateau to the west (Fig. 1). The Nauru Basin contains a lineated sequence of Jurassic-Early Cretaceous magnetic anomalies (Larson, 1976; Cande et al., 1978). The magnetic anomaly pattern trends east-northeast across the basin, with age of the lineations increasing northward (Fig. 1). Anomaly M16 has been reliably identified just north of Nauru. Identifications of M11 to M15 in the area of Nauru and Ocean Islands are more tenuous, being based on correlations with only one marine magnetic profile. Nevertheless, it is probable that Nauru overlies oceanic lithosphere accreted at a mid-ocean spreading ridge at about M15 time (132 Ma).

DSDP drilling at Site 462 (Larson, Schlanger et al., 1981) in the northern Nauru Basin was expected to bottom in Late Jurassic oceanic basement on the basis of observed magnetic lineations. Instead, over 500 m of mid-Cretaceous basalt sills, flows and some intercalated volcaniclastic sediments were penetrated without recovery of Jurassic material (Larson & Schlanger, 1981a). The apparent magnetic transparency of the Late Cretaceous volcanic complex is attributed (Larson & Schlanger, 1981b) to the sheet-like geometry, normal magnetization and fairly consistent magnetization direction of individual units. Such a magnetized sheet would produce anomalies only at its edges. Houtz & Ludwig (1979) have mapped an acoustically reverberant subbottom layer (believed to represent the volcanic complex) over most of the Nauru Basin. The layer is shown to be more than 600 m thick at DSDP Site 462 and also in the vicinity of Nauru.
At DSDP Site 289 on the Ontong Java Plateau (Fig. 1), basalt of pre-early Aptian (about 108 Ma) age was encountered beneath a cover of 1262 m of calcareous biogenic sediments (Andrews, Packham et al., 1975). The basalt appears to be typical oceanic tholeiite and probably represents true oceanic basement. The age disparity with that indicated by the magnetic lineations in the Nauru Basin to the east can be accounted for by a transform fault coinciding with the eastern margin of Ontong Java Plateau (Winterer et al, 1974). A reliable \(^{40}\text{Ar}/^{39}\text{Ar}\) age determination of 110 +/- 3 Ma on a sample from the upper sill unit drilled at Site 462 suggests that the Cretaceous volcanic complex of the Nauru Basin, and basaltic basement of the Ontong Java Plateau, are coeval. This extensive volcanic and seafloor spreading episode was the forerunner to even more widespread mid-plate volcanic activity during the rest of the Cretaceous, which involved most of the central Pacific including the Central Basin, Mid-Pacific Mountains and Line Islands as well as the Nauru Basin - Marshall Islands area. Regional thermally induced uplift associated with the volcanism began in Barremian-Aptian time (115-110 Ma) and continued until the Maestrichtian, about 70 Ma (Schlanger & Premoli Silva, 1981). Normal subsidence followed. Evolution of a broad oceanic swell in the late Mesozoic was first postulated by Menard (1964), who called it the Darwin Rise.

The Nauru area appears to have been tectonically stable except for minor adjustments throughout the Cenozoic, despite complex interactions at the Indo-Australian/Pacific plate boundary to the southwest. Subduction commenced at the North Solomon trench in late Eocene - early Oligocene. Ontong Java Plateau entered the trench in early Miocene (about 20-22 Ma) jamming the subduction process, with the result that convergence was taken up by reversal of subduction at
the South Solomon trench in the late Miocene (Kroenke, 1984). The Pacific plate at Nauru is presently moving northwest at about 104 mm/year (Minster & Jordon, 1978).

GRAVITY SURVEY

For the gravity survey of Nauru, a total of 228 gravity stations were occupied and the network was tied to three gravity base stations located at; (i) entrance to Meneng Hotel; (ii) benchmark at the Nauru Phosphate Commission Office; and (iii) entrance to the Survey Office in the Central Workshop area. For precise elevation control, gravity stations were sited at existing benchmarks wherever possible. Thirteen benchmarks, distributed mainly along the coastal ring road, were utilized. Elsewhere, gravity stations were located with the aid of detailed 1:1000 topographic map sheets. These sheets, produced in 1982 from aerial photography and geodetic ground surveys, provide 1 m contours and allowed elevations to be scaled to 0.1 m precision.

The gravity meter (Sharpe S145) used for this survey was not a geodetic instrument and because of its limited range could not be used to tie into the Australian network. Since it appears that no absolute gravity stations have been established on Nauru, the survey data are relative only. The gravity data were corrected for drift and reduced to relative Bouguer anomaly by applying elevation and latitude corrections and adopting a density of 2.1 t m\(^{-3}\) for above sea-level topography. A small number of stations located on or near steep slopes required terrain corrections.

The gravity map of Nauru (Fig. 4) shows a dome-shaped field with
contours concentric about the centre of the island. The relative gravity field ranges from a high of about 22 mgal at the centre to 4 mgal at the southwest coast. The pronounced gravity high over the centre of the island reflects the excess mass of the Nauru pedestal (density about 2.5 t m$^{-3}$) over that of the surrounding ocean (density 1.03 t m$^{-3}$). The central position of the gravity high and the almost circular, little-distorted nature of the contours indicate that the substructure of the island is approximately radially symmetrical.

**GRAVITY MODEL**

Three-dimensional gravity modelling of the submarine pedestal of Nauru was undertaken to obtain an estimate of mean density and to reduce the observed relative Bouguer anomalies to residual gravity values. Residual gravity is here defined as relative Bouguer anomaly minus the gravitational contribution of the island pedestal, and effectively represents gravity (with arbitrary base) corrected for submarine and subaerial terrain. The residual gravity field should highlight any anomalous density distributions within the edifice. The gravity modelling technique of Plouff (1976) was used for the analysis. By this method, the gravity field of geological bodies with complex shape can be calculated by representing them as an equivalent set of polygonal prisms. Least-squares inversion of the observed data allowed best-fit density to be estimated.

For the modelling, the submarine pedestal of Nauru was approximated by a stack of 10 polygonal slabs, each with 9-19 vertical sides and horizontal upper and lower surfaces. The slabs represent horizontal layers of the seamount as defined by the bathymetric contours of
Figure 2. Adopted depth extents of the slabs were: 0-200 m, 200-450m, 450-1000 m, 1000-1500 m, 1500-2000 m, 2000-2500 m, 2500-3000 m, 3000-3500 m, 3500-4000 m and 4000-4300 m. The gravity model assumes that the loading of the crust by the Nauru edifice is compensated by a regional, rather than local isostatic mechanism. Input to the least-squares inversion analysis included gravity data from all 228 gravity stations.

The calculations give the best-fit density of the island pedestal as 2.51 t m$^{-3}$ (assuming a sea water density of 1.03 t m$^{-3}$, with correlation coefficient of 0.94 and standard deviation 1.9 mgal. A by-product of the analysis is the derivation of a gravity datum, which allows approximate conversion of relative Bouguer gravity to absolute Bouguer gravity. For gravity values shown in Figure 4 conversion was effected by adding 183 mgal, yielding a Bouguer anomaly high over Nauru of 205 mgal. The calculated mean density of 2.51 t m$^{-3}$ is typical of Pacific seamounts and suggests a lithology composed largely of basalt flows beneath the carbonate platform.

Residual (observed minus calculated) gravity is depicted in Figure 5 as 0.5 mgal contours. The residual gravity field is fairly flat with a range of only -2.5 to +3.5 mgal. A broad high of about 2.4 mgal is located over the central part of the island and extends towards the eastern coast. Arcuate lows, about 0.8 km inland and sub-parallel to the coast, are present in the southwest and northern parts of the island. A band of high residual gravity extends along the eastern coastline. The calculation of residual gravity for stations along the coastal strip is sensitive to near-shore variations in bathymetry. Only the bathymetry off the southwest coast is well known, and the remainder of the coastal bathymetry has had to be interpolated. It is
possible that some of the relatively high residual gravity values along the coastline may partly be due to the submarine slope being more gentle than assumed.

The residual gravity contours indicate no other prominent local or linear structural features that can be interpreted without further control such as exploratory deep drilling. The variations in residual gravity over Nauru probably reflect minor density changes within the carbonate platform and/or interior of the volcanic edifice produced by the heterogeneous distribution of diverse lithologies. Ash, hyaloclastite and vesicular basalts are generally less dense than basaltic intrusions and non-vesicular flows erupted in deep water. Such lower density materials may be responsible for the arcuate gravity lows over the northern and southwest parts of the island. The central gravity high may be the expression of a volcanic plug or intrusive complex with density slightly higher than that of surrounding flows and tuffs. The diameter of the upper part of this structure is about 3 km. Assuming that the structure has steeply sloping sides and extends through the full height of the volcanic pedestal, the observed gravity anomaly represents a positive density contrast of less than 0.05 $t\ m^{-3}$.

**MAGNETIC SURVEY**

Magnetic measurements were made by proton precession magnetometer at a total of 1360 stations. Sensor height was 2.5 m. The strategy adopted was to complete a series of relatively long detailed traverses, in combination with isolated stations to fill data gaps over the island. It was envisaged that high resolution mapping of local anomalies along traverses would yield source depths, while the
combination of traverses and isolated stations would provide the overall field of the volcanic substructure for mathematical modelling of its magnetic and geometric parameters.

Along traverses, readings were taken at 18 m intervals. To correct for diurnal variation, a main base station was set up near Meneng Hotel and subsidiary base stations were occupied at points along the N-S mine road in the interior of the island. Station readings were corrected for diurnal variation relative to a datum taken as the mean value of readings at the Meneng base station (35870 nT).

Much of Nauru is prone to man-made magnetic noise, owing to urban development along the coastal terrace and intensive mining activity in the interior. The problem is compounded in the interior by the fact that access is restricted to the mining roads because of the impenetrable karst pinnacle terrain. Abandoned mining equipment and other ferrous rubbish line these roads and are often difficult to detect visually because they have become overgrown by lush vegetation or have been incorporated as backfill in the road construction. Although the magnetic noise was a problem, the ultimate success of the survey was not affected. Steps taken to reduce the problem included: (i) taking a number of readings at scattered points around the isolated stations to check for large gradients indicative of non-geologic magnetic sources; and (ii) surveying on the reef-flats at low tide. The outer reef-flats were an almost magnetic noise-free environment because of the absence of metallic litter.

The results of the survey are shown in Figure 6, which depicts the data distribution and magnetic contours at 50 nT intervals. The magnetic field has a range of 830 nT, going from a low of about 35140
nT over the NNE coastline to an E-W elongated high of 35970 nT over the southern sector of Nauru. The pattern suggests a reversely magnetized volcanic pedestal. Examination of the magnetic profiles along the detailed traverses indicates a smoothly varying, relatively long-wavelength field, ignoring the localized non-geological noise. This implies a fairly deep magnetic basement although the paucity of distinct high-amplitude anomalies makes estimates of depth difficult. However, from the observed gradients and subtle variation in the profiles, the source depth is estimated to be about 400-500 m.

MAGNETIC MODEL

Magnetic modelling of the volcanic substructure of Nauru was done in a manner analogous to that of the gravity modelling, i.e. by approximating body shape by a set of polygonal prisms and estimating magnetization parameters by least-squares inversion (Plouff, 1976). Uniform magnetization of the magnetic body being modelled is assumed in the analysis. The input data for the modelling comprised 115 magnetic data points interpolated at 2 km intervals onto 9 lines distributed over Nauru. The lines were selected so that representative data coverage of Nauru was achieved, and so as to coincide with actual field traverses.

For the initial stage of modelling the magnetic field over Nauru, a uniformly magnetized island pedestal with a non-magnetic top section to represent the carbonate capping, was assumed. The same polygonal representation of the island's submarine topography as in the gravity modelling was used, except that some of the upper polygonal slabs were omitted to simulate a non-magnetic capping. This model produced an unsatisfactory fit to the observed field, indicating that the volcanic
edifice of Nauru is not uniformly magnetized, and that the anomalous field is probably due to a magnetic core within a mainly non-magnetic or weakly magnetized volcanic pedestal.

By trial-and-error variations in geometry of the body representing a magnetized volcanic core, two simple but geologically plausible models were derived which produced an excellent fit to the observed data. The fit of Model 1 is marginally better than that of Model 2. The geometry of Models 1 and 2, together with their calculated magnetic fields are presented in Figures 7 and 8. Magnetization parameters for these magnetic models are given in Table 1.

Model 1 represents a vertically-sided core laterally confined within the coastline of Nauru except for a minor excursion beneath the eastern side of the island. The top surface is horizontal but lies at two levels, 400 m deep beneath southern Nauru, and 600 m deep beneath northern Nauru. Model 2 is similar to Model 1, though less angular in overall shape. The sides of Model 2 slope at about $35^\circ$ in contrast to the vertical sides of Model 1. The shallowest level of the upper surface of Model 2 is also located beneath southern Nauru at a depth of 400 m.

The magnetic core represented by Model 1 may consist of an intra-volcanic intrusive complex while Model 2 suggests an intrusive complex or lava dome. The relatively flat upper surface of the core may have been produced by planation of the original volcano to sea-level. The surface shows a general dip to the north, possibly caused by slight tilting of the whole edifice in this direction or by a system of E-W striking normal faults with downthrow to the north.
According to the International Geomagnetic Reference Field model (Peddie, 1982), total magnetic intensity at the Meneng Hotel base-station for the time of the survey is estimated as 36171 nT. The reference field over the rest of the island is expected to be within about 20 nT of this figure. The magnetic base levels derived by modelling (Table 2) should correspond approximately to the reference field, but are in fact about 700 nT lower. The reason for this is not clear. It may be that the reference field is not a good representation of the magnetic field in the Nauru region because of inadequate local magnetic station control. Alternatively, Nauru may be located within a trough of the seafloor magnetic lineation pattern (Larson, 1976), or the base of the island pedestal may possess a significant normal-polarity magnetization.

PALAEOMAGNETISM AND AGE OF NAURU

The magnetic lineation pattern in the Nauru Basin (Larson, 1976) implies that Nauru was constructed on oceanic crust formed at about the time of magnetic anomaly M15 i.e. 132 Ma ago (Early Cretaceous). This represents the maximum possible age for Nauru.

Assuming that the geomagnetic field is essentially that of an axial geocentric dipole when averaged over a period of several thousand years, latitude and orientation of a seamount at the time of its construction can be determined from its magnetization direction. This is subject to the additional provisos: (i) that the seamount must have formed over a long enough time interval so that geomagnetic secular variations are averaged out; and (ii) the magnetic anomaly must be primarily due to thermoremanent magnetization acquired by the volcanic rocks as they cooled through the Curie temperature. The
seamount palaeomagnetic technique is discussed by Grossling (1967) and has been successfully employed by a number of workers (e.g., Francheteau et al., 1970; McNutt & Batiza, 1981; Sager, 1987).

The two models for the magnetic core yield slightly different magnetization vectors, but an estimate of the age of Nauru can still be obtained. Calculated palaeomagnetic poles, and palaeolatitudes for Model 1 and Model 2 are shown in Table 2. The palaeomagnetic poles for Nauru plot just west of the apparent polar wander curve for the Pacific plate (Suarez & Molnar, 1980).

The degree of misfit implies a possible rotation of the Nauru area by about 8° in a clockwise direction relative to the rest of the Pacific plate. There is some evidence for differential movement within the Pacific plate during early-mid Tertiary (Francheteau et al., 1970; Gordon & Cox, 1980; Gordon & Jurdy, 1986) basically involving a split between the northeastern and southwestern parts of the plate. The location of the postulated plate boundary (or boundaries) is not well constrained, nor is the relative motion between the sub-plates. The palaeolatitude data indicate that since its evolution, Nauru has drifted northward on the Pacific plate by 12.3° according to Model 1 or 6.6° according to Model 2. From absolute Cainozoic angular velocities of the Pacific plate given by Gordon & Jurdy (1986), the northward movement of the Pacific plate in the region of Nauru's current position is estimated as 25 mm/year from 0 to 43 Ma and 72 mm/year from 43 to 64 Ma. It follows, therefore, that the age of Nauru is 47 Ma according to Model 1 and 29 m.y. according to Model 2. Combination of the two results puts the probable time of Nauru's evolution in the period mid-Eocene to Oligocene. This assumes that the northward drift rate of Nauru on a
possible sub-plate was much the same as that of the main (northeastern) Pacific plate. During this period, the earth's field was of reverse polarity over at least 17 intervals (LaBrecque et al., 1977), with a particularly long interval of mainly reverse polarity between 34-38 Ma. The bend in the Hawaiian-Emperor chain marks a major change in the motion of the Pacific plate at 43 Ma (Dalrymple et al., 1980), and plate re-organisation associated with this event may have triggered the volcanism which created Nauru. In the local region, such plate re-organisation was tectonically expressed as initiation of subduction in the late Eocene at the North Solomon Trough to the southwest of Ontong Java Plateau (Kroenke, 1984). Subduction continued throughout the Oligocene.

Assuming fixed mantle hot spots, Duncan & Clague (1985) have reconstructed the position and age of hot spot traces on the Pacific plate. Their plot (Fig. 9) shows the Samoa hot spot passing within a few hundred kilometres of Nauru's present position at about 38 Ma. The bend in the predicted hot spot trace corresponding to a change in plate motion at 42-43 Ma also lies close to Nauru. This suggests that Nauru may have been constructed by hot spot volcanism, possibly coinciding with major plate re-organisation as mentioned previously. The palaeolatitude of 12.8° derived from Model 1 closely matches the 14° latitude of the eastern end of the Samoan island chain where the hot spot is presently located.

DISCUSSION: SUBSIDENCE AND UPLIFT

The abyssal plain around Nauru lies at a depth of about 4300 m. According to the seafloor subsidence/crustal age relationship of Parsons & Sclater (1977), 132 Ma oceanic crust should be 6000 m deep.
The discrepancy could partly be explained by shallowing of seafloor due to build-up of seafloor material since the crust formed at a mid-ocean ridge crest. This material would include the Cretaceous volcanic complex and overlying Tertiary pelagic sediments. Isostatic adjustment of this crustal load means that the effective elevation of the seafloor would be significantly less than the total thickness of the volcanic/sediment pile. Build-up of seafloor to 4300 m would require the volcanic/sediment layer to have developed a thickness well in excess of 1700 m, which appears unreasonable. The vertical tectonics in the Nauru Basin are better explained by thermal uplift in the mid- to Late Cretaceous, for which there is evidence from drilling in the northern Marshall Islands and also Site 462 (Detrick & Crough, 1978; Schlanger & Premoli Silva, 1981).

The regional volcano-thermal episode ended about 70 Ma and was followed by continued subsidence amounting to 1600 m (Schlanger & Premoli Silva, 1981). The subsidence curve for Bikini and Enewetak atolls suggest that the thermal age of the lithosphere had been reset to 25 Ma by the event (Detrick & Crough, 1978). The curve indicates a fairly constant rate of subsidence since late Eocene averaging about 18 m/Ma.

From the geophysical data, the thickness of the carbonate platform on Nauru is interpreted as about 550 m. With the assumption of 18 m/Ma subsidence, the age of Nauru works out to be 31 Ma. However, since Nauru occupies a position on the former upper flank, rather than the crest, of the postulated Darwin Rise (Menard, 1964) the subsidence rate may have been lower, perhaps three-quarters of that given by the curve. This would yield an estimated age of about 41 Ma for Nauru, which is consistent with the palaeomagnetically derived age.
Some modification of the subsidence history outlined above may be required if Nauru was created by hot spot volcanism. An increased subsidence rate would be expected due to resetting of a thermal age of the lithosphere (Detrick & Crough, 1978). This would have the effect of lowering the above subsidence-based age estimates. An extreme example of a large thermal anomaly is that producing the Hawaiian chain. Detrick & Crough (1978) established that lithosphere which has drifted over the top of this hot spot begins to subside at a rate equivalent to that of normal 30 Ma old crust. At this rate 550 m of subsidence would result in about 17 Ma. The Samoan chain differs in its physiography and pattern of volcanism from other Pacific island chains, including the Hawaiian chain (Keating, 1985). An important difference is the apparent absence of a well defined bathymetric swell along the Samoa hot spot trace and no clear pattern of subsidence along the chain. In addition, there appear to be large gaps in the chain, particularly to the southeast of Nauru. These anomalies suggest that the Samoa hot spot may be weak and/or intermittent, with recent extensive volcanism perhaps mainly attributable to plate deformation associated with subduction in the nearby Tonga Trench (Natland, 1980). If, as seems likely, the Samoa hot spot represents only a minor thermal phenomenon and if hot spot volcanism was responsible for creating Nauru, then an age estimate for Nauru based on subsidence would be much closer to the 41 Ma age previously derived, than a younger age (17 Ma) appropriate only for major hot spots.

The emergence of Nauru and Ocean Island 300 km to the ESE is attributed by Menard (1973) and McNutt & Menard (1978) to uplift of the seafloor as the Pacific Plate rides over a 'bump' in the
underlying asthenosphere. Convection cells in the mantle are believed to produce the crustal warping reflected as depth anomalies which generally have amplitudes of about 300 m and wavelengths of 500-1000 km. The bumps are weakly correlated with free-air gravity anomalies. The uplift of Nauru and Ocean Island is ascribed to their location on the flank of a gravity high lying to the northwest, in the direction of plate motion. However, examination of SEASAT-derived free-air gravity images of the region (Haxby, 1987) shows that the gravity field is complex. A gravity high of about 15 mgal does exist northwest of Nauru, but both Nauru and Ocean Island lie within gravity depressions. A further complicating factor is the elevation difference between the two islands. Nauru would be expected to be the higher of the two islands if they were both rising up the side of a positive depth anomaly centred to the northwest. Ocean Island is a coral dome reaching a height of about 80 m above sea-level (Power, 1905; Nugent 1948), while the higher knolls on Nauru are only about 55 m above sea-level although a peak on the raised rim in the southwest of the island does attain a height of 70 m.

Nevertheless, uplift of the Pacific plate as it rides over a minor mantle thermal anomaly is still the most plausible explanation for the emergence of Nauru (and Ocean Island). The presence of an undulating surface on the depth anomaly could account for the apparent inconsistencies and it could be argued that, assuming Nauru was the first to be affected by the uplift, it has suffered the greater height reduction by longer exposure to subaerial erosion and dissolution.

CONCLUSIONS

1. Gravity and magnetic surveys show that the substructure of
Nauru Island is approximately radially symmetrical, and has a density of 2.5 t m⁻³.

2. Magnetic modelling indicates that the island has a reversely magnetized volcanic core with magnetization 1.5 - 1.9 A m⁻¹, at a depth of 500 m beneath the limestone capping, the upper part of which is late Miocene or younger.

3. The age of the volcanic core is estimated as mid-Eocene to Oligocene, based on the orientation of the magnetization vector, and an estimate of island subsidence.

4. The emergence of the island is probably due to uplift of the Pacific plate as it rides over a mantle thermal anomaly.

ACKNOWLEDGEMENTS

Fieldwork was facilitated by the Nauru Phosphate Corporation, and we thank David Newick (General Manager), Don Lauder (Chemist) and Stan Nowak (Surveyor) for their assistance in this regard. The investigation was undertaken on behalf of the Commission of Inquiry into Rehabilitation of the Worked-out Phosphate Lands in Nauru.

George Chaproniere of BMR undertook foraminiferal determinations, and carbon dating of coral samples was done by the University of Waikato, New Zealand. Bathymetric data was provided by the Defense Mapping Agency, Aerospace Center, St. Louis, U.S.A. and by the National Geophysical Data Center, National Oceanic and Atmospheric Administration, Boulder, Colorado, U.S.A. The paper has benefited
from critical review by Mart Idnurm and Jim Colwell (both of BMR).

Joan Brushett typed the manuscript and Joe Mifsud drafted the figures.

The authors publish with the permission of the Director, BMR.
Table 1. Radiocarbon dates on drillcores
(by University of Waikato Radiocarbon Dating Laboratory)

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Drillhole</th>
<th>Depth (m)</th>
<th>Conventional age (^1) (years B.P.)</th>
<th>True age (^2) (years B.P.)</th>
<th>^^14C depletion (^3)</th>
<th>Delta (^{13})C</th>
<th>Material</th>
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<tbody>
<tr>
<td></td>
<td>WK 1149</td>
<td>Q1</td>
<td>2660 +/- 60</td>
<td>2730 +/- 60</td>
<td>-281 +/- 4.6</td>
<td>-0.1 (^\circ)/oo</td>
<td>Coral</td>
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<tr>
<td></td>
<td>WK 1150</td>
<td>Q1</td>
<td>2740 +/- 60</td>
<td>2823 +/- 60</td>
<td>-288.9 +/- 4.6</td>
<td>0</td>
<td>Coral</td>
</tr>
</tbody>
</table>

\(^1\) 'Conventional age' implies use of Libby half-life (5568 years) with AD 1950 as reference year, and assumed constancy of atmospheric radiocarbon levels.
'True age' is based on new half-life of 5730 years.

Carbon-14 depletion expressed as ‰ with respect to 0.95 oxalic acid.
### Table 2. Magnetisation parameters for Nauru magnetic models

<table>
<thead>
<tr>
<th></th>
<th>Model 1</th>
<th>Model 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Inclination (+ down)</td>
<td>24.4°</td>
<td>14.0°</td>
</tr>
<tr>
<td>Declination (+ east)</td>
<td>186°</td>
<td>188°</td>
</tr>
<tr>
<td>Intensity (A m(^{-1}))</td>
<td>1.47</td>
<td>1.93</td>
</tr>
<tr>
<td>Standard deviation, 115 data points (nT)</td>
<td>40</td>
<td>49</td>
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<tr>
<td>Multiple correlation coefficient</td>
<td>0.98</td>
<td>0.97</td>
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<tr>
<td>Magnetic datum (nT)</td>
<td>35495</td>
<td>35348</td>
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<tr>
<td>Palaeomagnetic pole</td>
<td>76°N 321°E</td>
<td>80°N 297°E</td>
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<tr>
<td>Palaeolatitude (based on inclination)</td>
<td>12.8°S</td>
<td>7.1°S</td>
</tr>
</tbody>
</table>


HUTCHINSON, G.E., 1950. Survey of existing knowledge of biogeochemistry; 3. The biogeochemistry of vertebrate


SAGER, W.W., 1987. Late Eocene and Maastrichtian paleomagnetic poles for the Pacific plate: implications for the validity of
seamount paleomagnetic data. *Tectonophysics* 144, 301-314.


FIGURE CAPTIONS

Fig. 1  Locality map and tectonic setting. Magnetic lineations after Larson (1976); absolute plate movement (broad arrows) after Minster & Jordon (1978).

Fig. 2  Nauru bathymetry.

Fig. 3  Geology and drillhole locations.

Fig. 4  Relative Bouguer gravity anomaly (terrain corrected for a.s.l. topography).

Fig. 5  Residual gravity.

Fig. 6  Total magnetic intensity - corrected field results.

Fig. 7  Theoretical magnetic field for magnetic core - Model 1.

Fig. 8  Theoretical magnetic anomaly field for magnetic core - Model 2.

Fig. 9  Predicted hot spot traces on the Pacific plate (after Duncan & Clague, 1985) and relationship to location of Nauru. Hotspots include Samoa (SM), Louisville Ridge (LV), Macdonald Seamount in the Austral Islands (MC), Pitcairn Island (PT), Mehetia in the Society Islands (ME), the Caroline Islands (CA), the Marquesas Islands (MQ), Easter Island (EA), Sala y Gomez (SG), Hawaii (HW),
Socorro Island in the Revillagigedos Islands (SC), Cobb Seamount (CB) and Dellwood Knolls (DK).
Ontong Java Plateau

Locality map & tectonic setting

FIGURE 1 (S7).
FIGURE 2 (S7).
PACIFIC OCEAN

**Figure 3 (S7).**

Geology and drillhole locations
FIGURE 4 (S7).
FIGURE 5 (S7).
Total magnetic intensity, corrected field results

FIGURE 6 (S7).
Intensity of magnetisation 1.47 A m$^{-1}$  Declination 186°  Inclination 24°  Regional magnetic correction of 495 nT added
Magnetic core is represented by polygonal bodies with depth extents (km)  G 0.4 - 0.6  H 0.6 - 4.3
Theoretical magnetic anomaly field for magnetic core — Model 1

FIGURE 7 (S7).
Intensity of magnetization: 1.93 A m⁻¹  
Declination: 188°  
Inclination: 14°  
Regional magnetic correction of 348 nT added

Magnetic core is represented by polygonal bodies with depth extents (km)

A 0.4 – 0.5  
B 0.5 – 0.7  
C 0.7 – 1.0  
D 1.0 – 1.5  
E 1.5 – 3.0  
F 3.0 – 4.3

Theoretical magnetic anomaly field for magnetic core – Model 2

FIGURE 8 (S7).
SM  Samoa hot spot  Ages in Ma

Predicted hot spot traces

FIGURE 9 (S7).
HYDROGEOLOGY OF NAURU - RESULTS OF A GEOPHYSICAL, DRILLING
AND HYDROLOGICAL INVESTIGATION

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ABSTRACT

Nauru, in the central Pacific Ocean, is a raised atoll capping a
volcanic seamount arising from an ocean floor depth of 4300 m. The
land area is 22 km$^2$, and the island rises to 70 m above sea level.
Drilling has proved dolomitised limestone of upper Miocene or younger
age to a depth of 55 m below sea level. Gravity and magnetic surveys
indicate that the limestone probably overlies volcanic bedrock at a
depth of about 500 m. Reverse-circulation drilling and geoelectrical
probes indicate that there is a discontinuous freshwater layer in the
limestone averaging 5 m thick and lying at a depth close to sea level.
This layer is underlain by a mixing zone of brackish water, 60-70 m
thick which in turn is underlain by sea water. The exceptional
thickness of the mixing zone is ascribed to high permeability of the karstified limestone. Radioactivity of the soils in the central part of the island is anomalously high, but is not expected to pose a threat to groundwater potability. The forthcoming cessation of phosphate mining will mean a shortfall in water supply which will probably have to be met by the desalination of brackish water. Groundwater beneath the mined-out area, and the settled coastal terrace, is highly vulnerable to pollution, and waste disposal management needs to be considered in relation to groundwater protection.

INTRODUCTION

Nauru, which supports a population of 8500, occupies a land area of 22 km$^2$ in the central Pacific Ocean at $0^\circ$ 32'S, 166°56'E (Fig. 1). The island has been mined for its surficial phosphate deposits for about 80 years, and the current reserves indicate that the mine has approximately 9 years' life left. About 80 percent of the land area has had its vegetation, soil cover and phosphate removed leaving an exposed limestone pinnacle surface, or karrenfeld. The pinnacles are residual features left after karstic solution of the limestone, their surface relief accentuated by the excavation of interjacent phosphate.

The present water supply is derived from rainwater tanks supplemented by dugwells and by water imported as ballast on the phosphate ships. Rehabilitation of the mined out land is under consideration and this is expected to increase the demand for water at a time when imported water will no longer be available. In this context an investigation of the hydrogeology and groundwater resources of Nauru was undertaken by a BMR team on behalf of the Commission of Inquiry into
Rehabilitation of the Worked-out Phosphate Lands in Nauru. The field investigation included inspections and sampling of springs and wells; gravity and magnetic surveys to determine the depth and shape of the island's basement; electrical resistivity surveys and drilling to determine the thickness of the freshwater layer; a gamma-ray spectrometer survey of the island to establish radioactivity and radio-element concentrations in the soil; and measurements of tidal response in bores.

Previous investigations of Nauru's groundwater resources were undertaken by the British Phosphate Commission (1965) and Australian Groundwater Consultants (1972) but results were inclusive. Our combined use of reverse-circulation drilling and geophysical techniques has enabled an estimate of Nauru's groundwater resources to be made. Details of the island's substructure and evolution are reported in a companion paper (Hill & Jacobson, in review). We report here the significant aspects of the island's hydrogeology.

HYDROGEOLOGY

Nauru is a raised coral atoll (Fig. 2) with maximum elevation 70 m above sea level. It is underlaid by a volcanic seamount that rises 4300 m from the floor of the Pacific Ocean.

The results of gravity and magnetic surveys in the present investigation indicate that about 500 m of dolomitised limestone cap the seamount. The limestone has been drilled in the present investigation to a depth of 55 m below sea level, and is intensely karstified to that depth, with phosphate cavity fillings. Micro-palaeontological examination of core samples by G.C.H. Chaproniere
(BMR) indicates that the limestone is of upper Miocene to Quaternary age to the depth tested by drilling. The base of the limestone platform may be as old as mid-Eocene (Hill & Jacobson, in review).

Field investigation - geoelectrical soundings and drilling

With a reverse circulation drilling rig, fragmented drillcore was obtained, and water samples were obtained at intervals for electrical conductivity measurement. Some difficulty was experienced in delineating groundwater from drilling water which was used in the section of the drillhole above the aquifer. A total of 12 holes (see Figure 2 for locations) were drilled to depths ranging from 26 to 83 m (Table 1). Three of the holes were completed as open-hole piezometers with 40 mm diameter plastic casing to enable the measurement of standing water level. Caving ground prevented the installation of casing in the other drillholes. Geological logs of the drillholes and electrical conductivity measurements of groundwater samples are given in Figures 3 and 4.

Locations of the Schlumberger-configuration geoelectrical soundings are shown in Figure 2 and a list is given in Table 2. Theoretical and practical aspects of the 4-electrode Schlumberger sounding method are described by Koefoed (1979). A total of 10 soundings were made in three hydrogeological environments:

(i) in the interior of Nauru, at elevations 12-27 m (depth probes DP 1, 2, 6 & 8);
(ii) adjacent to Buada Lagoon, elevations 2-3 m (depth probes DP 3, 4, 5 & 7); and
(iii) on the coastal terrace, elevations 3-4 m (depth probes DP 9 & 10).
Field curves of apparent resistivity and their interpretation for DP 1-10 are presented in the Appendix. Interpretation was done by forward modelling, computing the apparent resistivity model curves by the linear filter method of O'Neill (1975).

The instrument used was an ABEM Terrameter SAS 300B with Booster SAS 2000. A typical field curve of apparent resistivity and its interpretation shown in Figure 5. Drilling results show that the water-table invariably lies just above mean sea-level. The resistivity models were constrained to comply with this observation, and the depth of resistivity reduction associated with the top of the aquifer was set close to mean sea level (i.e. set to R.L. 1.0 m).

Routine interpretation of subsurface resistivity structure is practical only for horizontal layering because of the complex mathematical analysis and modified field techniques required for more complicated configurations. Therefore in selecting electrical sounding sites on Nauru, the requirement for at least approximate horizontal resistivity stratification was an important consideration. Site selection was accordingly made on the basis of available geological information, particularly surface indications. With the shallow to medium-depth soundings there appear to be no problems. However, two deeper soundings, DP6 (AB/2 = 400 m) and DP8 (AB/2 = 900 m) were completed in the interior of Nauru where long electrical arrays are possible only along linear sections of the mine roads and in these cases some departure from horizontal stratification is evident from distortion of the field curves. Theoretically, if horizontal stratification applies, the apparent resistivity curves cannot rise at slopes greater than 45°. The distortion of the field
curves is caused by the channelling of electrical current through the relatively low-resistivity soil and phosphate forming the road foundation. In contrast, the pinnacled, worked-out areas adjacent to the roads are of significantly higher electrical resistivity.

Because of the distortion, the field curves for DP6 and DP8 cannot be interpreted accurately, although they do provide an approximate model for resistivity layering. The interpretations shown for DP6 and DP8 indicate the probable resistivity structure, based partly on inference and partly on the field curves. The freshwater layer thickness, in particular, cannot be determined precisely. An important feature of the DP6 and DP8 field curves is that the curves descend rapidly at large electrode spacings, without any apparent up-turning. This confirms gravity and magnetic survey results that any dense volcanic core must be deeper than about 200 m. Hydrogeological interpretations of the resistivity models for DP1-10 are provided in Table 3.

**Freshwater layer and mixing zone**

Drilling and geoelectric soundings show that Nauru is underlain by a discontinuous layer of freshwater up to 7 m thick (Fig. 6), overlying a mixing zone of brackish water up to 60 m thick, which in turn overlies sea water. Salinity increases gradationally downwards as shown in cross-sections (Figs. 7 and 8).

The water-table, the upper boundary of the freshwater layer, is at an average elevation of R.L. 1.50, or 0.30 m above mean sea level, throughout Nauru, with the exception of the Buada Lagoon catchment which is a different hydrological system. According to topographical survey data Buada Lagoon is at an elevation of R.L. 2.40; it is
perched above the regional water-table, on impermeable phosphatic alluvium (Fig. 8).

The average thickness of the freshwater layer is 4.7 m based on intersection in 11 drillholes (Figs 3 and 4). Its lower boundary is defined at a salinity level of 1500 mg/L total dissolved solids, which is equivalent to electrical conductivity of 2200 microSiemens/cm and is the upper limit for drinking water.

The unusually thick mixing zone of brackish water is due to high permeability in the limestone; open karst fissures allow intrusion of seawater beneath the island and diffusion to form a zone of brackish water. Quantitative estimates of hydraulic conductivity have not been undertaken on Nauru, but by analogy with similar raised limestone islands and plateaux elsewhere it is probably 3000-3500 m/d. Estimates have been made for Barbados, 3400 m/d (Goodwin, 1980); Florida, 3000 m/d (Kohout, 1960); and Northern Guam, up to 3500 m/d (Goodrich and Mink, 1983).

**Groundwater recharge**

The annual rainfall for Nauru is shown in Figure 9. Records are available for 60 years, but there are significant gaps with no information. The available records indicate a mean annual rainfall of 1994 mm, with a high degree of variability; the maximum recorded was 4590 mm in 1930 and the minimum 280 mm in 1950.

High annual rainfalls commonly occur in the years corresponding to, or immediately following, major El Nino-Southern Oscillation Events (Philander, 1983) such as those of 1957, 1964, 1972 and 1982 (Fig.
The occurrence of this phenomenon every few years is probably responsible for the bimodal frequency distribution of Nauru annual rainfall (Fig. 10).

Ten of the 60 years for which records are available, had less than 1000 mm of rain. Drought periods with less than 1000 mm in 3 months occurred in 34 of the 60 years. The most severe historical droughts occurred in 1916-17, when 95 mm of rain was recorded in 8 months; 1938, when 97 mm was recorded in 8 months; and 1950, when 83 mm was recorded in 6 months.

The seasonality of the rainfall is illustrated in Fig. 11. The mean monthly rainfall ranges from 118 mm in May to 279 mm in January, with the months from December to March each averaging more than 200 mm.

Potential evapotranspiration has been calculated on the basis of Fleming's (1987) formula, derived for Tarawa atoll which has a similar climate to Nauru and is 700 km away. Figure 12 shows the empirical relationship between monthly rainfall and potential evapotranspiration for Tarawa, which is

\[ E = 115 + (300 - P)^2/1286 \]

where P is the monthly rainfall, which is less than 300 mm, and E is the monthly potential evapotranspiration. On this basis, using monthly data, the potential evapotranspiration estimated for Nauru ranges from 115 mm in January to 141 mm in May (Table 4) and the mean annual total is 1547 mm.

Actual evapotranspiration is less than potential evapotranspiration. To estimate actual evapotranspiration requires information on soil moisture and vegetation characteristics which is not available for
Nauru. Disregarding some groundwater discharge to, and surface water evaporation from, lagoons, the approximate water balance for Nauru can be considered as:

\[ \text{Rainfall} = \text{Actual evapotranspiration} + \text{Groundwater recharge}. \]

A detailed estimate of actual evapotranspiration for another central Pacific atoll, Kiritimati, showed that recharge is 29% of rainfall for a freshwater lens where the land is not vegetated (Falkland, 1984). This atoll is appreciably drier than Nauru. On Tarawa, the closest atoll to Nauru where information is available, and with comparable annual rainfall, recharge was estimated at 34% of rainfall (Daniell 1983). This coefficient was adopted for an atoll with 80% coconut tree cover and 20% grass. On Nauru, with its extensive bare karrenfield surface, recharge is probably about 40% of rainfall. Adopting this latter value for the recharge/rainfall ratio, we derive an annual water balance for Nauru:

\[ \text{Rainfall (1994 mm)} = \text{Actual evapotranspiration (1196 mm)} + \text{groundwater recharge (796 mm)}. \]

Recharge is assumed to be uniform throughout the island although it is probably greater inland, beneath the mined-out area.

**Groundwater flow and discharge**

The amount of recharge indicates that a substantial amount of groundwater flows through the Nauru groundwater system. Water-table measurements in present and previous investigation drillholes indicate that the groundwater head throughout most of the interior of Nauru is
about R.L. 1.50 which is about 0.30 m above mean sea level. There are some difficulties in establishing this, as tidal oscillations of groundwater-level are significant, and mean sea level is also not static. Figure 13 shows the relationship of adopted mean sea level on Nauru to the island datum.

Averaged elevations of the water-table are shown in Figure 14. Groundwater flow is probably radially outwards to the sea, and is generated by the 0.30 m head differential between the inland water-table and mean sea level. Buada Lagoon is perched (R.L. 2.40) above the general water-table (R.L. 1.50) which extends throughout the island. Measurements of Buada Lagoon water levels in October 1987 indicated that the lagoon is not tidal but its level is affected by rainfall and evaporation.

Groundwater discharges around the circumference of the island, to the chain of lagoons at Anabar in the northeast, and elsewhere to springs in the reef (Fig. 14). Several caves at the inland edge of the coastal terrace provide a window on the water-table close to the discharge end of the flow system. The largest of these is Maqua Pool in the southwest of Nauru.

**Tidal fluctuations**

Tidal information for Nauru is available for 13 years. The mean monthly sea level is 1.169 m above tide gauge zero. This compares with the adopted mean sea level for survey purposes on Nauru, of 1.186 m above tide gauge zero (Fig. 13), equivalent to R.L. 1.352.

The tidal range at Nauru is from R.L. 0.86 m, the lowest mean daily
tide, to R.L. 1.70 m, the highest mean daily tide, over the 13 years records. The effect of daily and longer term fluctuations in ocean tide level is shown in Figure 15. There is a reversal of hydraulic gradient at the shore line with drainage outwards at low tide, and seawater flow inwards at high tide.

Tidal effects in observation bore P3, 800 m inland, were measured during the present investigation, and additional information on this phenomenon is available from an unpublished report (Gormley, 1987). Tidal effects on groundwater levels are substantial, being close to half the amplitude of the ocean tidal stage throughout the island (Fig. 16). The tidal movement of the water-table is commonly of order of 0.5 m and the lag of the tidal peak in water bores is generally 1.5-3 hours.

Figure 17 shows the tidal efficiency, which is the ratio of tidal movement in groundwater to that in the ocean, plotted against distance from the sea; and the time lag plotted against distance from the sea.

HYDROCHEMISTRY

Typical chemical analyses of Nauru waters are shown in Table 5.

The relationship between electrical conductivity and total dissolved solids for Nauru waters is shown in Figure 18. The total dissolved solids content in mg/L is 0.69 x the electrical conductivity in microS/cm, with a high degree of correlation.

Nauru rainwater contains about 10 mg/L total dissolved solids, and is slightly acid, and bicarbonate-rich. Buada Lagoon was fresh in
October 1987, with about 200 mg/L total dissolved solids; it is believed to become brackish with evaporation in long dry periods. The Anabar Lagoons are brackish with about 5000 mg/L total dissolved solids. Accessible cave waters at Ijuh and Anatan are potable and used for small-scale water supplies. The largest cave supply, at Maqua Cave, is brackish, containing about 1750 mg/L total dissolved solids.

Samples from the freshwater layer taken during drilling (P6 and W1) are in the range 85-295 mg/L total dissolved solids. The freshwater is bicarbonate rich, and is moderately hard (55-108 mg/L). The pH is between 6.90 and 7.80 and temperature 25°-26°. Low nitrate, fluoride and iron contents confirm its suitability for drinking water.

Samples from the brackish water zone (drillholes P4, H10, P5, P3, W3) are increasingly saline with depth, and the water is very hard with more than 400 mg/L total hardness. These waters are alkaline with pH between 7.4 and 8.9. With increasing salinity the groundwater approaches seawater composition and becomes a chloride water, with sodium the dominant cation. The nitrate content is low, less than 13 mg/L, with the exception of samples from drillhole H10, in the centre of the island, which might be polluted by septic tanks. Fluoride and iron levels are also low in the brackish groundwater samples.

Chemical analyses of 22 wells sampled in the coastal terrace (Fig. 19) are in the range 290-3245 mg/L total dissolved solids. Of these wells 17 were within the limit for drinking water of 1500 mg/L dissolved solids, and contain more saline water. The variation in salinity is due to factors such as the depth of the well, and its pumping rate; it is likely that nearly everywhere in the coastal terrace a thin layer
of freshwater overlies a mixing zone of brackish water. Composition ranges from bicarbonate dominant in the fresher waters to chloride dominant in the saltier waters (Fig. 20). Nearly all the coastal plain groundwaters are hard or very hard, with total hardness between 172 and 776 mg/L. Most of these waters are slightly acid: pH ranges from 6.70 to 7.16. The temperature of the coastal terrace groundwater is constant at 28°C. Nitrate levels are generally low, up to 35 mg/L and all are within recommended limits for drinking water. Fluoride and iron levels are low. However 14 wells were bacteriologically polluted, when sampled.

WATER SUPPLY

The present water consumption is estimated as about 1300 m³/day, derived from the following sources:

Rainwater catchments 700
Imported water (phosphate ships) 400
Groundwater (coastal terrace wells) 200

These estimates are based on calculations at 140 L/capita/day including industrial use but excluding sewage (Hadwen, 1986). Predictions of future water demands are difficult, but the imported water will not be available when phosphate mining ceases in the next decade. Thus at present consumption rates, there would be a shortfall of 400 m³/day. If more water is required for irrigation of land rehabilitation projects then the shortfall would be greater than 400 m³/day. Additional sources that should be considered are groundwater from the inland plateau, and desalination.
Rainwater

Rainwater is particularly important in the island context as it is the best quality water available for domestic use and some industrial purposes. On Nauru it presently forms more than half the water supply. Rainwater supply systems are likely to fail in severe droughts, and further study is needed on Nauru to assess total catchment area and storage capacity installation in relation to water demand and probability of failure. Construction of additional storage capacity is desirable but this may be costly compared with groundwater extraction or desalination.

Rainwater catchments could be extended by construction of special purpose catchments and storage tanks in the interior of the island, for reticulation to the coastal terrace.

Imported water

To continue importing water by ship when mining ceases would be feasible, but may be costly in comparison with groundwater extraction or desalination.

Groundwater (coastal terrace)

The coastal terrace groundwater is abstracted from several hundred wells; it is regarded as a second class water source, being used for sewage and other secondary domestic purposes and as a backup domestic supply in drought years. The high salinity in some wells is due either to their being constructed too deep, or to overpumping. Some improvement to the present extraction of coastal plain groundwater
could be made by the introduction of skimming well/infiltration gallery technology.

Treatment would be necessary for the coastal terrace groundwater to be widely used for drinking water, owing to the variations in salinity and the incidence of bacteriological pollution. The coastal terrace groundwater would be suitable as brackish water feedstock for desalination.

**Groundwater (inland plateau)**

A large amount of groundwater is available beneath the inland plateau of Nauru. The thinness of the freshwater layer (4-5 m) precludes the use of pumping bores because of the likelihood of upconing saltwater and disrupting the freshwater layer. Bores could however be used to pump brackish water as feedstock for desalination.

The possibility of excavating shafts and infiltration galleries (tunnels) to skim the freshwater layer, requires further assessment. This technology is used in some islands to abstract water beneath thick limestone cover and has previously been proposed for Nauru (Gormley, 1987). In Barbados, horizontal galleries are developed at the bottom of shafts 40 m deep; however the minimum thickness of the freshwater layer is 13 m so that there is a buffer against upconing of saltwater (Goodwin, 1980). Freshwater lenses 8-10 m thick have been safely skimmed in some atolls, but on Nauru the thin lens and open fissures, make the prognosis doubtful.

**Desalination**
Desalination of groundwater may be the best long term option for Nauru's water supply, and costs could be more favourable than importing water or rainwater systems. The electrodialysis, reverse osmosis and Sirotherm technologies are appropriate for brackish water feedstock which would be available on the coastal terrace or inland plateau.

Reverse osmosis systems have been used successfully in many places, with either brackish water or seawater as feedstock. However this technology has relatively high energy requirements. Reverse osmosis systems need a constant salinity of feedstock water, which may not be attainable with brackish water on Nauru, so that a seawater-based system may be economic. Desalination technology is advancing rapidly, and serious consideration of this option for Nauru would require an updated technical and economic evaluation.

**WATER QUALITY CONSIDERATIONS**

Future groundwater development on Nauru requires consideration of the possibility of saltwater intrusion, and of pollution from sewage and other wastes and natural radioactive elements.

**Saltwater intrusion**

The main threat to future groundwater development would be saltwater intrusion through overpumping, and for this reason careful management of groundwater extraction will be necessary. At the present time, some saltwater intrusion is evident in the deeper and more heavily pumped coastal wells. There is also some possible contamination from overflowing saltwater storage tanks.
Sewage pollution

In order to assess the effects of pollution from sewage and animal wastes, 23 wells and 3 cave water supplies on the coastal terrace were sampled. These samples were taken at 1 km intervals around the island.

Of the 26 groundwater sources sampled, 12 had bacterial counts below the MPN index (Most Probable Number of e coli) of 23, indicating the suitability of the water for human consumption, untreated. The 14 samples with bacterial counts greater than the MPN index of 23 indicate that these water supplies would require chlorination. One of these samples had a count of 1100, and this water supply would require double chlorination. The wells with high bacterial counts are located at intervals around the coastal plain and the contamination is probably of local origin, i.e. sewage and animal wastes.

Thus, it is likely that approximately half the wells in the coastal plain aquifer are bacteriologically polluted by sewage and animal wastes. Nitrate levels in the groundwater remain low at the present time.

Waste disposal

Pollution of the main freshwater layer is possible from waste deposited in mined out karst limestone areas; known waste dumps include slimes containing zinc and cadmium, and also domestic garbage (Fig. 21).
Natural soils on Nauru contain from 6 to 173 ppm cadmium with a mean content of 70 ppm at 16 sampling points. In the calcination process of Nauru, raw phosphate is roasted at 1050° to remove cadmium and organic carbon. Sludge from the calcination plant contains about 200 ppm cadmium and about 2000 ppm zinc. The sludge has been dumped in mined out areas where leaching to the water-table could readily occur. Municipal garbage is also dumped in mined out areas.

Groundwater samples collected during the present investigation showed low levels of zinc and less than 0.01 mg/L cadmium. The upper limits in drinking water are generally taken as 5 mg/L zinc and 0.01 mg/L cadmium; the upper limit of cadmium in irrigation water is also 0.01 mg/L (Hem, 1985).

Although groundwater is not polluted from this source at present, a waste disposal policy must be considered in conjunction with proposals for future water supply development.

Radioactive elements

Some phosphatic soils are known to contain radioactive elements deleterious to health. For this reason, and because the radiometric method is potentially useful for delineating mineralization or variations in near-surface lithology, we carried out a gamma-ray spectrometer survey of Nauru soils and mapped the distribution of radiometric properties. In general, only soils that appeared to be in-situ were radiometrically sampled. This was done in an attempt to map the original, pre-mining radiometric pattern over the island.

The 4-channel 'Exploranium' DISA-400A instrument used for the survey
measures radioactive intensity over the following spectrum ranges -

(i) energy above 0.4 MeV (total count).
(ii) the K-40 1.46 MeV peak, window width 200 KeV.
(iii) the Bi-214 1.767 MeV peak, window width 200 KeV.
(iv) the Tl-208 2.62 MeV peak, window width 400 KeV.

The instrument was calibrated to yield potassium (K), uranium (U) and thorium (Th) concentrations as given by the formulae below (after channel readings have been corrected for background count, and assuming radioactive equilibrium).

\[
K(\%) = 0.2962 Kc - 0.3311 Uc - 0.0224 Thc \\
U(\text{ppm}) = 0.0254 Kc + 3.543 Uc - 1.802 Thc \\
Th(\text{ppm}) = -0.0273 Kc - 0.3709 Uc + 8.003 Thc
\]

where Kc, Uc and Thc are measured window count rates in counts/second.

Measurements were made at 58 stations fairly evenly distributed over the island. To obtain representative measurements, two or more readings were taken over different parts of the soil surface at each location. The survey results are presented in Figures 22-25 showing gamma-ray total count, calculated K(\%), U(\text{ppm}) and Th(\text{ppm}) concentrations (respectively) as well as station locations.

Measurements were also made on the new phosphate stockpile (station 19), on top of the black soil stockpile (station 20) and at a freshly exposed part of the base of the black soil stockpile (station 21).
Results for these locations are given below.

<table>
<thead>
<tr>
<th>Stn</th>
<th>Total count</th>
<th>K%</th>
<th>U(ppm)</th>
<th>Th(ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>19</td>
<td>570</td>
<td>3.14</td>
<td>37</td>
<td>40</td>
</tr>
<tr>
<td>20</td>
<td>660</td>
<td>3.67</td>
<td>49</td>
<td>46</td>
</tr>
<tr>
<td>21</td>
<td>710</td>
<td>4.12</td>
<td>55</td>
<td>49</td>
</tr>
</tbody>
</table>

These materials are not in-situ and so these data were not used in producing the contours of Figures 22-25. Data from station 55 (total count 210, K 1.23%, U 17 ppm and Th 10 ppm) were also not incorporated since station 55 was located on the golf course, on which the soil may have been introduced. Measurements made on limestone/dolomite blocks in the mined areas show a reduction of radioactivity (all 4 channels) to only about 10-40% of values obtained from the surrounding phosphatic soils. Obviously, much of the radioactivity in the soils is related to their phosphate content, as well as their organic carbon content, as indicated by the high values for the black soil stockpile.

The total count contours (Fig. 22) show a very low level of radioactivity along the coastal strip, increasing rapidly across the coastal escarpment and forming a broad high over the island's interior. The area of highest radioactivity is located to the NE of Buada Lagoon (920 cps maximum), while a subsidiary high occurs in the central NE of the island (720 cps maximum). The patterns for K, U and Th concentrations (Figs. 23-25) differ in detail, but show similar distribution. Maximum recorded concentrations are: K 4.8%, U 84 ppm, and Th 67 ppm. As a comparison, the average abundances of these elements in the earth's crust are K 2.1%, U 2.7 ppm and Th 9.6 ppm.
The concentration of uranium in the soil is relatively high, about 31 times that in average crust. The uranium is likely to be 'fixed' in the soil profile by absorbing action of the phosphate and organic carbon, and it is unlikely that sufficient leaches down to the aquifer to cause a groundwater pollution problem.

CONCLUSIONS

1. Drilling and the geoelectrical soundings indicate that the freshwater layer on Nauru is discontinuous and averages 4-5 m in thickness with a maximum of 7 m. The brackish water zone beneath the freshwater layer is unusually thick (60 m) owing to high-permeability limestone allowing ingress of seawater beneath the island. Salinity increases progressively downwards in the brackish water zone.

2. The substructure of Nauru is approximately radially symmetrical with the volcanic core at a depth of 500 m beneath the limestone, which is of mid-Eocene(?) to Quaternary age.

3. The conjunctive use of rainwater catchments and coastal terrace groundwater wells will form the basis of Nauru's water supply for the foreseeable future, and some improvements to these systems are desirable.

4. Additional groundwater development will be necessary to make up the shortfall in supply when importation of water ceases. The freshwater layer beneath the island is too thin to sustain heavy pumping from bores without upconing of saltwater. Desalination of brackish groundwater, which is available in
large quantities, is the most likely option for future development.

5. The coastal plain aquifer is polluted in part from sewage and animal wastes, as evidenced by bacterial counts in a number of wells. The level of radioactivity in Nauru soils is low enough not to pose a threat to groundwater development. Waste disposal in mined out areas is a potential hazard, and a waste disposal policy should be developed to protect future groundwater resources from pollution.

ACKNOWLEDGEMENTS

The investigation was undertaken on behalf of the Commission of Inquiry into Rehabilitation of the Worked-out Phosphate Lands in Nauru. Fieldwork was facilitated by the Nauru Phosphate Corporation, and we thank David Newick (General Manager), Don Lauder (Chemist) and Stan Nowak (Surveyor) for their assistance. Drilling was undertaken by Jetstream Exploration Testing of Brisbane and arranged and supervised by Peter Barratt (Consulting Geologist). Tidal information was supplied by Prof. K. Wyrtki of the University of Hawaii, and rainfall information was supplied by the Australian Bureau of Meteorology and the Nauru Phosphate Corporation. We thank Tony Falkland (Australian Capital Territory Administration) and Fred Ghassemi (Australian National University) for discussion of aspects of island hydrology.
REFERENCES


GOODWIN, R.S., 1980 - Water assessment and development in Barbados.


<table>
<thead>
<tr>
<th>Bore</th>
<th>R.L. (m)</th>
<th>Total depth (m)</th>
<th>S.W.L. (m below ground)</th>
<th>Estimated freshwater layer thickness (m)</th>
<th>E.C Bottom of hole (microS/cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>P1</td>
<td>34.53</td>
<td>30</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>P2</td>
<td>26.56</td>
<td>70</td>
<td>-</td>
<td>7</td>
<td>19500</td>
</tr>
<tr>
<td>P3</td>
<td>12.74</td>
<td>54</td>
<td>11.23</td>
<td>3</td>
<td>33400</td>
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<tr>
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<td>30.89</td>
<td>43</td>
<td>-</td>
<td>7</td>
<td>6000</td>
</tr>
<tr>
<td>P5</td>
<td>27.27</td>
<td>50</td>
<td>-</td>
<td>0.5</td>
<td>22100</td>
</tr>
<tr>
<td>P6</td>
<td>27.18</td>
<td>70</td>
<td>-</td>
<td>7</td>
<td>43500</td>
</tr>
<tr>
<td>P7</td>
<td>28.98</td>
<td>83</td>
<td>-</td>
<td>4.5</td>
<td>39500</td>
</tr>
<tr>
<td>Q1</td>
<td>4.80</td>
<td>26</td>
<td>-</td>
<td>3.5</td>
<td>27000</td>
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<td>W1</td>
<td>25.32</td>
<td>32</td>
<td>-</td>
<td>6.5</td>
<td>1000</td>
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<td>W2</td>
<td>35.43</td>
<td>65</td>
<td>-</td>
<td>3.5</td>
<td>22000</td>
</tr>
<tr>
<td>W3</td>
<td>39.34</td>
<td>65</td>
<td>37.85</td>
<td>6.5</td>
<td>30300</td>
</tr>
<tr>
<td>W4</td>
<td>26.14</td>
<td>55</td>
<td>24.56</td>
<td>3.5</td>
<td>27000</td>
</tr>
</tbody>
</table>
### TABLE 2

**LIST OF GEOELECTRICAL SOUNDING LOCATIONS**

<table>
<thead>
<tr>
<th>Location</th>
<th>R.L. (m)*</th>
<th>AB/2 (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>DP1 Flat-bottomed depression, northern Nauru</td>
<td>12.8</td>
<td>150</td>
</tr>
<tr>
<td>DP2 Topside Sports Oval, mid-southern Nauru</td>
<td>26.4</td>
<td>180</td>
</tr>
<tr>
<td>DP3 Buada Lagoon</td>
<td>3.3</td>
<td>100</td>
</tr>
<tr>
<td>DP4 Buada Lagoon</td>
<td>4.0</td>
<td>100</td>
</tr>
<tr>
<td>DP5 North of Buada Lagoon</td>
<td>3.4</td>
<td>100</td>
</tr>
<tr>
<td>DP6 Stockpile area, central Nauru</td>
<td>28.0</td>
<td>400</td>
</tr>
<tr>
<td>DP7 Buada Lagoon</td>
<td>3.0</td>
<td>80</td>
</tr>
<tr>
<td>DP8 Radio transmitter, central Nauru</td>
<td>26.3</td>
<td>900</td>
</tr>
<tr>
<td>DP9 Anetan coastal strip, northern Nauru</td>
<td>3.9</td>
<td>100</td>
</tr>
<tr>
<td>DP10 Coastal strip, east Nauru</td>
<td>4.9</td>
<td>90</td>
</tr>
</tbody>
</table>

*Mean sea level = R.L. 1.34 m.*
<table>
<thead>
<tr>
<th>Depth Probe</th>
<th>Layer</th>
<th>Resistivity (ohm-m)</th>
<th>Thickness (m)</th>
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TABLE 4

MONTHLY RAINFALL AND POTENTIAL EVAPORATION (in mm)

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<td>August</td>
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* Estimated from Fleming's (1987) formula
### TABLE 5

CHEMICAL ANALYSES OF NAURU GROUNDWATERS

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<th>P6 25.5m</th>
<th>W1 25.5m</th>
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* field pH

Elemental/ionic analyses in mg/L
Electrical conductivity (E.C.) in microS/cm
FIGURE 1 (S8).
Geology, geoelectrical sounding & drillhole locations

FIGURE 2 (S8).
FIGURE 3 (S8).
Logs of drillholes

FIGURE 4 (S8).
Apparent resistivity and Model resistivity ($\Omega$ m)

Layer 1: 3.2 m, 270 $\Omega$ m
Layer 2: 7.0 m, 1100 $\Omega$ m
Layer 3: 15.0 m, 10000 $\Omega$ m
Layer 4: 6.0 m, 1000 $\Omega$ m
Layer 5: 20 m, 200 $\Omega$ m
Layer 6: 6 $\Omega$ m

Drillhole W1

Phosphate
Limestone
Water table
EC = 2300 $\mu$S/cm
Fresh
Brackish
Seawater

Resistivity layered model (DP2) compared with drill log (W1)

FIGURE 5 (S8).
Thickness of fresh water layer

FIGURE 6 (S8).
Cross section showing groundwater salinity (Electrical conductivity in $\mu$S/cm)

**FIGURE 7 (S8).**

Cross section showing groundwater salinity (Electrical conductivity in $\mu$S/cm)

**FIGURE 8 (S8).**
Nauru annual rainfall 1916–1986

FIGURE 9 (S8).

Nauru annual rainfall frequency distribution

FIGURE 10 (S8).
Nauru mean monthly rainfall and potential evaporation 1916–86

FIGURE 11 (S8).

Relationship between monthly rainfall and potential evapotranspiration derived for Tarawa (after Fleming, 1987)

FIGURE 12 (S8).
Reduced level
(m above mean)

Tidal height (Slater gauge at Bul Harbor)

High tide (1975)
RL 2.000

Highest daily msl (1979)
1.373

Highest monthly msl (1986)
E 1.352

Adopted mean sea level (1979)
Rl 1.000

Mean monthly msl (1974 - 1986)
E...

Lowest monthly msl (1983)
RL 1.000

Low tide (1975)
RL 0.166

Lowest daily msl (1983)
0.091

Lowest monthly msl (1983)
RL 1.000

Mean sea level at Nauru (after S. Nowak, NPC and K. Wyrski UH)

Range of daily msl

Range of monthly msl

FIGURE 13
(S8)
Surface catchment of Buade Lagoon

Nauru groundwater flow system

FIGURE 14 (S8).
Reversal of hydraulic gradient with tidal fluctuation

FIGURE 15 (S8).

Tidal fluctuations in observation bore P3, 800 m inland 18 October 1987

FIGURE 16 (S8).
Tidal efficiency and tidal lag in Nauru bores and wells

FIGURE 17 (S8).
Relationship of total dissolved solids to electrical conductivity

\[ y = 0.63x \quad r = 0.999 \]

Limit for drinking water

Seawater

FIGURE 18 (S8).
Well with high bacterial count (MPN ≥ 23)
• Cave with high bacterial count (MPN ≥ 23)
△ Well, low or nil bacterial count
460 Total dissolved solids (mg/L)

Nauru coastal plain groundwater quality

FIGURE 19 (S8).
Trilinear diagram showing ionic composition of Nauru groundwaters, in percent milliequivalents/litre

FIGURE 20 (S8).
PACIFIC OCEAN

FIGURE 21 (S8).

Coastal plain — septic tanks, animal wastes, underground petrol tanks
General direction of groundwater flow

Nauru groundwater pollution hazard

Slimes (cadmium) dump
Saltwater tanks
Dilatanks
Garbage dump
Nuatu International Airport

Buada Lagoon
Surface catchment of Buada Lagoon

166°55'00" 188°57'00"
0°30'30" 0°32'00" 0°33'30"
FIGURE 22 (S8).
Potassium (K) concentration in soil (%)
Uranium (U) concentration in soil (ppm)

Contour interval 20 ppm
Station number (ppm U)

FIGURE 24 (S8).
Thorium (Th) concentration in soil (ppm)

Contour interval 20 ppm

Station number

(ppm Th)

FIGURE 25 (S8).
APPENDIX

Apparent resistivity field plots and their interpretation,
Schlumberger geoelectrical soundings DP 1-10.

(i) Field data points are shown as crosses (+).
(ii) The continuous curve represents apparent resistivity computed for the interpreted layered resistivity model.
(iii) AB is the current electrode spacing (metres).
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DP 5
DP 6

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SECTION 2

SEDIMENTATION PATTERNS AND STRUCTURE OF TOFUA TROUGH - A FOREARC BASIN ON THE TONGA RIDGE

Introduction

Tofua Trough, located in the Kingdom of Tonga, is a linear sediment-filled depression within the Tonga Ridge between the chain of active volcanoes of the Tofua magmatic arc and the high-standing Tonga platform to the east (Figs 1 & 2). The structure is about 500 km long, extending from Ata Island in the south to Fonualei Island in the north. It is best developed, however, north of Tongatapu between latitudes 19°00'S and 20°30'S - to the west of the Ha'apai group of coral islands on the Tonga platform. Here the trough is 30-40 km wide, lies in a water depth of about 1800 m and contains about 2 km of sediments.

The Tonga arc-trench system, together with the Mariana arc-trench system, has been widely recognised as type example of a simple island arc with an active marginal basin (Karig, 1970; Packham, 1978; Hawkins et al., 1984). Tofua Trough forms an integral and distinct architectural element of the Tonga arc. For this reason, study of its structural and stratigraphic development in what is perceived as a relatively uncomplicated tectonic setting contributes to the general understanding of geotectonic processes and their interactions in modern island arc environments, and assists

As yet no significant mineral resource potential has been identified in the trough itself. The Tonga platform, adjacent to the trough, is considered prospective for petroleum and has been actively explored by oil companies (Maung et al., 1981). Hydrothermal mineral deposits associated with volcanic activity along the Tofua magmatic arc may have formed at depth within the Tofua Trough sediment pile (Cronan et al., 1984).

In 1986 a detailed geophysical survey of Tofua Trough and its margins was undertaken by HMNZS 'Tui' as part of the Tripartite II marine research program co-ordinated by CCOP/SOPAC (Suva). Geophysical systems operated included single-channel seismic with 120 cu inch airgun source, magnetics, gravity and 12-kHz bathymetry. Profiles of 500 km total length were recorded over a pattern of eleven survey lines (A-K) criss-crossing the trough as indicated in Figure 3. The results of the investigation are presented in this paper.

**Tectonic and Geological Framework**

The region lies at the extreme eastern boundary between the Indo-Australian and Pacific plates (Fig. 2). The convergent plate boundary is marked by the 10500 m deep, NNE trending Tonga Trench at which the Pacific plate is being subducted beneath the Tonga Ridge along a Wadati-Benioff zone dipping to the west at about 45° (Isacks et al., 1969; Hanus & Vanek, 1979; Hamburger & Isacks, 1987). The Tonga Ridge consists of two principal structural components, (i) the reef and coral island studded Tonga platform, and (ii) the
Pliocene(?)-Holocene, mainly andesitic volcanic cones of the Tofua magmatic arc located along the western margin of the ridge. The platform and volcanic chain are separated by Tofua Trough. Extending westward of the Tonga Ridge is the V-shaped, 2000-3000 m deep Lau Basin, which is bounded to the west by the meridional Lau Ridge.

The Lau Basin is characterized by relatively shallow depth (2000-3000 m compared with the 5000-6000 m deep Pacific Basin to the east), rough acoustic basement, thin sediment cover, and high but variable heatflow. On the basis of this, and the recovery by dredging of fresh tholeiitic, midocean ridge-type basalts (MORB), the Lau Basin has been interpreted as an actively spreading backarc basin separating the old forearc platform and active volcanoes of the Tonga arc from the remnant volcanic arc of the Lau Ridge (Karig, 1970; Sclater et al., 1972; Hawkins, 1974; Gill, 1976).

Various models for opening of the Lau Basin have been proposed (Sclater et al., 1972; Lawver et al., 1976; Weiszel, 1977; Falvey, 1978; Eguchi, 1984), or implied from maps of tectonic fabric presented (Cherkis, 1980; Larue et al., 1982; Malahoff et al., 1982). The models are mainly based on evidence from magnetic lineation patterns, and to less extent on seismicity, earthquake focal mechanisms, seafloor morphology, sediment thickness, age and petrology of seafloor samples, and palaeomagnetism. A major problem in elucidating the detailed tectonic evolution of the Lau Basin has been that the magnetic lineations which reflect crustal accretion and seafloor spreading processes have been difficult to map. This difficulty in identifying and correlating magnetic lineations suggests short spreading ridge segments and general structural complexity of the basin, and is responsible for the diversity in interpretations.
Proposed spreading geometries include, (i) a Y-shaped configuration of spreading ridges and transforms involving interaction of a number of micro-plates (Weissel 1977; Falvey, 1978), (ii) a curved spreading ridge disrupted only in a minor way by transform offsets (Cherkis, 1980), and (iii) a uniform system of NW-SE trending transforms and NE-SW spreading axes continuous along the length of the basin (Sclater et al., 1972; Eguchi, 1984). The spreading geometry depicted in Figure 2 is based on the seismotectonic interpretation of Eguchi (1984).

Identification of the oldest magnetic anomaly in the central Lau Basin (south of Peggy Ridge and west of Tofua Trough) has varied among investigators working on different magnetic data sets. The oldest anomaly according to Weissel (1972) is 2A, according to Cherkis (1980) it is 3A, while Larue & others (1982) map it as 3. Thus from magnetic data alone, onset of seafloor spreading is the central Lau Basin is placed at 3.5-5 Ma with opening proceeding at a total rate of 50-76 mm/year. Weissel (1977) suggests, however, that a deep 80 km-wide subsidiary basin adjacent to the Lau Ridge may have been the site of initial separation of the Lau and Tonga Ridges at 5-6 Ma, but that it was abandoned by a ridge crest jump to the east (at about 3.5 Ma).

The Lau Basin is geochemically zoned (Hawkins & Melchoir, 1985), with normal type (N-MORB) basalts forming the modern central part of the basin and transitional type basalts (similar to those of the Mariana Trough) underlying the basin's older western margin. The eastern side of the basin close to the active Tofua magmatic arc is also considered to be floored by transitional type basalts (Falloon et al., 1987), but this remains untested due to burial by products of
the recent volcanism.

Forearc sediments on the Tonga Ridge are up to 5 km thick and consist of a near continuous sequence of middle Eocene-Holocene volcaniclastics and carbonates overlying igneous basement of pre-middle Eocene age (Cunningham & Ascombe, 1985). The sediment pile thins towards the eastern rim of the platform; Oligocene beds are thin or absent. Basement is represented by exposures of a basalt-rhyolite suite of island arc tholeiitic affinity on the island of Eua.

The geological evolution of the Tonga platform has been described in recent accounts (Maung et al., 1981; Kroenke, 1984; Scholl et al., 1985b; Exon et al., 1985; Herzer & Exon, 1985; Lewis, 1985). Tonga platform basement is believed to have originally formed part of an Eocene forearc adjacent to New Caledonia on the ancestral Inner Melanesian Arc. The Oligocene geological record shows that this period was one of limestone deposition and quiescence in volcanic activity, with widespread uplift and emergence and generally thin sediment accumulation. Opening of the South Fiji Basin in the late Oligocene rafted the 'Eua arc fragment (Lau-Tonga ridge) eastward. Arc volcanism began by early Middle Miocene with volcanic centres located on the western side of the Lau-Tonga ridge. At this time the Tonga platform was a forearc basin resembling the present one, with volcaniclastic deposits thickly accumulating near the volcanic arc and thinning eastwards towards an outer high. Intrusive activity was common during this period.

A decrease in volcanic activity is indicated in the late Miocene. However, evidence from wells drilled on Tongatapu suggests that volcaniclastic sedimentation continued well into early Pliocene.
before a change to limestone deposition in late Pliocene-Holocene, though there is uncertainty as to how much of the volcaniclastic sequence may represent older, reworked material. As the Lau Basin opened in early Pliocene, the forearc (now the Tonga platform) was rafted away to the east resulting in extinction of the volcanic arc left behind on the Lau Ridge. Volcanism reappeared in the late Pliocene at the western margin of the Tonga platform with construction of the volcanic chain of the Tofua magmatic arc along which activity continues to the present time.

During Pliocene-Holocene, thermal uplift and subsidence associated with rifting and spreading in the Lau Basin has been complicated by tectonic adjustments to oblique subduction of the Louisville Ridge beneath the platform (Dupont & Herzer, 1985). Uplifts, both regional and local, have produced a widespread unconformity in the early Pliocene. Two principal sets of normal faults have developed since late Pliocene, (i) a westerly dipping set related mainly to Lau Basin rifting and subsidence, and (ii) a set cutting orthogonally across the NNE trending axis of the forearc basin probably activated in response to Louisville Ridge subduction.

The Sediments and Arc Volcanism

Seafloor sediment samples were recovered by coring and dredging from stations distributed over an extensive area of Tofua Trough during the 1981 and 1982 research cruises of R/V 'Tangaroa' (Cronan et al., 1984). Sediments were generally found to comprise greater than 75% volcanic debris - fine sand to silt size, composed of acidic glass shards, calcic plagioclase and clinopyroxene with minor amounts of opaques. Rapid accumulation rates of volcanic material was indicated
by the low concentration of biogenic components. The seafloor
sampling results are consistent with the numerous charted sediment
The annotations indicate a bottom predominantly of "volcanic sand"
with some areas of "cinders" and "pumice", and only a few locations
marked as "coral". The volcanic sediments are undoubtedly sourced by
volcanic activity along the adjacent Tofua magmatic arc - both because
of the proximity of the arc and the silicic nature of the sediments.
Coral fragments were recovered from the middle of the trough (water
depth about 1700 m) as well as along its margins. This material was
probably transported by sediment gravity flows from the shallow waters
of the Tonga platform and from reefs developed along the arc - reefs
that have formed either as local narrow fringing reefs around
subaerial volcanoes or were constructed on shallow volcanic banks.

Some insight into the sedimentology of Tofua Trough may be
obtained by briefly reviewing what is known of the geology and
volcanic processes taking place along the Tofua volcanic arc, since it
is the arc that is the primary source of sediment. Bryan et al.
(1972) estimate that volcanic eruptions along the arc have averaged
about one every four years during this century, discounting small
submarine eruptions that may have passed unnoticed or unrecorded.
Volcanic events other than mild solfataric activity appear to have a
periodicity of 20-50 years, and most eruptions are of short duration,
lasting from a few months to a year. Existing records do not reveal
any close relationship in periods of activity between adjacent
volcanic centres. Geological and/or petrological studies of most of
the volcanic islands and shoals adjacent to our area of investigation
have been undertaken; those studied include (from north to south)
Late Island, Home Reef, Metis Shoal, Kao, Tofua, Falcon, Hunga Tonga
and Hunga Ha'apai Islands. A summary of the more important aspects of the islands' geology and volcanic activity relevant to sedimentation in the adjoining Tofua Trough is provided below. For more detailed discussions the reader is referred to Schofield (1967), Bauer (1970), Melson et al. (1970), Baker et al. (1971), Mulder & Nieuwenhuiizen (1971), Bryan et al. (1972), Ewart et al. (1977), and Cunningham & Anscombe (1985).

**Late**: The island is a conical volcanic peak about 540 m high and 6 km in diameter at sea level, with a well-developed central crater at its summit. The volcano last erupted in 1854 and still shows some weak solfataric activity in the crater. A number of small cinder cones are located on the western flank of the volcano. The extensive lowland areas to the north and east of the central cone are composed of rough, blocky lava flows mantled by ash deposits. The lavas on Late are all of basaltic andesite or andesite.

**Home Reef**: Very little is known about this structure. British Admiralty Chart 2421 indicates that the seafloor on the flanks of the seamount is composed mainly of patches of volcanic sands and pumice; coralline sediment was recorded at several sites on the southeast flank. A reported submarine eruption in 1984 is located about 6 km southeast of the reef.

**Metis Shoal**: The shoal represents a submarine volcano that last erupted in 1967-68 with explosions of steam, ash and bombs. An island at least 0.5 km long and 24 m high was produced, but was soon eroded to beneath wave base. The erupted product consisted of vesicular rhyolite glass.
Kao: Kao is the highest volcano of the Tofua chain, and consists of a steep-sided symmetrical strato-volcano rising to an elevation of about 1125 m. The remnants of a crater or coalescing craters can be seen near the summit. The slopes of Kao are mantled by ash, cinder and flows - all apparently of recent origin since erosion has not incised deeply into the flanks of the volcano. No historic eruptions are known on Kao. The exposed lava flows are mainly augite andesite.

Tofua: Tofua is an active composite volcano 8-10 km across, with a large (4 km diameter) central collapsed caldera containing a fresh-water lake and recent volcanic deposits (tuff, lapilli tuff, cinder and some lava flows). Lofia cone within the caldera last erupted in 1959.

Four stratigraphic units have been mapped - Hamatua formation of pre-caldera age (?late Pleistocene), Hokula froth lava - possibly contemporaneous with the caldera collapse, post-caldera Kolo formation, and Lofia formation comprising deposits recently erupted from within the caldera. Erosional unconformities separate the Hokula froth lava from the underlying Hamatua formation and overlying Kolo formation. The Hamatua formation consists of lava flows of basaltic andesite, hyperthene-bearing augite andesite, andesite and augite dacite. It is thought to have been part of an older cone, possibly resembling Kao, which was largely destroyed as the caldera was created. A 500 m thick section of the Hamatua formation is exposed in the walls of the caldera, where individual flow units are seen to average 20 m in thickness.

The Hokula froth lava comprises a number of microvesiculated flow units of andesitic composition, each unit being about 5-6 m
thick. The formation is exposed in sea cliffs along the north-west coast of Tofua. The post-caldera Kolo formation is composed of air-laid lapilli tuff-breccia, tuff, unconsolidated ash and cinder and andesite lava flows up to 20 m thick. A consolidated lapilli tuff-breccia member lies unconformably on the Hamatua formation on the southern, eastern and northern flanks of Tofua - forming a blanket more than 30 m thick over the entire area. The recently-deposited Lofia formation is composed of lithologies similar to that of the Kolo formation, but its distribution is limited essentially to the confines of the caldera.

Falcon Island (also called Fonua Foou): Falcon Island is the top of an active volcano that, like Metis Shoal, is successively built up above sea level and then eroded back down. It has had two major periods of activity during historic times, the first between 1877-1894 and the second between 1921-1936. Such long periods of sustained activity are unusual for volcanoes of the Tofua chain. An augite andesite was erupted in 1885 and in 1927 andesitic pyroclastic debris was ejected.

Hunga Tonga & Hunga Ha'apai: These islands appear to be the northern and western subaerial erosional remnants of a once much larger active volcanic cone. Submarine eruptions were recorded in 1912 and 1937 at a site several kilometres to the south of Hunga Tonga. The geology of the islands consists of alternating layers of andesitic lava flows and pyroclastics (scoria, lapilli and ash), which dip gently away from the centre of the primary founding volcanic edifice. Much of the surface of the islands is covered by a reddish volcanic ash.
Previous Geophysical Investigations

Early work in the area included that of the Capricorn Expedition (1952-53) during which a deep seismic refraction sounding was made along Tofua Trough near Kao Island; two similar soundings were made farther to the east in the Tonga Trench (Raitt et al., 1955). In 1972, the Mobil Oil research ship 'Fred H. Moore' recorded a number of multichannel seismic lines across Tonga Ridge - an interpretation of two of the lines (72-200 and 72-202) has been published (Kroenke & Tongilava, 1975). Between 1975-1978, ORSTOM (Office de la Recherche Scientifique et Technique Outre-Mer, Noumea) surveyed Tonga Ridge with an extensive network of single-channel seismic, magnetic and gravity traverses during the AUSTRADEC IV, EVA III and EVA VII research programs. During the EVA VII program, seven seismic refraction profiles were recorded along an east-west transect across Tonga Ridge at about 11°30'S (Pontoise et al., 1980; Pontoise & Latham, 1982), just to the north of the Capricorn Expedition refraction transect.

Petroleum exploration began in Tonga in 1970 following the discovery of thermogenic crude oil seepages on Tongatapu and Eua. Much of the Tonga platform has been mapped by oil company seismic lines, and five exploration wells have been drilled on Tongatapu (Tongilava & Kroenke, 1975; Maung et al., 1981). The southern Tonga platform was investigated in 1982 by R.V. 'S.P. Lee' (Scholl & Vallier, 1985) using multichannel seismic, gravity, magnetics, sonobuoy seismic refraction and seafloor sampling methods. During this cruise single-channel seismic and several sonobuoy refraction profiles were recorded along the axis of Tofua Trough (Fig. 3) while the ship was in transit to the main study area south of Tongatapu.
The M.S. 'Natsuchima', in 1984, completed a geophysical and geological research study of the Tonga Ridge north of Vava'u (Honza, Lewis et al., 1985). Though the area investigated lies just beyond the northern extent of the Tofua Trough, the results of this cruise are useful for comparing the geology of sections of the Tonga Ridge with and without a forearc Tofua Trough.

**Bathymetry**

The bathymetry of Tofua Trough and adjacent areas shown at 200 m contour interval (Fig. 3) is based largely on the compilation by Chase et al. (1982) modified by the new bathymetric control provided by the HMNZS 'Tui' survey. Additional useful sources of data were the 1:200 000 chart by Eade (1972) and British Admiralty chart 2421 'Tonga or Friendly Islands' printed in 1984.

The deepest part of Tofua Trough lies between 19°05' - 20°05'S and is outlined by a closed 1800 m isobath. Within this area the seafloor is fairly flat, sloping down only very gently to the middle of the trough. Maximum depths attained are 1840 m east of Tofua Island and 1870 m southeast of Metis Shoal. The Tonga platform, which bounds the trough to the east, is generally shallower than 500 m. The trough/platform boundary is formed by a WNW-facing submarine escarpment with 1200-1600 m relief and average slope of 8°-18°. Numerous submarine and subaerial edifices of the NNE trending Tofua volcanic chain (including Home Reef, Metis Shoal, Kao, Tofua, Falcon, Hunga Tonga and Hunga Ha'apai Islands) separate Tofua Trough from the 2200+ m deep Lau Basin to the west. The submarine slopes on the volcanic edifices are moderately steep with gradients mainly in the range 6°-15°. Because the volcanic chain is not
strictly linear but consists of a band of edifices 20-30 km wide, the western margin of the trough is not clearly defined. The trough appears to become narrower towards its southern end. This is probably partly due to the volcanic chain being more diffuse in this region, with some edifices located within the trough itself.

**Seismic, Gravity and Magnetic Survey Data**

Line drawing interpretations of the seismic reflection records for Lines A-C, D-F, G-H and I-K are shown in Figures 4, 5, 6 and 7 (respectively). Free-air gravity and magnetic anomaly profiles for these lines are presented above corresponding seismic sections. The magnetic anomaly data are relative to IGRF80 (Peddie, 1982).

**Seismic Results**

The seismic sections across Tofua Trough (Figs 4-7) show the sediment fill in the trough to be highly stratified, with beds generally sub-horizontal or dipping at shallow angles. Structural disruption of the strata is evident, and varies over the trough from very minor to locally severe. The observed structures are primarily normal faults and piercements, produced by tectonism and volcanic activity. Basement is not clearly defined in our data, but a sedimentary section of at least 1.8 s (twt) - corresponding to a thickness of about 1800 m (assuming a velocity of 2.0 km/s) is indicated. This is in accord with interpretations of EVA729 and AUS401 seismic profiles over the deepest part of the trough north-east of Kao/Tofua Island (Dupont, 1982), showing about 2.2 s (twt) of sediment fill. Refraction data from the R.V. 'S.P. Lee' cruise (Childs, 1985; Childs, pers. comm.) - sonobuoys S2, S3 & S4 (Fig. 3) -
indicate low velocity sub-bottom layers of velocity 1.6-2.1 km/s and combined thickness 2.05 km overlying a 4.1 km/s substratum of presumed basement, thus confirming that the sediment fill in the trough is about 2.0 km thick.

The strata in the trough appear, from the seismic, to consist of depositional units with thickness on average of 25-35 m. The thickness of these primary beds is difficult to estimate with accuracy because of resolution limits imposed by the frequency content of the seismic signal, and because some apparent stratification may be due to effects of airgun bubble-pulse oscillation as well as inter-layer reverberation. Each of the observed layers may represent a deposition cycle associated with a major eruptive episode from a volcanic centre on an adjacent section of the chain. The beds lap onto the downfaulted Tonga platform which forms the eastern boundary of the trough, and generally inter-finger with the more chaotic proximal units of the volcanic constructions which are part of the volcanic arc, forming the trough's western boundary.

The beds dip gently from the Tonga platform boundary, and also generally from the volcanic arc side. On the arc side the beds are commonly truncated by normal faults. Some apparent truncation may be due obliteration of the underlying section by thick lava flows emanating from the nearby volcanic centres. Where such truncation occurs, the deeper beds often show a continuing westward dip (eg Line E, Fig. 5; Line K, Fig. 7). Discounting local structural variations, the overall pattern seen in the trough is an intrabasinal sag, most pronounced in the deeper strata and almost imperceptible in the near surface beds which lie close to horizontal - at least where the trough is relatively wide. This sediment sag is well developed along Line B
(Figs 4 & 8) and Line C (Fig. 4), and has a magnitude of about 300 m
in the deeper part of the section.

The progressive down-bowing of the sediment pile appears to
be an accommodation response to excess space being generated near the
base of the accumulating pile. Mechanisms producing this extra volume
include, (1) compaction by sediment loading, particularly of relatively
porous strata such as ash beds, and (2) trough dilation through - (1)
subsidence of the trough floor, perhaps by continued down-faulting
along the eastern boundary fault system - partly as an isostatic
response to loading of the crust to the west by the arc volcanoes and
their volcanogenic sediment aprons, and/or (ii) tectonic widening of
the trough in an intra-island arc tectonic stress field acting
transverse to the arc.

The middle and eastern half (Tonga platform side) of the Trough
is characterized by closely-spaced sub-parallel reflectors with high
continuity of up to several tens of kilometers, and sequences which
show low-angle onlap at their base. Such seismic patterns are best
seen in the EVA729 (Dupont, 1982 - Fig. V-11) and 1982 'S.P. Lee'
reflection profiles which were shot along the trough axis, but are
also well illustrated in the seismic profile of Line B (Figs 4 & 8).
Such patterns are typical of onlapping-fill seismic facies (Sangree
& Widmier, 1977) deposited in basin slope and floor environments.
In Tofua Trough they probably consist predominantly of volcaniclastic
deposits transported down from the island-arc volcanoes by low-velocity
turbidity currents. Carbonate debris from reef buildups and to minor
extent, pelagic deposits, are also likely to be carried into the
trough at less frequent intervals by sediment gravity flow from high
on the Tonga platform as well as from parts of submarine volcanic
edifices within the photic zone where coral reefs have developed. The trough deposits are expected to contain interbeds of hemipelagic sediment and ash fallout. Though sediment grain size in an onlapping-fill seismic facies is predominantly fine (silt - fine sand in Tofua Trough, judging from the seafloor sampling results), some thin beds of coarser volcaniclastic sand borne by intermittent high-energy turbidity currents are likely to be present, particularly on the arc side of the trough.

Where structural modification has not severely disrupted the strata, some beds of the Tofua Trough can be traced beneath the lower flanks of volcanic edifices (Line C, Fig. 4; Line E, Fig. 5) and can be seen to thin and converge upwards. These up-slope depositional units may be proximal volcaniclastic turbidites, or alternatively may represent debris flows or pyroclastic flows that have been partly converted by surface transformation to turbidites (Fisher, 1984) and subsequently deposited more distally on the floor of the trough.

Tofua Trough reflectors can also be traced beneath the lower flank of the Home Reef submarine volcano (Lines A & B, Fig. 4 & Fig. 8). This seismic transparency suggests that the flanks are composed of volcaniclastic sediments and are largely devoid of massive lava flows. The beds within the edifice and beneath it, to a level about 300 m below the floor of the adjacent Tofua Trough, appear highly deformed. The volcanic sediments are probably of relatively low strength and may have been affected by slumping at or soon after deposition - or been deformed at a later stage by differential loading as the volcanic pile grew upwards with continuing eruptions. The hummocky surface topography on the lower flank of the Home Reef volcano may be caused by slumping of recently deposited
volcaniclastics - the volcano still being quite active, the last eruption (submarine) having been recorded in 1984.

There is further evidence for extensive and large scale mass wasting on the submarine flanks of the volcanic edifices. The lower, trough-facing flank of the large submarine volcano just north of Kao has similar irregular topography as seen on the Home Reef volcano. The seismic profile (line C, Fig. 4) shows that the topographic relief is due to the slumping and sliding of several sediment sheets, 50 m thick and up to 4 km long, down the side of the edifice. Line D (Fig. 5) also shows a large hummocky deposit of slumped material at the base of the slope. Slightly older, lensoid slump deposits can be seen in the section at this location. On the Lau Basin-facing lower flank of this volcano, chaotic bedding extending deep into the section suggests that slumping has been active here for some time. Other examples of chaotic seismic expression signifying probable slump deposits can be seen in the sections at the base (Lau Basin side) of submarine volcanoes crossed by lines I and J (Fig. 7).

The hummocky build-up seen on Line H (Fig. 6), at the base of the Tonga platform escarpment may represent material that has slumped from high on the platform, but may also consist of lavas or pyroclastics that have been extruded out onto the trough floor from an underlying volcanic intrusion for which there is evidence in the section. Volcanic structures can also be seen within the Tofua Trough section on Lines B (Figs 4 & 8) and F (Fig. 5).

Only limited seismic coverage of the Tonga platform was obtained during our survey; parts of Lines A, D and E (Fig. 3) extend over its western margin. Seismic detail below 0.5-1 s (twt)
sub-bottom has been lost because of the strong water-bottom multiple. The seismic profiles (Figs 4 & 5) show a pinnacled seafloor on the platform, probably representing reef constructions, below which lies about 100-200 m of recent sediments. This ?late Pliocene-Holocene unit is underlain by an erosional unconformity that dips gently to the west, and truncates block-faulted older rocks of the Tonga platform succession (Eocene-early Pliocene).

Gravity and Magnetic Expression

The free-air gravity field over Tofua Trough and adjacent marginal zones is highly anomalous (Figs 4-7), varying in amplitude from about 50 mgals to 150 mgals. The variations in gravity can be attributed largely to seafloor topography and partly to lateral density variations in the sub-bottom, while the positive gravity bias is a reflection of the major positive gravity anomaly normally present along the crest of active island arcs (Talwani, 1970).

The broad gravity low observed over Tofua Trough is due to the combined effect of deeper water in the trough and the thick sequence of young, relatively low density volcanogenic/hemipelagic sediment that has accumulated in the trough. This is confirmed by 2-dimensional gravity modelling (Fig. 8) along Line D, which transects the central part of the trough almost orthogonally and also crosses the adjacent volcanic chain to the west and the Tonga platform on the eastern side of the trough. Following Misseque & Malahoff (1982), a density of 2.0 \( \text{tm}^{-3} \) was adopted for the arc-derived volcanioclastic sediment pile within the trough and that deposited as a sediment wedge on the Lau Basin side of the volcanic chain. Similarly, 2.4 \( \text{tm}^{-3} \) was taken
as the density of the upper section of the Tonga platform, the 
basement beneath the trough and also the submarine volcano just to the 
north of Kao Island. A good match between observed and calculated 
gravity was obtained for a modelled sediment thickness of 2.0 km in 
the trough - a result which supports the seismic refraction and 
reflection evidence.

The magnetic field is also highly anomalous over the Tofua 
Trough region. Magnetic variations of up to about 1200 nT were 
recorded, though anomaly amplitudes are more commonly in the order of 
several hundred nT. The larger anomalies are mainly associated with 
submarine and subaerial volcanoes of the Tofua volcanic chain.

The limited magnetic coverage of the Tonga platform adjacent to 
Tofua Trough shows a fairly subdued field with anomalies mainly 
significantly less than 200 nT in amplitude. The field is 
predominantly of moderate wavelength and there is no appreciable 
expression of seafloor topography in the profiles. Even the major 
trough/platform boundary faults are not strongly expressed in the 
magnetics, though there is some indication of a possibly related general 
increase in field strength across the boundary from west to east. 
Interpretation of the magnetics suggests that, (i) (?Eocene) magnetic 
basement is weakly magnetized, or is located at depth (say >1-2 km) 
beneath the platform in the area investigated, (ii) much of the magnetic 
variation is due to sources at intermediate depth within the section, 
probably the Miocene volcanioclastics and intrusions, and (iii) units at 
shallow depth (approximately 0-150 m) beneath the platform are 
non-magnetic or only weakly magnetic and probably comprise 
volcanipelagic sediments and reefal carbonates.
The volcanoes of the Tofua chain are mainly andesitic in composition and are considered to be composite structures built up by a combination of lava flows and pyroclastic deposits, with minor intrusions (mainly dykes) emplaced as feeders for the extrusives. Because of the high proportion of pyroclastic material, both induced and remanant magnetization are expected to be significant potential contributors to the overall magnetic field produced by these structures. This contrasts with the magnetization of mid-ocean basaltic volcanoes for which remanent magnetization is generally by far the most dominant component on account of the higher lava:clastic ratio.

The geomagnetic field in the Tofua Trough area presently has an inclination of $-40^\circ$ and declination $13.5^\circ$E (Fabiano et al., 1983), an orientation not appreciably different to that of a geocentric dipole field which at latitude $20^\circ$S would have an inclination of $-36^\circ$. Thus a young submarine volcanic cone in the Tofua Trough area would be expected to produce a bipolar magnetic anomaly pattern, aligned N-S with the positive anomaly to the north, at sea level directly above the cone. This assumes normal polarity of the earth's field during construction of the volcano. A different, more complex magnetic anomaly pattern is likely if the earth's field was of reverse polarity at the time, due to a mixture of opposing induced and remanant magnetization directions within the interior of the cone.

The four submarine volcanoes crossed on Lines G, I, J and K (Fig. 3) show strong magnetic anomalies consistent with normal magnetization, suggesting that these volcanoes formed in the past 0.7 Ma during the Brunhes normal chron rather than during an earlier period of reversed polarity. Palaeomagnetic measurements made on lavas from Tofua and the island of Hunga Ha'apai (Tarling, 1965) indicate that
the remanent magnetization coincides fairly closely with the direction of the present earth's field, implying a recent age. A small discrepancy is attributed to recent secular variation. The subbottom volcanic structure within the Tofua Trough sediments at the NWW end of Line B also appears to be normally magnetized, though the local field is disturbed to some extent by nearby volcanic edifices of Home Reef and Metis Shoal. The magnetic anomaly associated with the large unnamed seamount located just north of Kao Island (and crossed by Lines C and D) is of complex shape. Based on the morphology of the seamount, the anomaly is not compatible with a 100% normal magnetization. This infers some internal reverse magnetization, implying that parts of this edifice may be older than 0.7 Ma.

There is a possibility of interference between magnetic anomalies produced by andesitic volcanics of the Tofua chain and magnetic lineations resulting from Lau Basin seafloor spreading. Cherkis (1980) shows magnetic anomaly 3A coinciding with Tofua Trough in this area, while Larue et al. (1982) plot magnetic anomaly 3 in the same position. The observed magnetic field correlates very poorly with theoretical seafloor spreading anomaly profiles calculated assuming a fairly simple spreading geometry; the conflicting interpretations being symptomatic of this.

Structure

The principal geological structures mapped in the Tofua Trough area are shown in Figure 10. Normal faults and volcanic build-ups are the main structural elements. The NNE trending system of normal faults which marks the western limit of the Tonga platform and controls the eastern extent of Tofua Trough is a prominent structural
feature of regional scale. The Tonga platform/Tofua Trough boundary slopes WNW at about 20°-25° and is formed by a series of normal faults with dips of 45°-60° which step the Tonga Platform block down to depths in excess of 1.8 km below seafloor in the trough, making the combined throw greater than 3.4 km. The faults are segmented by poorly defined WNW and NW striking faults which cut across the Tonga platform, and were probably produced by differential uplift of the forearc as it swept over subducting ridges and seamounts of the Louisville Ridge.

The seismic sections show steeply dipping normal faults in places offsetting Tofua Trough sediments in the middle of the trough, particularly along the central part of the trough in the vicinity of the large Kao and Tofua volcanoes (Figs 5 & 6). They are common adjacent to the volcanic chain, both on the Lau Basin side as well as in Tofua Trough. Some appear to be growth faults produced as volcaniclastics are shed off the arc, while others are associated with volcanic diapirism.

Volcanic centres - subaerial, submarine and sub-bottom - are distributed along a 40 km wide band trending NNE (Fig.10). They extend from close to the Tofua Trough/Tonga platform boundary fault(s) to about 15 km west of the main line of subaerial volcanoes. The locations of the volcanic centres are probably controlled by NNE striking basement faults, though there is no compelling evidence for this in our data.

**Refraction/Reflection Data and Crustal Structure**

Sonobuoy refraction data are available which allow Tofua Trough
to be examined in the broader context of the overall crustal structure of the Tonga island arc. The data sources, to which reference has already been made, include the Capricorn Expedition (Raitt et al., 1955), the French EVA VII program (Pontoise et al., 1980; Pontoise & Latham, 1982), and the more recent Tripartite I 'S.P. Lee' cruise (Childs, 1985). The sonobuoy sounding locations are distributed latitudinally across the Tonga arc just to the north of Kao/Tofua Islands. The crustal cross-section of Figure 11 is oriented normal to the arc and located roughly along an extended Line D (Fig. 3). The bathymetry beyond our data from Line D has been taken from Chase et al. (1982), while the seismic velocity structure is based on the sonobuoy data projected onto the cross-section.

Revised interpretations of the 'S.P. Lee' Tofua Trough sonobuoy soundings (S2, S3 & S4 - Fig. 3) have been provided by Jon Childs (pers. comm.). These new interpretations give the sub-bottom velocity structure as:

<table>
<thead>
<tr>
<th>Layer</th>
<th>Velocity (km/s)</th>
<th>Thickness (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>S2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>1.6</td>
<td>1.08</td>
</tr>
<tr>
<td>2</td>
<td>2.9</td>
<td>0.97</td>
</tr>
<tr>
<td>3</td>
<td>4.1</td>
<td></td>
</tr>
<tr>
<td>S3</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>1.6</td>
<td>1.07</td>
</tr>
<tr>
<td>2</td>
<td>2.9</td>
<td></td>
</tr>
<tr>
<td>S4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>1.6</td>
<td>1.08</td>
</tr>
<tr>
<td>2</td>
<td>2.4</td>
<td></td>
</tr>
</tbody>
</table>

The accuracy of the S2 interpretation for the deeper layers may be
affected, however, by the presence of an underlying buried volcanic cone or intrusion seen in the seismic section of line B (Fig. 8).

The crustal velocity structure as portrayed in Figure 11 shows a marked discontinuity across the major boundary faults at the western margin of the Tonga platform. The data suggest that the volcanic chain and Tofua Trough are founded on transitional oceanic crust of the Lau Basin. This hypothesis is consistent with interpretations of Lau Basin seafloor spreading anomalies (Cherkis, 1980; Larue & others, 1982), but conflicts with the observations of Kroenke & Tongilava (1975) who believe that they can trace middle Eocene-lower Miocene strata of the Tonga platform beneath Tofua Trough on two oil company seismic lines crossing the forearc just north and south of Tongatapu. North of Vava'u the volcanic arc is located at the crest of the Tonga Ridge, there being no structural equivalent to Tofua Trough. The arc is, however, adjoined on both sides by a thick sequence of young volcaniclastic sediments within an ?isostatically downwarped basin (Kitekei'aho et al., 1985). This sediment pile is up to 2.0 s (twt) thick - much the same thickness as deposits along the volcanic chain adjacent to Tofua Trough. On the northern Tonga Ridge, the sediments are unconformably underlain by upper Miocene and older rocks similar to those beneath the Tonga platform. In their interpretation of 'S.P. Lee' multi-channel seismic profiles across the southern Tonga platform, Herzer & Exon (1985) show up to 1.6 s (twt) of sediment in the southern extension of Tofua Trough. Their reflector horizon A, representing the latest Miocene/early Pliocene unconformity, is mapped within the sedimentary section. The seismic basement is not identified, though Scholl et al. (1985a) interpret it equivocally as 'late Cenozoic arc basement' in their Figure 3.

Clearly, additional deep seismic refraction and reflection
investigations of Tofua Trough are required to establish more convincingly whether the young, mainly volcanogenic sediments in the trough are underlain by a down-faulted Tonga platform type section (essentially Eocene-Miocene) or by back-arc basalts of the Lau Basin. The sediments in the trough may be floored by both types of 'basement', each underlying different parts of the trough - as the data seem to suggest.

The crustal structure shown in Figure 11 suggests that Tofua Trough developed by subsidence of the trough/volcanic arc area relative to the Tonga platform by normal faulting at the major boundary fault system, in a manner analogous to half-graben formation. The mechanisms by which this may have occurred (and probably continues at present) include, (i) isostatic adjustment of the crust due to additional loading imposed by the build-up of arc-derived volcanic products, and (ii) thermal subsidence of the crust on cooling as the Lau Basin seafloor spreading axis moves progressively farther relatively westward. Tofua Trough appears to be a subsiding rift basin, contained on its western side by the growing volcanic build-ups of the Tofua volcanic chain.

Acknowledgements

We thank Jon Childs of the U.S. Geological Survey (Menlo Park, California) for providing revised interpretations of the 1982 R.V. 'S.P. Lee' sonobuoy refraction data.

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Figure 1. Location map. Bathymetry after Kroenke et al. (1983); isobaths in metres.

Figure 2. Regional tectonic setting and major structural elements. Lau Basin seafloor spreading geometry after Eguchi (1984) and Lewis (1985). Small circles along the Tofua magmatic arc indicate major Pliocene and Holocene volcanic centres.

Figure 3. Tofua Trough geophysical survey lines and bathymetry.

Figure 4. Seismic line drawing interpretations, free-air gravity profiles and magnetic anomaly profiles - Lines A, B and C.
Figure 5. Seismic line drawing interpretations, free-air gravity profiles and magnetic anomaly profiles - Lines D, E and F.

Figure 6. Seismic line drawing interpretations, free-air gravity profiles and magnetic anomaly profiles - Lines G and H.

Figure 7. Seismic line drawing interpretations, free-air gravity profiles and magnetic anomaly profiles - Lines I, J and K.

Figure 8. Seismic profile across Tofua Trough (Line B) showing thick well-stratified sediment fill (greater than 1.6 sec twt) and volcanic intrusion or buried volcano within the section.

Figure 9. Tofua Trough gravity model (Line D). The gravity profile calculated for the interpreted 2-dimensional model is compared to the observed data.

Figure 10. Map of principal structural features. The encircled V's correspond to Pliocene - Recent volcanic centres. Major faults are represented by heavy lines ticked on the downthrown side; inferred faults are shown as broken lines. Bathymetric contours are at 1000 m contour interval.

Figure 11. Crustal cross-section through the Tonga arc, based on seismic refraction and reflection data. Broad black arrows show direction of relative plate
motion - (i) convergence by subduction of the Pacific plate at the Tonga Trench on the right of the figure, and (ii) opening of the Lau Basin on the left.
FIGURE 1 (S9).
Regional tectonic setting & major structural elements

FIGURE 2 (S9).
Seismic, gravity & magnetic profiles

Lines A, B & C

FIGURE 4 (S9).
Seismic, gravity & magnetic profiles

Lines D, E & F

FIGURE 5 (S9).
Seismic, gravity
& magnetic
profiles

Lines G & H

FIGURE 6 (S9).
Seismic, gravity & magnetic profiles

Lines I, J & K

FIGURE 7 (S9)
FIGURE 9 (S9).

Seamount north of Kao I.

Tonga Platform

Lau Basin

Tofua Trough

Calculated

Observed

Gravity Model

V/H = 6.9

Horizontal Scale
0 10 20 km

Gravity (mgal)

FA

0 80 120 160

0 20 80 120

Sea-level
FIGURE 10 (S9).
Mantle partial melts from melting of subducted lithosphere.

Active volcanic chain.

Magmas derived from melting of subducted lithosphere.

Tonga Arc crustal cross-section

FIGURE 11 (S9).
SECTION 10

The Structurally Complex Eastern Margin of Manihiki Plateau - SeaMARC II, Geophysics and Geological Interpretation

Introduction

Manihiki Plateau is a large, anomalously shallow area of seafloor, about 500,000 km² in size, located in the central Pacific Ocean (Fig. 1). It is one of a number of such features in the Pacific that total about 2% of the ocean basin - other similar plateaus being Ontong Java Plateau, Shatsky, Hess and Magellan Rises. The most elevated part of the plateau, the south-east sector, is known as the Manihiki High Plateau. It rises from adjacent ocean basins, which are about 5500 m deep, to a relatively smooth, dome-shaped upper surface lying mainly 2400-3000 m below sea-level. The Manihiki High Plateau is bounded by the Danger Islands and Suwarrow Troughs to the west, Samoan Basin to the south, Central Pacific Basin to the north and Penrhyn Basin to the east. The margins of Manihiki Plateau all appear to be structurally controlled with structural development most spectacularly expressed along the eastern margin of the plateau, where the High Plateau plunges 3500 m into the depths of the Penrhyn Basin along a descending series of high-relief ridges and troughs. The structural evolution of this eastern margin has not been well understood.
Manihiki Plateau is capped by a thick section of sediment with a well-developed internal acoustic stratigraphy (Winterer et al., 1974). The section on the High Plateau was drilled in 1973 at Site 317 (see Figure 1 for location) as part of the Deep Sea Drilling Project (DSDP). Three holes (317, 317A & 317B) were drilled in a water-depth of 2622 m and penetrated 943.5 m sub-bottom (Schlanger, Jackson et al., 1976). The section was found to consist of, (i) 424.5 m of late Pleistocene-early Eocene nannofossil and foraminiferal ooze, chalks and cherts, (ii) an uncored gap of 129.5 m probably occupied by Palaeocene pelagic sediments, (iii) Maestrichtian through Aptian-Barremian (?) chalk, chert, limestone and siltstone in the interval 554-677.5 m, (iv) volcaniclastic sandstones and siltstones (mollusc-bearing in the uppermost part of this unit, but otherwise barren of fossils) between 677.5-910 m, and (v) tholeitic, MORB-type basalt flows with intercalated thin volcanogenic siltstone/sandstone beds from 910 m to the bottom of the hole at 943.5 m sub-bottom. The age of the basalts is probably 110-112 Ma (Jackson & Schlanger, 1976).

The basalt flows and overlying volcaniclastic sediments appear to have been deposited in relatively shallow water (several hundred metres depth to ?subaerial). This is evidenced by the extreme vesicularity of the basalt, the presence of hyaloclastics and the occurrence of bivalves and gastropods of shallow-shelf (<200 m water depth) aspect in the uppermost section of the volcanics (Kauffman, 1976). The absence of fossils in all but the upper part of the volcanioclastics suggests an initial rapid accumulation rate as large volumes of basaltic lavas erupted explosively in shallow water. Then, as volcanism waned, the deposits were reworked by currents and colonized by benthonic fauna. According to Winterer et al (1974) the volcanic foundation of Manihiki Plateau evolved at a seafloor
spreading centre near the Pacific/Antarctic/Farallon plate triple junction in the Early Cretaceous (Fig. 2). The early evolutionary history of Manihiki Plateau could be likened to the modern development of Iceland.

The volcanioclastic sequence overlying the basalts is about 235 m thick and is of considerable interest because of its native copper content. The native copper, which occurs as disseminated blebs and strands, is distributed in zones throughout most of the volcanioclastic section. The average copper content of the sediment is not remarkably high - 150 ppm Cu from 6 analyses, though one analysis yielded a value of several thousand ppm Cu. Chalcopyrite is present in the underlying basalts. Jenkyns (1976) attributes the native copper to precipitation from cupriferous fluids driven through the sediment column by a convective hydrothermal system. Clay seams observed to traverse the volcanioclastic sediments may have acted as conduits for the mineralized solutions. Relatively large hydrothermal mineral deposits have been discovered along presently active seafloor spreading centres (Rona, 1984) - similar deposits may exist within the Manihiki Plateau volcanioclastic unit.

Seismic lines have been shot over the Manihiki Plateau by a number of scientific expeditions (eg Winterer et al., 1974; Schlanger, Jackson et al., 1976; Mizuno & Nakao, 1982). Some of the profiles show strata correlative with the volcanioclastics at DSDP 317 outcropping on fault scarps at places on the margin of the plateau as well as on the walls of troughs cutting across the plateau. By dredging such exposures and analysing the samples recovered, it may be possible to better define the extent of mineralization within the Manihiki Plateau volcanioclastics.
Such an experiment was conducted on the north-east margin of Manihiki Plateau during the 1984 R/V 'Sonne' research cruise SO 35-1 (Beiersdorf & Erzinger, 1986). In this paper we report on two cruises - one by HMNZS 'Tui' in 1986 and the other by R/V 'Moana Wave' in 1987 - which investigated the rugged and highly structured eastern margin of the plateau. Both cruises were conducted under the joint Australian/New Zealand/USA Tripartite II marine geoscience research program coordinated by the Committee for Co-ordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas (CCOP/SOPAC).

The investigation of the eastern Manihiki Plateau margin was aimed at multiple objectives - (i) to map the structure and stratigraphy of the area in order to try to reconstruct its geological evolution, (ii) to sample the Cu-bearing (?) volcaniclastic strata, and (iii) to sample for manganese nodules and ferromanganese-oxide crusts, study their environment of deposition and distribution, and assess their mineral resource potential. This paper concentrates on aspect (i) - the structure, stratigraphy and evolution of the margin.

**Geotectonic Background**

Winterer et al. (1974) proposed two alternative models for the evolution of Manihiki Plateau and surrounding region. These models (Fig. 2), denoted as A and B, show plate configurations at 110, 105 and 100 Ma. A third model has subsequently been presented by Winterer (1976) which is a slightly revised version of model B, but which differs little in its depiction of the evolution of Manihiki Plateau.

In the early Cretaceous, to Barremian time, new oceanic
lithosphere was being generated in the region by seafloor spreading at an active ridge-ridge-ridge triple junction between the Pacific, Farallon and Phoenix plates (n.b. the Phoenix plate is shown as the 'Antarctic' plate in Figure 2). At about 110 Ma the triple junction rapidly migrated south to a point in the vicinity of present-day Palmerston Atoll. It is postulated that the triple junction passed over a major melting anomaly in the mantle at an early stage in the southward migration. This resulted in massive outpouring of tholeiitic lavas at overlying spreading centres and the construction of Manihiki Plateau. This phase of constructional seafloor spreading was short-lived, and after about 5 Ma the plateau was abandoned on the Pacific plate by ridge crest jumps to the east and south associated with a major re-orientation of the Pacific/Farallon/Phoenix plate boundaries.

Models A and B represent different plate boundary configurations on the Manihiki plateau. In model A, a NW trending Farallon/Phoenix spreading centre is located in the middle of the High Plateau. The eastern margin is formed by a long transform fault (fracture zone) that offsets the Farallon/Phoenix ridge crest to the SSW. Model B shows the spreading centre on the High Plateau as a NNE trending segment offset from a slightly earlier or contemporaneous Pacific/Phoenix boundary to the NW by two linking transform faults striking NW. The spreading segment on the High Plateau is shown as an en echelon structure with short offsets produced by a number of minor NW trending transform faults. Thus, in both model A and B, the eastern margin of the High Plateau represents a former plate boundary across which the difference in crustal age is probably no greater than 15 Ma. According to model A the structure at the eastern margin is a transform fault separating normal oceanic crust from thickened oceanic
crust of the plateau, while model B depicts the margin as a rift margin between normal and thickened oceanic crust.

The ENE trends of the great fracture zones such as Marquisas, Galapagos and Clipperton may indicate the direction of spreading of the Pacific/Farallon ridge between 100 Ma and the imprinting of magnetic anomaly 32 at about 73 Ma. The absence of magnetic lineations over seafloor created during the long Cretaceous interval of normal polarity spanning 112-82 Ma makes interpretation of seafloor spreading geometrics a difficult task for ocean floor of this age. Little is known about the seafloor spreading history of the Penrhyn Basin adjacent to Manihiki Plateau - the basin has been poorly surveyed and much of it may lie within the Cretaceous quiet zone. Both model A and B indicate that a small block of Farallon or Phoenix plate was left abandoned adjacent to the eastern margin of the High Plateau during a major shift in plate boundaries at about 105 Ma.

Other more regional interpretations of the tectonic evolution of the Pacific area have been made (Hilde et al., 1976; Watts et al., 1980; Farrer & Dixon, 1981; Barron, 1987), but these provide neither conclusive new evidence nor details relating specifically to the evolution of Manihiki Plateau. The multiplicity of interpretations reflects the inadequate data control available for precise reconstructions.

Survey Methods and Coverage

The 'Tui' and 'Moana Wave' surveys included seismic profiling (single-channel with 120 cu. inch airgun source), gravity and magnetic
profiling, plus seafloor sampling (mainly dredging). During the 'Moana Wave' cruise an area over the eastern margin about 40 x 90 km in size, was mapped in detail using the SeaMARC II bathymetric and side-scan seafloor mapping system (Blackington et al., 1983), in conjunction with the other geophysical methods. The survey coverage is indicated in Figures 3 & 4 - the geophysical survey lines include Line T1 ('Tui') and Lines L1-L7 ('Moana Wave').

Navigation aboard 'Tui' was by the TRANSIT satellite system, while 'Moana Wave' navigation was by both TRANSIT and GPS satellite systems.

**SeaMARC II Bathymetric and Side-scan Mapping**

The 11/12 kHz SeaMARC II seafloor swath-mapping system was operated along Lines L1-L7 (Fig. 4) over the eastern margin of the plateau. With a swath-width of about 10 km in deep water, overlapping SeaMARC II image coverage was obtained between adjacent survey lines over the grid pattern selected. The longer NW-SE survey lines were oriented at a slightly oblique angle to the main structural trend to enhance response and recognition of seafloor features. A mosaic of colour-encoded bathymetric images was produced, then contoured by hand to create a map of detailed bathymetry at 100 m contour interval (Fig. 5). A mosaic of side-scan images over the survey area was also produced (Fig. 6). Typical side-scan detail of the seafloor across the margin is shown in Figure 7 by a strip of side-scan record from the central section of Line 5.

The overall morphology and structure of the margin over the
L1-L7 grid area is seen more clearly in the 500 m bathymetric contours (Fig. 8). A high-standing marginal ridge in water-depths of 2100-2200 m separates the relatively smooth surface of the High Plateau to the west, from the extremely rugged ridge and trough topography of the marginal slope descending eastward into the Penrhyn Basin. The eastern side of the marginal ridge is marked by a steep escarpment with 1600-1800 m vertical relief (Fig. 5). Smaller escarpments, 400-500 m in height, form much of the western slope of the marginal ridge. A number of high-relief ridges and troughs step the marginal slope down onto the floor of the Penrhyn Basin - this stepping-down being effected in places by major east-facing escarpments some 600-1200 m in height, which parallel the marginal ridge escarpment and cut right across the L1-L7 grid area. Troughs between the ridges are generally 3-7 km wide and have relatively flat floors; closed depressions in the trough floors are not uncommon.

The seafloor morphology observed along Line T1 (Fig. 9) is very similar to that seen over the L1-L7 grid area. On Line T1 the marginal ridge is seen to rise 700 m above the smooth plateau surface, while the Penrhyn Basin side of the ridge consists of a 1600 m high escarpment. The floor of the Penrhyn Basin at the eastern end of line T1 lies at a depth of about 5200 m, and is of uneven topography with relief of about 150 m.

Our survey results indicate that the marginal ridge is a long, extremely linear seafloor structure striking 008° (Fig. 3). This is confirmed by data from a number of previous expeditions to the eastern Manihiki Plateau region. From these data, the ridge can be traced to the north and south of our area of investigation, as well as between the L1-L7 grid and Line T1. The ridge is well defined from about
10°00' latitude (and 50 km east of Rakahanga Atoll) to 13°30' latitude (and 190 km east of Suwarrow Island), making it at least 400 km long. The ridge appears to deviate from linearity by less than a few kilometres. Poorly mapped ridges and seamounts mark what appear to be extensions of the ridge structure to the north and south (see Kroenke et al., 1985 - Fig. 1). The structure may extend as far north as the extreme north-east corner of Manihiki Plateau and as far south as Palmerston Atoll, giving it a possible overall length of 1200 km.

An interpretative map of seafloor structures observed in the SeaMARC II bathymetric mosaic (derived contours shown in Figure 5) and side-scan mosaic (Fig. 6) is presented in Figure 10. The Manihiki Plateau adjacent to the margin is seen to be fairly flat and featureless. The side-scan images show the plateau as a mainly white or light grey surface because of its low relief and the low reflectivity of surficial carbonate ooze or chalk. A number of seafloor erosional channels, up to about 100 m deep and emanating from the marginal ridge, have incised the western flank of the ridge and surface of the adjacent plateau. The cone of a mud volcano, about 200 m high and located on the plateau 20 km west of the marginal ridge, is clearly seen in the SeaMARC II bathymetric and side-scan images (Figs 5, 6, 10). This mud volcano is similar to those in a field of about 100 discovered 50 km southwest of Rakahanga Atoll at the north-eastern margin of Manihiki Plateau (Coulbourn, Hill et al., 1987; Section 2 of this thesis).

About a third of the summit areas of the marginal ridge is imaged in the side-scan records as being covered by large 'snowy' patches, which probably consist of carbonate ooze. Other areas of the ridge have a grey tone and medium-finely textured appearance,
suggesting that they may represent exposures of lithified strata.

Side-scan images of the major escarpment immediately east of the marginal ridge show it to be strongly streaked in contrasting tones in a down-hill direction (e.g., Fig. 7). This suggests intense gully development, and is confirmed by the bathymetric images which show a steep slope incised to depths mainly in the order of 100-200 m. Large fans and debris flow deposits extend out onto the trough floor at the base of the escarpment.

The side-scan mosaic (Fig. 6) exhibits a pattern of highly contrasting tones over the marginal slope of the plateau. The tonal diversity is due to both high topographic relief and variations in seafloor reflectivity. Theoretically, white areas could represent shadow zones behind topographic highs lying along-swath, or alternatively, seafloor of low reflectivity (e.g., soft sediment). Most of the larger, lighter toned areas on the marginal slope are thought to represent patches or blankets of unconsolidated sediment on the seafloor. These sediments probably comprise mainly carbonate oozes on the mid and upper parts of the slope and deep-sea red clays on the lower slope below the CCD (calcite compensation depth) of approximately 4300 m. Many areas of the relatively flat trough floors appear darker on the side-scan images than expected for ponded sediment. This may be an artifact of gain settings applied during shipboard recording, but is more likely to indicate a higher seafloor reflectivity due to an abundance of manganese nodules on the seafloor.

Most of the ridge areas show a textured pattern in medium grey to black tones on the side-scan images - with a highly contrasting dappled texture commonly present (Fig. 7). The seafloor features
producing the dark-printing patches are mainly of 200-1000 m diameter, and are believed to be hills and hillocks of rocky outcrop (? basalt). Though thin sheets of unconsolidated sediment blanket some parts of the ridge areas, surficial sediment deposits are mainly confined to hollows and channels within the rugged ridge topography.

A series of long linear structures, which trend NNE (parallel to the marginal ridge) and extend across all or much of the L1-L7 grid area, are particularly prominent in the side-scan mosaic (Fig. 6). These lineaments have been mapped in Figure 10. Their considerable extent (some are more than 35 km long) testifies to their structural importance. They have, in general, surprisingly little topographic expression - mainly less than 200 m associated relief, and appear as both narrow ridges and troughs 150-400 m wide. The structures are commonly seen as white stripes in the side-scan images, and in such cases seem to represent channels within bedrock, partially infilled by low reflectivity pelagic ooze or deep-sea clays. The most plausible interpretation of these structures is that they are fracture zones produced by transcurrent faulting in a transform fault system.

Not all structural trends seen in the SeaMARC II data are aligned sub-parallel with the plateau margin. A major fault zone, seen as a 5 km wide band of disrupted seafloor structure (Fig. 10) extending from a local eastward protrusion of the marginal ridge to the lower marginal slope, strikes E-W - approximately normal to the main marginal structures. The fault may be a transfer fault, a structure common along rifted continental margins.
Seismic Results

The seismic data, shown as line drawing interpretations, are presented in Figure 11 ('Moana Wave' lines L1-L7) and Figure 9 ('Tui' Line T1).

Three major reflecting horizons are indicated in the sections shot over the plateau - horizons b, m and k. From seismic reflector/drilling data correlations made at DSDP Site 317 (Schlanger et al., 1976), we interpret these horizons as representing the following lithostratigraphic boundaries.

<table>
<thead>
<tr>
<th>Reflector</th>
<th>Unit description</th>
<th>Depth below seafloor at DSDP 317 (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seafloor (s)</td>
<td>-</td>
<td>0</td>
</tr>
</tbody>
</table>

Foraminiferal/nannofossil ooze, chalk & chert.
Palaeocene - Quaternary.

<table>
<thead>
<tr>
<th>k</th>
<th>Nannofossil chalk, claystones and limestone - very minor interbedded volcanogenic sandstone to siltstone (near base) ?Barremian/Aptian-Maestrichtian.</th>
<th>576</th>
</tr>
</thead>
</table>

356
Volcanogenic sandstone to siltstone, containing mineralized zones.

?Barremian/Aptian (oldest fossil date 107 Ma)

Tholeiitic basalt flows. 110-112 Ma (106 +/- 3.5 Ma K/Ar minimum age).

The sedimentary section on Manihiki Plateau adjacent to the eastern margin is 0.5-0.8s (twt) thick. Deep seafloor erosion has considerably thinned parts of the section just west of the marginal ridge in the area of the L1-L7 grid. The scale of the erosion appears to increase to the north. No erosion is evident on Line T1, but as much as three-quarters of the s(seafloor)-k sequence seems to have been removed on line L1 (Fig. 11). The thickness of sedimentary section beneath the marginal ridge is often difficult to estimate from the seismic data because of faulting and acoustic scatter produced by the high relief of the feature. On Line L6 the ridge summit is underlain by about 0.75 s (twt) of sediment; much of the Manihiki Plateau section, including sequences k-m and m-b is exposed in the fault scarp on the eastern side of the ridge. East of the marginal ridge where the rugged marginal slope descends to the depths of the Penrhyn Basin, there is no evidence in the seismic data of any substantial thickness of sediment. It is conceivable, however, that some potential sub-bottom reflectors have been masked by effects of the high relief topography. Some of the deep troughs on the marginal
slope appear to contain ponded sediment, but the thickness of these deposits is difficult to ascertain. There is evidence from the T1 seismic profile that basement in troughs and beneath relatively flat-lying areas of the deeper parts of the Penrhyn Basin, is covered by about 50-100 m of sediment - probably red clays.

Normal faulting is commonly seen in the seismic sections suggesting that extensional tectonics or vertical adjustments have played an important role in the structural evolution of the Manihiki Plateau's eastern margin region. Basement faulting is often seen to extend into the m-b sequence, and sometimes right to the m boundary, which it offsets (see L6 & L7, Fig. 11; T1, Fig. 9). This suggests that the Manihiki volcanic pile was undergoing structural modification during the time, and soon after, the m-b volcaniclastic sequence was deposited. There is also evidence of some recent faulting (see L1, Fig. 11) which offsets the sedimentary section to the seafloor (albeit eroded).

As noted by Winterer et al. (1974) horizon k appears in places as an angular unconformity (see L6 & L7, Fig. 11; T1, Fig. 9). In addition, normal faults are sometimes seen to extend to the k boundary and offset it (T1, Fig. 9). This evidence, together with evidence of apparent tilting of sub-k beds (L6, Fig. 11) indicates tectonic upheaval at Maestrichtian/Palaeocene time - about 65 Ma. Such tectonic activity coincides with the onset of subsidence of the postulated regionally-extensive Darwin Rise (Menard, 1964; Schlanger & Premoli Silva, 1981), which, it is believed, incorporated Manihiki Plateau. The geological record from cores recovered at drill site DSDP 317 shows a dramatic increase in sediment accumulation rate at this time (Schlanger, Jackson et al., 1976). Sedimentation, which was entirely
pelagic, jumped from a rate of about 1 m/Ma to 10 m/Ma. This may be related to vertical tectonics and ocean current circulation patterns, but could also be due to Manihiki Plateau's northward movement on the Pacific plate into the equatorial high-productivity zone where the supply of nannofossil and foraminiferal carbonate is abundant.

**Sampling Results**

On completing the geophysical line T1 during the 'Tui' cruise, the line was retraced to the west and the seafloor sampled at a number of locations - stations U296a to U302 (Fig. 3). The results are given below:

<table>
<thead>
<tr>
<th>Station</th>
<th>Sampling Device</th>
<th>Depth (m)</th>
<th>Material Recovered</th>
</tr>
</thead>
<tbody>
<tr>
<td>U296a</td>
<td>Pipe dredge</td>
<td>5130</td>
<td>Abundant spheroidal, polynucleate Mn nodules (574 total; 10-83 mm max. diam.).</td>
</tr>
<tr>
<td>U297a</td>
<td>Short corer</td>
<td>5227</td>
<td>0.5 m core of red clay.</td>
</tr>
<tr>
<td>U298</td>
<td>Pipe dredge</td>
<td>4996</td>
<td>Abundant (574) basaltic scoria pebbles (to 34 mm diam.); spheroidal Mn nodules (10 total; 16-59 mm max. diam.); red clay.</td>
</tr>
<tr>
<td>U299a</td>
<td>Rock dredge</td>
<td>4637</td>
<td>Spheroidal Mn nodules (4 total; 16-</td>
</tr>
</tbody>
</table>
The 'Tui' sampling results indicate that, (i) the floor of the Penrhyn Basin at the lower slope of the Manihiki margin consists of red clays bearing a high concentration of Mn nodules, (ii) Mn nodule fields blanket much of the rugged seafloor of the eastern marginal slope; Mn crusts of at least 35 mm thickness have been deposited in this area; and evidence of local basaltic volcanic activity is indicated by a large haul of scoria, and (iii) the relatively flat surface of the plateau adjacent to the eastern margin is covered by nanno-foraminiferal ooze.

The objective of the 'Moana Wave' rock dredge at station RD16 (Fig. 4) was to sample volcaniclastic sediments of the Manihiki section exposed on the major escarpment immediately east of the marginal ridge (Fig. 11, Line L6). The dredge was lowered in 2600 m of water and hauled up the escarpment. A total of 20-25 kg of rocks was recovered, including cobble-size unconsolidated and consolidated
volcanic sandstones, siltstones and claystones. Most of the specimens were covered by 0.5-2 cm of Mn crust; two samples had very well developed crusts to 4.5 cm thick. The largest rock in the haul was a 40 x 30 x 30 cm limestone boulder pock-marked with bore-holes and partly encrusted by a Mn deposit up to 3 mm thick. The lithologies recovered were as anticipated from the seismic data, with seismic horizons correlated across to DSDP 317. The stratigraphic interval sampled is believed to correlate with the Barremian-Aptian section drilled at DSDP 317, comprising a 235 m thick basal sequence of volcanioclastics (sandstone-claystone) overlain by 50 m of nannofossil micritic limestone. Preliminary microscopic examination of the volcanioclastic samples in both thin and slab section revealed no evidence of native copper particles. A similar absence of native copper was observed in volcanioclastic sediments dredged from the north-east margin of Manihiki Plateau (Beiersdorf & Erzinger, 1986). Either copper mineralization was never developed, or quite possibly, was leached out by circulation of sea-water through submarine exposures of the volcanioclastic sediments.

Comparison with sampling results of previous expeditions

Several previous research cruises have conducted seafloor sampling operations over the eastern margin of the High Plateau in the vicinity of our area of investigation. During the 1962 'Vema' Cruise 18 (Heezen et al., 1966) the margin was sampled by piston corer and dredge; bottom photographs were also taken. Sample stations were located, (i) from the lower marginal slope to the marginal ridge along a transect almost coincident with Line T1, (ii) on the marginal ridge about 15 km south of the L1-L7 grid, and (iii) on the east-facing marginal ridge escarpment and in the trough at the base of the escarpment - close to 'Moana Wave' RD16 dredge station. A polygonal
area of seafloor on the marginal slope, located close to the 'Vema'
transect and Line T1, was extensively sampled and photographed during
the 1968/70 research program of RV 'Vityaz' (Bezrukov, 1971; 1973).
Seafloor in the depth range of 2000-5250 m was investigated. The RV
'Hakurei-Maru' sampled two sites in the area during the 1980 GH80-1
breeze (Mizuno & Nakao, 1982). The first site (station 1615),
located on the plateau adjacent to the marginal ridge about 18 km
north of Line L1, was sampled by piston corer and free-fall grabs
with camera; while the second (station 1616) located
in the trough at the south-east corner of the L1-L7 grid, was sampled
by box corer and free-fall grabs with camera.

The findings of the earlier expeditions were similar to our own.
Sufficient data are now available to provide well-based insight into
the geology and character of the seafloor at the eastern margin of the
plateau. Some of the more significant findings of the previous work,
which extend or support our own results, are documented and
discussed below.

During 'Vema' Cruise 18 siltstone and claystone
pebbles/fragments (?from sequence m-b) were recovered from locations
(ii) and (iii) on or adjacent to the marginal ridge. The sampling
records indicate that all stations from which siltstone was recovered,
except one, are located close to the marginal ridge. The exception is
'Vema' V-18 core station 276 situated just north of Line T1 about 26
km east of the marginal ridge crest. The core, from a water-depth of
3587 m, contained siltstone fragments within a graded bed of manganese
nodules and fragments. It is quite possible that the siltstone-
bearing sediment was transported from the marginal ridge by turbidity
currents. Bezrukov (1973) records recovery of large blocks of
Mn-encrusted stratified tuff sandstones and tuff gravel-stones (mainly hyalobasalts) from the polygonal area investigated on the margin near Line T1. No station co-ordinates are given, so these sedimentary rocks may well have come from the marginal ridge.

Heezen et al. (1966) conclude that the eastern escarpment of Manihiki Plateau is underlain by basic igneous rocks. The recovery of gabbroic rock fragments and serpentinites is further evidence of major faulting. High heat flow in the area may also indicate the presence of a major fault zone (Bezrukov, 1971).

All stations on the High Plateau adjacent to the eastern margin indicate a surface of light brown nanno-foraminferal ooze devoid of manganese nodules. The presence of carbonate ooze even on those areas adjacent to the marginal ridge, shown by the seismic profiles to be severely eroded, may indicate a recent return to tranquil bottom-water conditions and renewed deposition. This is supported by bottom photographs which show a tracked and burrowed surface without any sign of strong current activity.

The absence of Mn nodules on the plateau contrasts with the situation on the marginal slope where nodules are abundant both on ridges and in troughs. It is not uncommon for nodule coverage of the seafloor to be as high as 50-70%. In addition, all exposed rock surfaces on the marginal slope are FeMn-oxide coated or encrusted - this is observed in bottom photographs and seen on dredged rock specimens. Bottom photographs taken on the upper marginal slope (Heezen et al., 1966) show foraminiferal sand surfaces to be rippled and current lineated, as well as scoured around nodules and boulders embedded in the sands. This evidence of vigorous bottom-water
movement suggests conditions of low sedimentation or even erosion. Such conditions are known to be conducive to manganese nodule development (Kennett & Wilkins, 1975; Pautot & Melguen, 1976), which probably explains the abundance of nodules and the well-developed FeMn-oxide crusts on the marginal slope.

**Gravity and Magnetic Surveys**

Gravity and magnetic data were acquired during both the 'Tui' and 'Moana Wave' cruises wherever such acquisition was possible and where such data would serve a useful purpose in interpretation of the marine geology. The L1-L7 grid area over the eastern margin of the plateau, shown in Figure 8 together with bathymetric contours at 500 m interval, was mapped in detail both gravimetrically and magnetically by 'Moana Wave'. The results of these surveys are presented in Figures 12 & 13 as gravity and magnetic contour maps which also indicate the data distribution. Figure 12 shows free-air gravity anomaly contours at 5 mgal interval, while Figure 13 depicts magnetic anomaly contours at 25 nT interval. The magnetic anomaly data for Lines L1 & L4-L7 can also be seen in profile form in Figures 14-23. Free-air gravity anomaly and magnetic anomaly profiles for the 'Tui' Line T1 are presented in Figure 9, together with the seismic interpretation for that line. In all cases, the magnetic anomaly field has been calculated by subtracting the IGRF80 global reference field (Peddie, 1982) from the observed data. However, in generating the contour maps (Figs. 12 & 13) some minor 'drift' adjustments were made to tie the data at line intersections.
Gravity field description and qualitative interpretation

The gravity field over the eastern margin of Manihiki Plateau and the western Penrhyn Basin shows a broad correlation with seafloor topography - but certainly not a direct correlation. Free-air gravity over Penrhyn Basin ranges between 0 and -15 mgal, and shows a general increase over the plateau margin towards the marginal ridge. The ridge and trough topography on the marginal rise is replicated to some extent by local anomalies in the gravity profiles. An obvious example of this is the large ridge with about 1000 m relief situated on the mid-rise (centre of Figs 5 & 8), which is expressed in the gravity field as a 30-40 mgal positive anomaly. The marginal ridge appears in the gravity data as a 35-55 mgal free-air gravity high. Farther west, over the plateau, the field decreases to about +15 mgal.

The difference in free-air anomaly between that in the Penrhyn Basin (-15 to 0 mgal) and that over the Manihiki Plateau (about 15 mgal) is relatively small considering the tremendous change in seafloor elevation, amounting to about 3000 m. The relatively subdued gravity expression of the plateau (see also Haxby, 1987) suggests that the mass of Manihiki Plateau is isostatically compensated. In addition, the absence of a well defined gravity 'moat' (such as surrounds the islands of Hawaii) beyond the plateau margin implies that compensation is local, signifying that the plateau was emplaced on relatively young oceanic crust of low flexural rigidity, i.e. close to a spreading centre. The fact that the plateau appears to be isostatically compensated, coupled with its considerable 3000 m relief above the adjacent ocean basins, indicates that the crustal thickness of Manihiki Plateau is substantially greater than that of normal oceanic crust.
Magnetic field description and qualitative interpretation

Magnetic anomaly, along the lines surveyed across the eastern margin and western Penrhyn Basin, ranges in amplitude from about -90 to +260 nT. A positive magnetic anomaly of 120-230 nT was recorded over the marginal ridge signifying an internal basaltic composition, with magnetization being of normal rather than reverse polarity. A magnetic high of about 200 nT also appears to be associated with a wide seafloor ridge of 1000 m relief just east of the middle of Line 4 (Figs 16 & 17). This anomaly does not extend farther SSW (Fig. 13) as would be expected from structural trends in the area. The structural discontinuity (?fault zone) of 106° strike, seen in the SeaMARC II data, also appears to coincide with a discontinuity in the magnetic pattern over the L1-L7 grid. An imaginary line drawn along the discontinuity passes between the two local positive anomalies located above the marginal ridge, truncates the fore-mentioned 200 nT anomaly to the south, and truncates the elongate 180 nT magnetic high recorded at the eastern ends of Lines L5, L6 & L7 to the north.

The origin of the large 260 nT positive anomaly over the middle of the marginal rise on Line T1 is problematic. The 180 nT high at the eastern ends of Lines L5, L6 & L7 may be a NNE extension of the Line T1 high. These anomalies do not appear to be related to any major change in seafloor topography, and therefore must be produced by lateral magnetization contrasts in the upper crust - probably as a result of faulting (?transform) or geomagnetic field reversal at the time of crustal accretion.

The magnetic field over the Manihiki Plateau is relatively subdued and predominantly of long wavelength, signifying that much of
the field variation is due to magnetic heterogeneity within basement.

**Werner Deconvolution Interpretation of Magnetic Profiles**

Magnetic data along the longer lines crossing the plateau margin at close to normal angles (Lines L1, L4-L7 & T1) were processed by the Werner deconvolution technique (Jain, 1978; Hsu & Tilbury, 1977) in an attempt to establish the depth of magnetic bodies producing the anomalies seen in the profiles and to determine the geometries of the magnetic structures if possible. The internal acoustic structure of many of the high-relief ridges along the margin is poorly defined because of the steep and rugged topography. For this reason, it was hoped that the Werner deconvolution processing would allow an estimate of depth to magnetic basement within the ridges and so give some idea of whether or not the ridges, believed to be fault blocks, are capped by sedimentary strata - possibly those of the Manihiki Plateau succession deposited prior to major structural development at the margin.

Werner deconvolution provides estimates of sub-surface position and magnetization direction for two types of two-dimensional magnetic model - thin sheet (representing dyke-like bodies) and interface (representing faults or dipping contacts). Because of the requirement that profiles run approximately normal to strike, profiles (magnetic and bathymetric) for Lines L1 & L4-L7 were projected onto an azimuth of 114° which corresponds approximately to the local structural cross-strike direction, while the Line T1 profiles were projected onto an azimuth of 098° which is normal to the regional strike of the margin structures. Werner deconvolution solutions for thin sheet and
interface models, together with respective magnetic anomaly and bathymetric profiles, are contained in Figures 14-25. In relation to assessment of Werner deconvolution plots, it should be noted that clusters of magnetic source estimates are indicative of reliable solutions, while isolated or scattered estimates are generally of low significance. Magnetic anomalies often yield depth estimates for both thin sheet and interface models, with the thin sheet estimates generally being in the order of 20% deeper than the interface estimates. Which of the two is the better solution depends on the shape of the causative magnetic body. A further point to note is that the depth of interpretation is reduced at the start and end of profiles because of limitations to the length of the data-point window, which scans the profile during Werner deconvolution, at the profile extremities.

The high amplitude, long-wavelength component of the field seen in this profile is probably of deep-seated origin - perhaps due to a magnetic discontinuity at 8-9 km depth indicated by clustering of interface-model estimates beneath the plateau half-way along the profile. The source estimates do not clearly outline basaltic basement beneath the plateau, but this is expected due to the fairly smooth upper surface seen in seismic profile. Several thin-sheet model clusters roughly coincide with top of basement. The grouping of estimates at about 1-2 km sub-bottom immediately west of the marginal ridge is attributed to basement faulting seen in the seismic section. The marginal scarp is located near the end of the profile and this may explain, to some extent, the few estimates associated with the ridge. There is some evidence that magnetic basement may lie about 200-300 m below the summit as suggested by the seismic.
A major basement fault marked by the marginal ridge escarpment is clearly indicated by the Werner deconvolution solutions. The low ridge about 10 km east of the escarpment appears to have moderate magnetic expression and probably consists of basalt, exposed or at shallow depth (less than 200 m sub-bottom). Tightly clustered, high intensity source estimates underlie the prominent ridge located just east of the centre of the profile, at a depth of about 6-8 km. The magnetic contours (Fig. 13) and anomaly profile show a local, high-amplitude (200 nT) anomaly at this location; the anomaly has no direct extension along regional structural strike to the south. Such departure from magnetic two-dimensionality casts some doubt on the accuracy of the depth determination, and raises the question of whether the clustered estimates represent a real magnetic source or are merely an artifact of unmapped high-relief basalt seafloor topography to the north or a magnetic body of limited strike length at depth. Estimates located close to seafloor just east of the ridge suggest seafloor basalts.

A cloud of source estimates surrounds the summit area of the marginal ridge; the estimates here include a tight cluster near the eastern edge of the summit. These indications suggest a ridge composed largely of basalt, with little or no sediment cover. The dense cluster of estimates which plot just above the 300 m high ridge to the east of the marginal ridge indicates a moderate magnetization and also probable basaltic composition. Scattered source estimates and some clusters suggest that the broad mid-slope ridge, just east of the centre of the profile, is composed of magnetic rocks. The
proximity of the large positive magnetic anomaly seen immediately to
the north on Line 4 has probably produced some distortion in the L5
profile, thereby reducing the accuracy of the source estimates. A
vertical band of thin-sheet source estimates is located beneath the
trough to the east of the broad ridge and extends from the seafloor to
a depth of almost 8 km. These estimates may represent a boundary
between contrasting magnetization in the upper crust (?transform
fault, ?geomagnetic reversal during seafloor spreading). A magnetic
anomaly high of about 150 nT is present immediately to the east,
towards the end of the line. The high cannot be readily correlated
with seafloor topography.

Line L6

Clusters of magnetic source estimates beneath the summit area
of the marginal ridge indicate a sub-bottom depth of magnetic basement
in the order of 400-650 m, which is in approximate agreement with the
seismic and dredging data. The band of estimates extending downwards
from the foot of the marginal ridge escarpment is probably the
expression of a major marginal fault, the presence of which was also
inferred from the Line L4 data. Source estimates farther to the east
along the line lie mainly close to the seafloor, or less than about 2
km sub-bottom. This suggests that the seafloor is likely to be
composed of basalts with little or no sediment cover - at least on the
ridges.

Line L7

The estimates lying up to 3.5 km sub-bottom west of the
marginal ridge may be linked with intense basement faulting seen in
the seismic data. A paucity of estimates at the ridge summit suggests
that basaltic basement rocks beneath the ridge may be overlain by
sedimentary strata - perhaps in the order of 200 m thick, though such an interpretation is based on meagre evidence. Lateral magnetic contrast at the high escarpment on the eastern side of the marginal ridge and/or magnetic contrast produced by nearby associated marginal faults appear to be the source of estimates located on the escarpment and up to about 1.5 km beneath its upper slope. Farther east, estimates occur both as clusters and as isolated points, with depth distribution generally ranging from seafloor to about 2 km sub-bottom. Basaltic seafloor is the probable source of these estimates. The 700 m high peak towards the eastern end of the line appears to have no magnetic expression, and may be composed of a weakly magnetic lithology - possibly a pyroclastic material such as hyaloclastite.

**Line T1**

Along this line many of the estimates plot in the lower half of the water column - this is particularly the case along the Penrhyn Basin half of the line. The reason for this is not clear, but reflects a slightly higher spatial frequency content in the magnetic profile than would be expected. It may have been caused by mild geomagnetic storm activity during the survey; a small shift to higher frequencies may also have been introduced in projecting the profile normal to regional structural strike. Nevertheless, most of the reliable source estimates (those grouped together and of high magnetization) along the Penrhyn Basin and lower marginal slope section of line are seen to plot close to seafloor or shallower than about 1.5 km sub-bottom, suggesting that this area is floored by basalts. The large escarpment on the eastern side of the marginal ridge and inferred fault or faults that extend into the sub-bottom from the escarpment, are clearly indicated by the steeply dipping line of source estimates. Estimates are tightly massed in the vicinity of...
the summit of the marginal ridge indicating it is a relatively strongly magnetized basement (?basalt) block with little or no sediment cover. The very steep gradient and 300 nT excursion seen in the magnetic profile about 20 km east of the marginal ridge is quite strongly expressed in the Werner deconvolution estimates as a band of possible sources extending from about seafloor level to at least 2 km sub-bottom. This band of estimates probably corresponds to a large magnetic discontinuity in the upper crust.

**Crustal Structure**

Sonobuoy seismic refraction soundings have been made at a number of locations on and adjacent to Manihiki Plateau (Sutton et al., Hussong et al., 1979). Some of the soundings (68, 70 & 71 - Fig. 3) were made within our area of investigation.

Sounding 68 in the Penryhn Basin indicates a typical ocean basin crust with a mantle depth of about 11.5 km, while beneath Manihiki Plateau the average structure derived from a number of soundings suggests a velocity structure similar to that of typical oceanic crustal layers 2A-3A but that beneath Manihiki Plateau each of these layers is about 3.1 times thicker than normal (Hussong et al., 1979). The average crustal model for Manihiki Plateau places the deepest recorded layer of 6.8 km/s at a depth of 10.7 km. No refraction data with deeper penetration to the lower crust and mantle are available. On the basis of the similarity of the shallow crustal structure between the Ontong Java and the Manihiki plateaus, Hussong et al. (1979) extrapolate a mantle depth of about 25-30 km beneath Manihiki Plateau.
Our gravity data and seismic profiles allow us, we believe, to make an improved estimate of the crustal thickness beneath the plateau and to construct an approximate crustal model of the eastern margin of the plateau. We have done this by forward gravity modelling based on the constraints tabled below.

Gravity model constraints

<table>
<thead>
<tr>
<th>Layer velocity (km/s)</th>
<th>Density (t/m³)</th>
<th>Thickness (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Manihiki Plateau</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2.0a (sediment)</td>
<td>1.9</td>
<td>0.5-0.9</td>
</tr>
<tr>
<td>4.6 (Layer 2A)</td>
<td>2.5</td>
<td>1.5 approx.</td>
</tr>
<tr>
<td>5.4 (2B)</td>
<td>2.65</td>
<td>3.1</td>
</tr>
<tr>
<td>6.2 (2C)</td>
<td>2.75</td>
<td>3.2</td>
</tr>
<tr>
<td>6.8 (3A)</td>
<td>2.9</td>
<td>*</td>
</tr>
<tr>
<td>7.4b (3B)</td>
<td>3.1</td>
<td>*</td>
</tr>
<tr>
<td>8.25b (Mantle)</td>
<td>3.35</td>
<td></td>
</tr>
<tr>
<td><strong>Penrhyn Basin</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4.2 (2A)</td>
<td>2.4</td>
<td>1.3</td>
</tr>
<tr>
<td>6.2 (2C)</td>
<td>2.75</td>
<td>1.3</td>
</tr>
<tr>
<td>6.8 (3A)</td>
<td>2.9</td>
<td>1.5</td>
</tr>
<tr>
<td>7.2 (3B)</td>
<td>3.1</td>
<td>2.2</td>
</tr>
<tr>
<td>8.2 (Mantle)</td>
<td>3.35</td>
<td></td>
</tr>
</tbody>
</table>

1 Based on experimental seismic velocity/density relationship of Ludwig et al. (1970).
2 Approximate value based on wide-angle reflection data (Winterer et al., 1974) and the sediment column velocity at DSDP 317 of 2.09 km/s (Schlanger,
Velocities inferred from Hussong (1972), Hussong et al. (1979) and other literature on Pacific refraction data.

To be determined by gravity modelling, but ratio of layer 3B thickness:
Layer 3A thickness is assumed to be 1.7 based on Pacific Basin data (Hussong, 1972; Hussong et al., 1979).

Figures 26 and 27 show profiles of free-air gravity anomaly and bathymetry for Lines L7 and T1 (respectively), projected onto lines normal to the general strike of the marginal structure (i.e. projected onto an azimuth of 098°). The two-dimensional forward modelling was done by varying the thicknesses of the lower crustal layers beneath Manihiki Plateau and also the geometry of the transition zone between Manihiki Plateau and Penrhyn Basin crust, until a good fit between observed and calculated gravity was achieved. The theoretical gravity for the various models tested was calculated by the method described by Talwani et al. (1959). The final crustal models depicted in Figures 26 and 27 provide close agreement between observed and calculated gravity profiles. Depth to mantle beneath Manihiki Plateau is interpreted as 19.5 km - at least 5.5 km shallower than predicted by Hussong et al. (1979). The zone of crustal thinning beneath the plateau margin is modelled as being 75 km wide. Crustal thickness changes across this zone from 17 km beneath the plateau to 6.3 km in the Penrhyn Basin. The gravity modelling cannot resolve with certainty whether the margin represents a rift or transform structure.
Discussion on Structural Evolution of the Margin

Two main phases of tectonism post-dating construction of the Manihiki volcanic edifice are evident in the data.

**Phase 1** is represented by normal faulting that affects basement and extends partly or completely through the overlying basal volcaniclastic sequence (m-b). These structures were probably created in response to late-stage oceanic rifting in the area and to differential subsidence as the volcanic pile began to cool and sink. The time of this episode of deformation would be late Barremian-early Aptian, about 112-107 Ma.

**Phase 2** is expressed as, (i) normal faulting that extends through the chalk/nannofossil claystone/limestone sequence k-m, (ii) tilting of the sub-k units at the plateau margin, and (iii) angular unconformity k. The onset of phase 2 probably occurred during the late Maestrichtian-Palaeocene (about 65 Ma).

Minor deformation affecting the Tertiary section is also seen on the High Plateau, but is restricted to small-scale local normal faulting, slumping and compaction structures. A 500 m thick pile of pelagic carbonate has accumulated on the plateau since the beginning of the Tertiary. Adjustment of basement to the additional load and internal soft-sediment deformation are likely causes of many of the observed structures.

There is no clear indication from the geophysical and geological sampling data that either sequence m-b or sequence k-m extend farther east than the graben at the base of the east-facing...
escarpment on the marginal ridge. The seismic profiles show both sequences to be truncated at the fault scarp. Truncation of these sequences is also indicated on the CATO-3 seismic profile across the plateau margin to the south of our area of investigation - at about 14°S (Winterer et al., 1974). Here there is no pronounced marginal ridge and the sediments simply off-lap at the marginal escarpment. No eastward continuation of the sequences is evident in the jumble of steep-sided ridges and troughs that constitute the marginal slope.

The extraordinary length (up to 1200 km) and linearity of the marginal ridge and structural extensions to the north and south, as well as the extensive set of side-scan lineations trending parallel to the marginal ridge, strongly suggest transform faulting as the primary structural mechanism. The apparent lateral truncation of sequences m-b and k-m is consistent with their being offset by transform movement. Thus the linear eastern margin of the High Plateau is tentatively interpreted as a fracture zone active at the time of phase 2 tectonism, i.e. from about 65 Ma.

Two additional, though approximate, controls on tectonism at the plateau margin are available. The first is the age of the Penrhyn Basin seafloor. The mean depth of the central part of the basin is about 5250 m. According to the Parsons & Sclater (1977) age/depth relationship for typical ocean floor, an ocean basin of such depth would be expected to be about 64 Ma old. Rough seafloor topography, thin sediment cover and the presence of large magnetic anomalies without associated seafloor topographic expression all argue in favour of the Penrhyn Basin being younger than mid-Cretaceous as proposed by Winterer et al. (1974). The second control is the time of faulting at the margin derived from the thickness of FeMn-oxide crust. The
maximum thickness of crust recovered from fault scarps at the margin during our cruises was 45 mm. If we follow Beiersdorf & Erzinger (1986) and assume an accretion rate of 1 mm/Ma, we arrive at an estimate of 45 Ma for the minimum age of faulting. Such a figure is, admittedly, a rough estimate being based on a number of assumptions, yet it still provides useful order-of-magnitude control.

There is further support for the concept that the area to the east of the High Plateau was the site of active seafloor spreading with associated transform faulting at the plateau margin sometime during the Maestrichtian-early Tertiary. Farrer & Dixon (1981) postulated that at about 67 Ma the Pacific plate ruptured along a fracture zone system comprising the Emperor trough and the Line Islands lineament, and that in the interval 67-40 Ma some 1700-1900 km of dextral offset was accumulated between the east and west Pacific plates. Their proposed model provides an integrated explanation of a number of otherwise disparate observations relating to the Tertiary evolution of the Pacific plate. What is lacking in their model is a more explicit explanation of the tectonic development at the south-east end of the fracture system where the model requires the opening of new ocean basins about 1700 km in total width. Farrer & Dixon (1981) suggest that the requisite seafloor may have been generated along the Nova Canton Trough (north of Manihiki Plateau), east of Manihiki Plateau (Penrhyn Basin) and to the south of the plateau. A major structural element controlling this phase of extensional tectonics may have been the interpreted transform fault zone at the eastern margin of the Manihiki Plateau.
Acknowledgements

D. Bergersen, graduate student of Hawaii Institute of Geophysics (HIG), assisted with the preparation of Figure 5 (bathymetry).

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Southwest Pacific. CCOP/SOPAC. Suva, Fiji.


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Figure 27. Interpreted crustal density structure across the eastern margin of Manihiki Plateau, based on Line T1 data and sonobuoy seismic refraction control.
FIGURE 1 (S10).
Plate tectonic evolution of Manihiki Plateau & its eastern margin

FIGURE 2 (S10).
FIGURE 3 (S10).
FIGURE 4 (S10).
SEAMARC II
DETAILED BATHYMETRY
100 M CONTOUR INTERVAL

Depths in km

FIGURE 5 (S10).
See also next page
FIGURE 6 (S10).
See also next page
SIDE-SCAN IMAGE
LINE 5 (CENTRAL SECTION)

PLATEAU

MARGINAL RIDGE

TROUGH

RIDGE

TROUGH

FIGURE 7 (S10).
FIGURE 8 (S10).

BATHYMETRY

0.5 KM CONTOURS
FIGURE 9 (S10). SEISMIC, MAGNETIC & GRAVITY PROFILES --- TUI LINE T1
Seafloor Structures

FIGURE 10 (S10).

Steep seafloor slopes (≈12°). Arrows indicate direction of fall.
SEISMIC PROFILES
MOANA WAVE LINES

FIGURE II (CISIO)

Sec (twt)
FIGURE 13 (S10).

MAGNETIC ANOMALY CONTOURS
MANIHIKI EAST

nT

0 20 Km

198° 45' 198° 50' 199° 00' 199° 10' 199° 20' 199° 30' 199° 40'

11° 45' 11° 50' 12° 00' 12° 05' 12° 10' 12° 15' 12° 20' 12° 25'
FIGURE 18 (S10).
Figure 21 (S10)
LINE T1 W.DECON

FIGURE 24 (S10).
FIGURE 25 (S10).