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Marine Geology of the New Zealand Subantarctic Sea Floor

by

C. P. SUMMERHAYES

New Zealand Oceanographic Institute
Memoir No. 50

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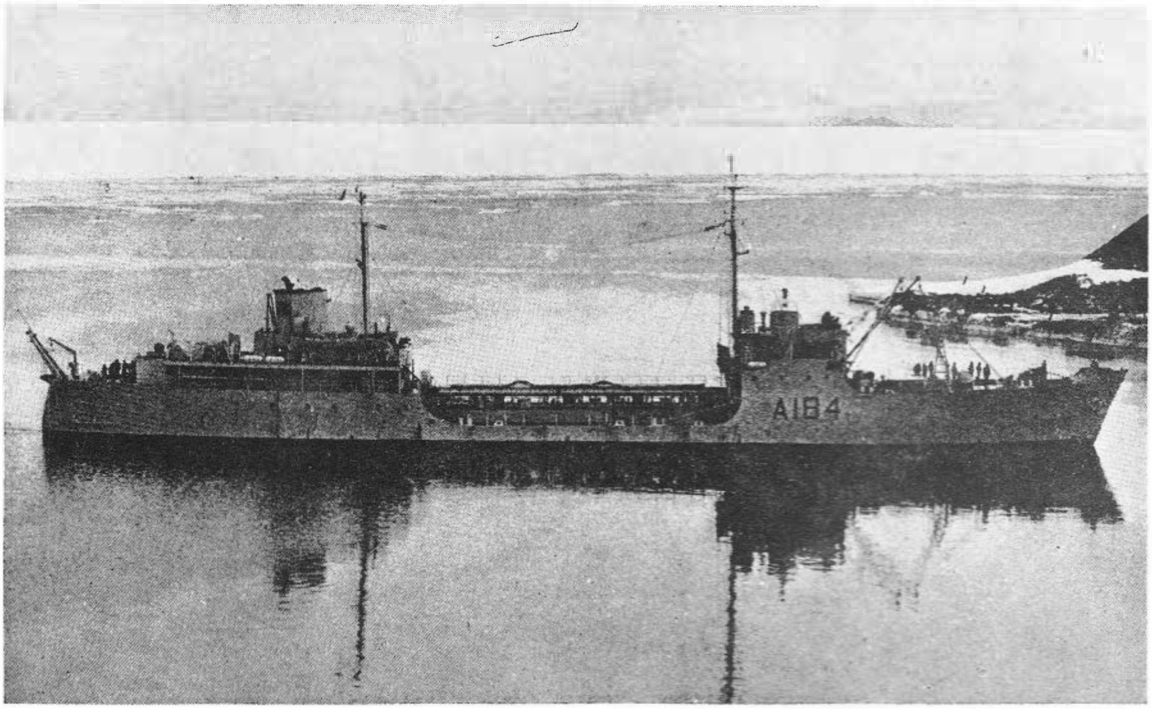
OF THE
NEW ZEALAND SUBANTARCTIC SEA FLOOR



Photograph: J. Calvert

HMNZS *Endeavour* in McMurdo Sound

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FOREWORD

Over the past four years Institute cruises on RNZN vessels, carried out as part of the N.Z. Antarctic programme, have been planned to provide data on the morphology, sediments, and structure of the New Zealand Subantarctic region. The two principal elements of this area—the Campbell Plateau and Macquarie Ridge—have been investigated during four cruises on HMNZS *Endeavour*.

The first of these carried out in October 1959 was led by D. E. Hurley; the second in April 1963 led by E. W. Dawson; the third in January 1964 led by E. W. Dawson; and the fourth in January 1965 led by I. N. Estcourt.

Additional data have come from an Institute cruise on USS *Glacier* in February 1966 led by E. W. Dawson.

The present memoir provides a description and interpretation of the resulting bathymetric, sedimentary, and structural data, covering this extensive area of the sea floor south of New Zealand.

Miss B. J. Davison assisted in the preparation of the manuscript for publication.

J. W. Brodie, Director,
New Zealand Oceanographic Institute, Wellington.

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FIGURES

Marine Geology of the New Zealand Subantarctic Sea Floor

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ABSTRACT

The sea floor between New Zealand and the flanks of the Pacific-Antarctic Ridge is diversified by an extensive submarine platform, the Campbell Plateau, and a narrow arcuate ridge, the Macquarie Ridge.

Campbell Plateau has a geological history extending at least to the Triassic, if not into the Precambrian. It is believed to be the extension of the Lord Howe Rise, displaced across the Alpine Fault. It formed part of the foreland for the New Zealand Mesozoic Geosyncline, was eroded to base level during the late Cretaceous, and has remained at about its present depth since the beginning of the Tertiary. Late Tertiary alkaline volcanicity is associated with elevation of rises on the plateau.

Phosphatised early and mid-Tertiary foraminiferal oozes from parts of the Campbell Plateau indicate that this type of sediment has been deposited here continually since the end of the Cretaceous. Authigenic glauconite, phosphorite, and manganese minerals are found locally on the plateau surface.

Macquarie Ridge is a Tertiary - Quaternary island arc extending as a geophysical and morphological entity into southern New Zealand. Pre-Miocene uplift and erosion were followed by Miocene and Pliocene tholeiitic volcanicity. The ridge is seismically active.

Between the ridge and plateau is the Solander Trough - Emerald Basin - Waiau Depression rift system, floored in part by oceanic crust and characterised near New Zealand by hornblende andesite volcanism at Solander Island.

Evolution of mid-ocean rises during the Permian in the Indian Ocean and during the Tertiary in the Pacific caused the fragmentation of Gondwanaland. Interaction in the New Zealand region between these two oppositely directed convective systems led to the present configuration of local structural elements. Disposition of the presently active island-arc system from Tonga to Macquarie Island is thought to be closely related to formation of the East Pacific Rise, to which it is subparallel for most of its length.

INTRODUCTION

AREA STUDIED

The area considered in this work (Fig. 1) is that covered by the Campbell, Pukaki, Auckland, and Macquarie Sheets of the N.Z. Oceanographic Institute 1 : 1,000,000 series oceanic charts. It covers some 500,000 square miles between latitudes 48° and 57° 30′ S and longitudes 157° and 180° E.

The personal interest of the writer in the structure of this region was aroused on the 1965 "Campbell Plateau" cruise on HMNZS *Endeavour* under the leadership of Mr I. N. Estcourt. Visits to the Campbell, Antipodes, and Auckland Islands allowed the writer to familiarise himself with the subaerial geology. Rock and sediment samples collected on the cruise contribute to the geological discussion.

Soundings from this area in the N.Z.O.I. collections date back to the 1933–35 cruise of the *Bear*. Following this, data were collected on specifically oriented scientific cruises and on supply trips to and from Antarctica and under joint N.Z. - U.S.A. Antarctic Research programmes (Table I, Appendix I).

Within the area studied are six islands or island groups: Snares Islands, Auckland Islands, Bounty Islands, Antipodes Islands, Campbell Island, and Macquarie Island. Emerald Island (reported by Mr C. J. Nockells of the vessel *Atlantic* in 1821) is no longer thought to exist (U.S. Navy Department Hydrographic Office, 1943).

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SUBMARINE FEATURES

Detailed morphological descriptions are given in Appendix IV and discussed in geomorphic terms on pages 32–48. The main sea-bed features recognised in this area are: Campbell Plateau, Macquarie Ridge, Solander Trough, Emerald Basin, Southwestern Pacific Basin, and Tasman Basin.

CAMPBELL PLATEAU

The Campbell Plateau was first named by Fleming and Reed (1951), who defined it as the irregular submarine platform, south of the Bounty Trough, on which the Subantarctic Islands of New Zealand are situated. Bounded by latitudes 48°–55° S and longitudes 166°–180° E, it forms part of the New Zealand Plateau (Brodie, 1964), a broad region less than 500 fathoms deep completely surrounding New Zealand. It covers some 200,000 sq. miles (about twice the area of New Zealand), and has a flat or gently undulating surface lying between 340 and 1,000 m. Being a shelf-like feature of greater depth and breadth than the continental shelf, from which it is separated by a low continental slope, it may be classified physiographically as a continental marginal plateau (Heezen, Tharp, and Ewing, 1959).

MACQUARIE RIDGE

This is a narrow ridge-like feature extending from about 60° S through Macquarie Island to the New Zealand Shelf off Puysegur Point (Brodie and Dawson, 1965). The ridge crest lies mainly in depths of less than 2,000 m and may be flat topped in 1,000 m or less. Its rugged topography has been briefly described by Brodie and Dawson (1965). Association of the ridge with earthquake epicentres has recently been indicated by Cooke (1966) and its magnetic character studied by Hatherton (1967).

SOLANDER TROUGH

A large linear depression extending south from the vicinity of Solander Island to the latitudes of the Auckland Islands was named the Solander Trough by Brodie (1958), but has not received any attention as a major structural feature. At approximately 51° S three seamounts rise from its centre, and Solander Island lies in a similar position at the head of the trough. The trough does not reach typically oceanic depths over most of its length.

EMERALD BASIN

This elongate, flat-floored basin stretches south to about 57° S from the end of the Solander Trough, and, like it, lies between the Macquarie Ridge and the Campbell Plateau. Its floor lies in 4,500–4,750 m, some 500–600 m shoaler than the floor of the Southwestern Pacific Basin, from which it is separated by a sill in 4,000 m. From its depth, the basin may be classified as typically oceanic.

SOUTHWESTERN PACIFIC BASIN

This covers a substantial part of the south-eastern part of the region. The basin floor is diversified by numerous abyssal hills and occasional seamounts. Between the hills and the continental rise flanking the Campbell Plateau is a narrow abyssal plain. The floor of the basin has a general level of 5,100 m.

TASMAN BASIN

A small part of the western area is occupied by the Tasman Basin. As there are very few echo-soundings from this region the morphology of the basin floor is not known in any detail.

BATHYMETRIC DATA

Bathymetric data have been compiled on a 1 : 1,000,000 scale and isobaths drawn at 250-m intervals. Data in the N.Z.O.I. collections are chiefly complete echo-sounding profiles from which depths are read at approximately 15-minute intervals (2.5 sea miles apart at a ship speed of 10 knots). Data supplied by the U.S. Oceanographic Office and other overseas sources are only in the form of depth records at specific time intervals taken from profiles to which the N.Z.O.I. does not readily have access.

The correction for the draught of the vessel and the Matthews correction for variations in the velocity of sound through sea water are applied to all soundings. Navigational positioning in this region is often inaccurate owing to cloudiness and poor visibility. The resulting uncertainty in the data is ± 5 miles, but sometimes as little as ± 2 miles. Intersections of complementary sounding traverses enable checks and correlations to be made.

Since the standard echo-sounder can record reflections through a 30° cone about the vertical, the average echo-sounding profile is rarely a true picture of the configuration of the sea bed. However, the divergence is insignificant unless very detailed studies are undertaken (Krause, 1962). Echo-sounding profiles have been chosen to illustrate different aspects of the morphology (Figs. 2a, 10d) and are referred to by the appropriate profile number when used in the text. Discussion of morphology (Appendix IV) is based entirely on charts and profiles.

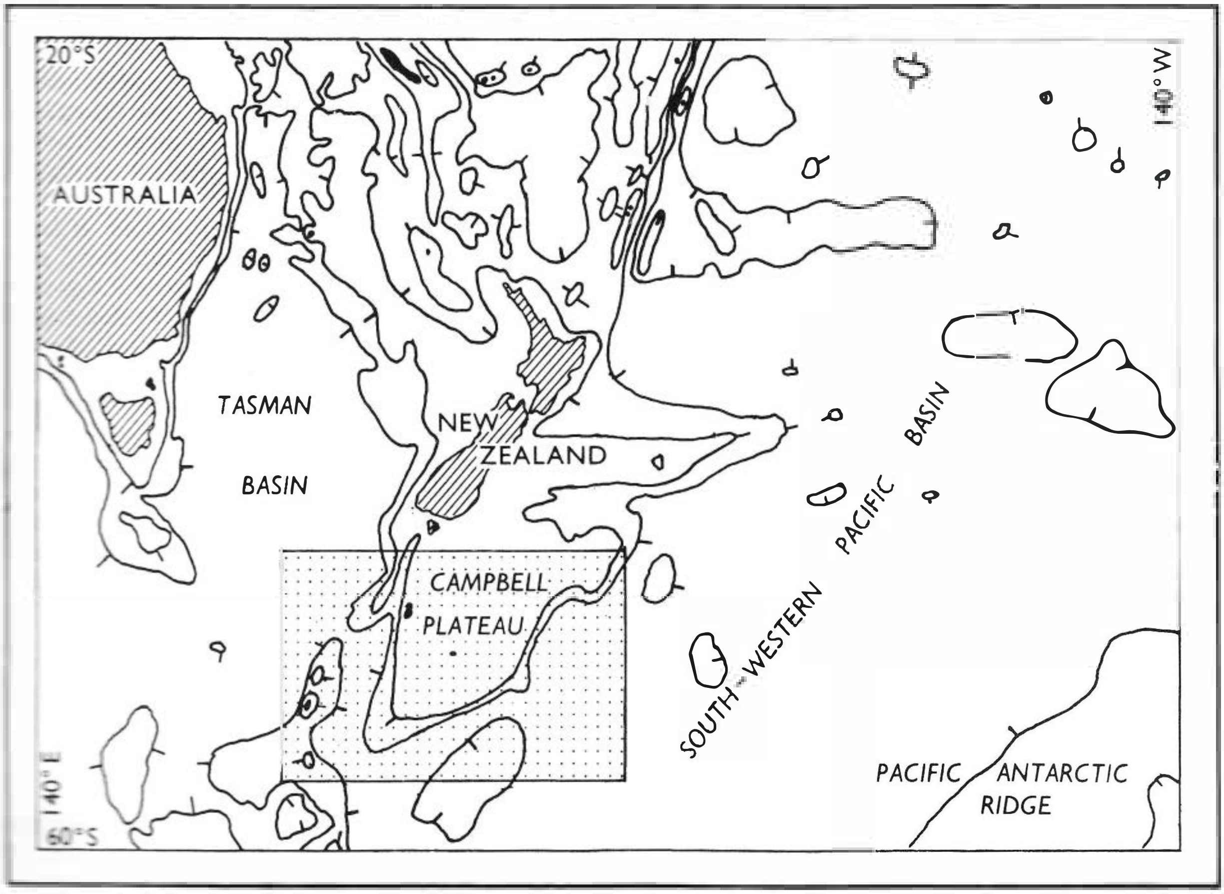

Fig. 1: Area studied (stippled). Contours in 2,000 m intervals.

GEOLOGY

REVIEW OF GEOLOGICAL HISTORY OF THE SUBANTARCTIC ISLANDS

The Subantarctic Islands south of New Zealand provide most of the data to which the geology of the surrounding sea floor may be related (Table 2). For this reason the geological history of each island will be briefly summarised and an attempt made later to correlate these histories in the light of the development of sea-floor structures.

AUCKLAND ISLANDS

(Ferrar, 1907; Speight and Finlayson, 1909; Marshall, 1912; Fleming, 1959, 1965, and in press; Wright, 1966.)

The Auckland Islands are some 200 miles south of Stewart Island at 51° 30′ S, 165° 50′ E. They are the remains of two submaturely dissected volcanic cones (Fig. 2, based on Fleming 1965 and Summerhayes 1967). Marine and glacial erosion have resulted in the development of cliffs up to 1,400 ft high on the west coast and drowned U-shaped valleys on the east coast of the main island. The stratigraphic sequence (Table 2) has been derived from a study of the pre-Pliocene rocks exposed in Carnley Harbour, the core of one of the island's volcanoes.

Biotite granite apparently forms the oldest rock on the island and, like the olivine gabbro that intrudes it, is considered Paleozoic (Speight and Finlayson, 1909). Trachyte dikes penetrate both granite and gabbro, and tuffs and flows of the same material overlie the granite. The age of these rocks is not known, but pebbles and boulders of trachyte, granite, and gabbro, with gneiss and mica schist, are found in the probably Cretaceous Camp Cove conglomerate. Marine sandstones originally thought to be Oligocene (Fleming, 1959) but now thought mid to late Miocene (Fleming, in press) are also found in Carnley Harbour. The major part of the islands consists of the remains of two volcanoes, probably Pliocene, of basalts, dolerites, and related pyroclastics, which are in turn intruded by acid dykes. Recent studies of the volcanic history of the Auckland Islands by Wright (1966) indicate that an extrusive sequence of alkali basanites, mugearites, trachytes, and comendites, followed by basanitic olivine basalts, form the Pliocene volcanics. Chemical analyses of rocks collected at the Auckland Islands by Speight and Finlayson (1909) are recorded in Table 3 (Appendix I).

CAMPBELL ISLAND

(Speight, 1905; Marshall, 1909; Oliver, Finlay, and Fleming, 1950; Fleming, 1962.)

Lying some 400 miles south of New Zealand, at 52° 30′ S, 169° 10′ E, this island is geologically similar to the Auckland Islands, consisting mainly of a much dissected volcanic cone. The most detailed account of the stratigraphy (summarised in Table 2) is by Oliver, Finlay, and Fleming (1950). Basement rocks consist of micaceous schists of the Complex Point Series. The series is not stratigraphically datable, but analogies have been drawn between these and the early Mesozoic or late Paleozoic Otago schists and the Paleozoic schists of northern Stewart Island and Fiordland (Fleming, 1962). The oldest overlying sediments, unconformable with the schists, are Upper Cretaceous quartz conglomerates and carbonaceous mudstones, which were deposited in fresh or brackish water. Conformable with these is a series of Lower Eocene to Middle Oligocene argillaceous limestones comparable to the Amuri Limestone of the New Zealand Eocene. Lower Pliocene tuffs, fossiliferous breccias, and lignites follow these, and their deposition was succeeded by a period of volcanic activity during which a basalt and andesite cone was formed. The Menhir Gabbro is thought by Oliver to be part of the pre-Cretaceous undermass upfaulted during the Pliocene. Chemical analyses on some basic igneous rocks from Campbell Island (Marshall, 1909) are recorded in Table 3 (Appendix I).

BOUNTY ISLANDS

(Speight and Finlayson, 1909)

This group of nine islands, with an average altitude of only 200 ft, is some 450 miles east of Stewart Island at 47° 42′ S, 179° 00′ E. They are not volcanic and are composed entirely of biotite granite. No extensive study of the islands has been made, but the position of the granite in the stratigraphic column has been fixed (Wasserburg *et al*., 1963) at 190 million years. This was thought to indicate Middle to late Triassic emplacement of the granite. However, on the Phanerozoic time scale (Geological Society of London, 1964) 190 million years indicates early Jurassic.

Cullen (1967) reports the occurrence of greywacke on the sea floor around the Bounty Islands.

Antipodes Islands

(Speight and Finlayson, 1909; Marshall, 1912; Fleming, 1959)

The Antipodes Islands are 420 miles south-east of Dunedin at 49° 40′ S, 178° 50′ E. The main island is a rough undulating plateau, up to 1,320 ft, surrounded by sheer, rocky cliffs. The first detailed geological study of the islands is at present being undertaken by Dr D. J. Cullen of the N.Z. Oceanographic Institute. Chief feature of the main island is its well preserved volcanic character, there being at least five cones in a good state of preservation. Scoriaceous and glassy basalts are prominent, and bedded tuffs and lavas are well exposed in coastal sections. Olivine basalt was recorded by Speight and Finlayson (1909). Owing to the good preservation of volcanic cones visible along the west coast of the main island the volcanic activity is thought to be late Tertiary or Quaternary.

Snares Islands

(Marshall, 1909; Fleming, Reed, and Harris, 1953)

This small group of islands 62 miles south-west of Stewart Island at 48° 00′ S, 166° 35′ E, lies on the continental shelf of the New Zealand mainland. They are formed from a gneissic muscovite granite, which is apparently in contact with mica schists. Their geology is similar to that of the Stewart Island - Fiordland complex and, by analogy, they are probably late Paleozoic.

Macquarie Island

(Mawson and Blake, 1943)

Macquarie Island is the only emergent part of the Macquarie Ridge, which stretches from the New Zealand mainland off Puysegur Point to about 60° S (Brodie and Dawson, 1965). It is some 570 miles south-south-west of Stewart Island at 54° 30·7′ S, 158° 57′ E (the location of the

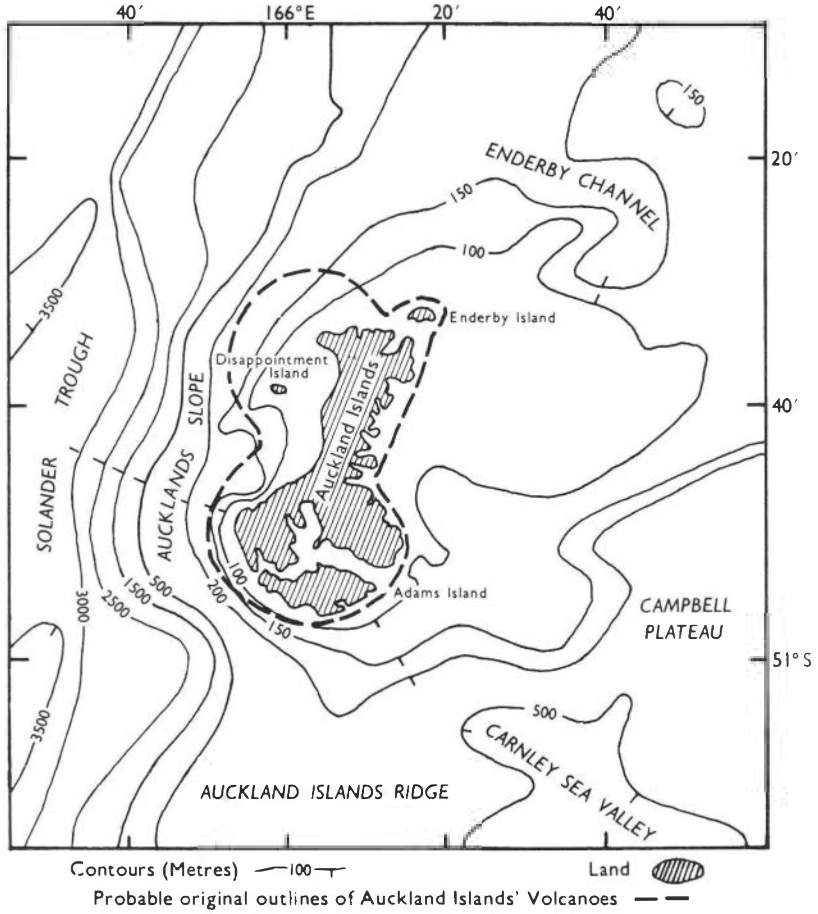

Fig. 2: Possible original dimensions of Auckland Islands volcanic centres (derived from Fleming, 1965, fig. 1) in relation to local bathymetry (derived from Summerhayes, 1967, fig. 1).

magnetic station determined in 1911; Mawson and Blake, 1943).

The present configuration of the island, elongate north–south, is thought to be due to structural control by longitudinal faults on the east and west coasts. This deduction is based on the occurrence of crush breccias and slickensided fragments on both coasts. The frequency and intensity of earthquakes indicate that fault movements are still in progress (Mawson and Blake, 1943).

An Older Basic Group of possibly Mesozoic dolerite lavas, tuffs, and tuffaceous sediments has been strongly folded and contemporaneously intruded by gabbroid masses. There followed a period of subaerial erosion, possibly late Cretaceous or early Tertiary, which uncovered the gabbroic rocks. Unconformable with these are the interbedded basaltic tuffs, lavas, and agglomerates of the Younger Basic Group. The younger lavas are mainly pillow lavas associated with globigerinal limestone of Miocene age similar, except in age, to the Campbell Island limestones. The similarity between the geology of the two islands has been pointed out by Oliver (Oliver, Finlay, and Fleming, 1950).

The surface of the island is mantled with till or boulder clay that appears to be locally derived. Four erratics, two sandstone and two granite boulders, were recovered from the beaches by Blake (Mawson and Blake, 1943). The sandstones were mainly quartz, with subsidiary decomposed feldspars. The granites retain characteristics typical of cataclasis. These rocks are possibly ice-rafted erratics from Antarctica.

The basic igneous rocks of both Older and Younger Basic Series were considered by Mawson and Blake (1943) to be calc-alkaline but are herein regarded as typically tholeiitic. Small amounts of alkaline rocks and a few more extreme alkali varieties within the Younger Basic Series also occur on the island.

Chemical analyses of selected rocks from Macquarie Island are presented in Table 4 (Appendix I).

DISCUSSION OF SUBMARINE ROCK SAMPLES

Fifty-five dredge stations yielded rock samples (Fig. 3). Rock was obtained mainly from the Macquarie Ridge, the island shelves and banks on the Campbell Plateau, and the continental shelf of New Zealand. Association of rock with shelves, banks, and ridges is to be expected, since these are typically rocky areas and areas of low sedimentation commonly swept by strong currents. Most of the shelves have been formed by the Pleistocene erosion of bedrock, which consequently may crop out on the shelf. Criteria for determining the probability of a sample being *in situ* on the sea floor can be listed (compare Emery, 1960). Some or all of them are applicable to the autochthonous rock samples listed in Appendix II.

Outcrops of bedrock may be indicated by:

1. freshly broken rock samples;
2. anchoring of the ship due to the sampler catching on rock outcrops during dredging;
3. a strong pull on the cable due to the sampler catching on rock outcrops during dredging;
4. bottom photographs;
5. the rugged appearance of the sea bed on the echo-sounding trace.

A local source may be indicated by:

6. abundant rocks of related lithology;
7. angular fragments;
8. fragile or poorly consolidated rocks.

An assumption that material is exotic must be based to a large extent on criteria involving gravity. Angular samples could be derived from a nearby talus slope flanking a submarine scarp, bank, or other rocky elevation, but where such a feature is not developed this mode of origin is inapplicable. If there is no evidence for the local occurrence of lithologies similar to the exotic fragments, then rafting, either by kelp holdfasts or iceberg drift, must be the gravitational factor.

The Campbell Plateau and Macquarie Ridge may be dumping grounds for ice-rafted glacial erratics from Antarctica. Glacial erratics have been recorded from the Chatham Rise further north (Cullen, 1965), which is approximately the present northern limit of Antarctic iceberg drift (U.S. Navy Hydrographic Office, 1957). Cullen (1962) has pointed out that during the lowered sea levels of the Pleistocene stranding of drifting icebergs may have occurred on the shoaler parts of the Chatham Rise. His arguments apply equally to the Campbell Plateau and Macquarie Ridge.

General characteristics of each rock sample are given in Appendix II. Paleontological analyses of limestones are given in Appendix III.

AUTOCHTHONOUS ROCKS FROM MACQUARIE RIDGE

Eighteen dredge stations on Macquarie Ridge yielded rock samples (Fig. 3). Dominant *in situ* rock types are volcanic or plutonic, basic igneous rocks including olivine basalts, basalts, dolerites, gabbros, peridotites, basic agglomerates, and undifferentiated basic volcanics (Table 5 and Appendix II). Other *in situ* rocks sampled were

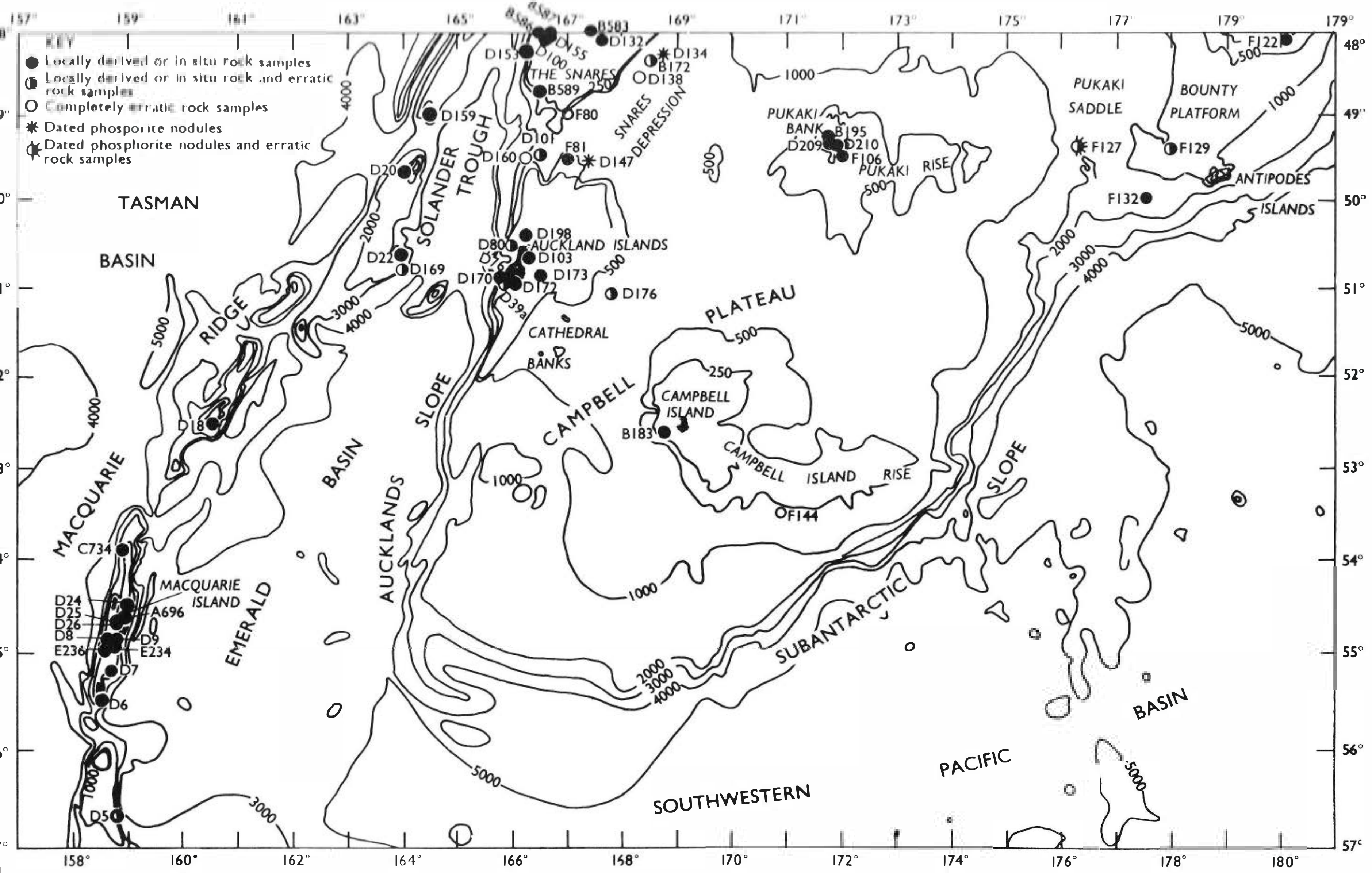

Fig. 3: Rock-sample distribution.

a friable volcanic ash (sta. D 5) and angular samples of polyzoan limestone (sta. D 159). Exotic fragments of acid igneous, metamorphic, and sedimentary rocks form subsamples at two sites on the ridge, D 5 and D 169.

Chemical analyses have been carried out on samples D 7, D 17, D 18, and D 6 by Mr J. Ritchie of the D.S.I.R. Chemistry Division, (Tables 4 and 6). Normative analyses have been computed for each sample (Table 6). Glassy olivine basalt samples D 17 and D 18 contain normative olivine and hypersthene. These analyses compare favourably with the modal analyses (Appendix II), although D 17 has no modal hypersthene.

In general, the modal minerals of submarine basalts from the Macquarie Ridge are augite, plagioclase (An_{50}), and iron oxides, with occasional small amounts of hypersthene and/or olivine. This mineral assemblage is typical of tholeiites as defined by Yoder and Tilley (1962, p. 353). It is considered that those basalts (D 6, D 20, C 734) and olivine basalts (D 6, D 18, D 169) containing modal hypersthene, and D 17, containing normative hypersthene, are tholeiitic basalts.

Basalt with a slightly different character occurs as subsamples at D 26, one containing phenocrysts of andesine (An_{32}) and the other with andesine (An_{40-50}). These contain no hypersthene and may be basalts with alkaline affinities. Sample D 7, an olivine basalt containing modal (?) analcime, also contains normative nepheline and so is an alkali basalt (compare Yoder and Tilley, 1962). The normative plagioclase of this sample is An_{45}, a calcic andesine.

The occurrence of palagonite and variolitic structure in many of the basalts shows them to be probably submarine extrusives. For further discussion of the petrography and petrochemistry of Macquarie Ridge igneous rocks see pp. 41–3.

Examination of the Foraminifera from friable angular fragments of polyzoan and foraminiferal limestone from D 159 (Hornibrook, Appendix III) indicates a probable lower Opoitian age. Small, subrounded fragments of basic volcanics within the limestone may have been derived by the action of currents on local volcanic outcrops. *In situ* fragments of bedded volcanic ash and microbreccia at sta. D 5 contain many planktonic Foraminifera, which indicate that volcanic eruption was probably Pliocene (Hornibrook, Appendix III). Olivine basalt fragments and large (2–3 mm) crystals of plagioclase (An_{60-70}) are the major constituents of the volcanic debris. As the volcanic rocks of Macquarie Island are characterised by similar bytownitic plagioclase, this part of the Macquarie Ridge (190 km to the south) evidently had a volcanic history similar to that of the island.

AUTOCHTHONOUS ROCKS FROM CAMPBELL PLATEAU

1. The Auckland Islands region (Table 5). Eight dredge stations from the shelf and Inner Aucklands Slope near the Auckland Islands yielded basic volcanics, chiefly olivine basalts (Appendix II). Most samples are lithologically homogeneous and consist of large angular and subangular fragments, probably locally derived.

A scarcity of plagioclase in samples such as D 198A suggests basanitic affinities. Olivine basalts forming the shelf around the islands are similar to late-stage olivine basalts from the islands, although the latter tend to be feldsparphyric with rare phenocrystic olivine. The absence of modal hypersthene from sea-floor samples makes them distinctly different from the tholeiitic basalts of the Macquarie Ridge discussed in the foregoing section.

Polyzoan and shell fragmental limestones were dredged from D 39 (30%) and D 173 (100%). The samples from D 173 (141 m) were freshly broken and *in situ*. Those from D 39 (549 m) are similar in lithology to D 173 and were probably deposited in a similar environment. If originally deposited on the shelf they may have reached their present site by earthquake-induced slumping or a similar mechanism. In view of the angularity of sample the source was probably local.

2. Campbell Island region (Table 5). Olivine basalt samples were dredged from Perseverance Harbour and from the shelf west of Campbell Island. They are closely similar to the Auckland Islands Shelf samples and contain olivine, augite, plagioclase, and iron ores. Sample B 183 is a large subangular pebble that probably has not undergone much transport and may reflect the westward extension of Campbell Island lava flows.

3. Pukaki Bank (Table 5). Four sites, D 209, D 210, B 195, and F 106, yielded glassy basalt and olivine basalt samples. Those from D 209 were freshly broken and are thought to have been *in situ*. A core from F 106 contains angular volcanic fragments to some depth, indicating deposition during a period of erosion or volcanic eruption. Since the angular and subangular samples obtained at B 195 and D 210 are also basic volcanics they are thought to be close to their source.

The palagonitic character of samples from B 195 and D 209 is taken to indicate submarine eruption. Olivine basalt from B 195 has a "shell" of well indurated foraminiferal limestone containing small angular fragments of altered volcanic minerals. It is inferred that the limestone was deposited during volcanic ash showers, after which it was indurated by the submarine extrusion of an olivine basalt lava.

Sample F 106 contains small, angular fragments of basic volcanics and friable foraminiferal limestones, which are both probably near their source area.

4. *Western Campbell Plateau* (Table 5). A large sample of mainly angular fragments of assorted basic volcanics, including olivine basalts, was dredged from station D 176, between the Auckland Islands and Campbell Island. Its lithological homogeneity and the angularity of fragments suggest that the sample is probably *in situ* or not far from its original source. The sample site lies on the line of the Cathedral Banks, a series of volcaniform banks some 10 miles further south. Thus these rocks are possibly directly related to some nearby source on the surface of the Campbell Plateau. The lineation of the Cathedral Banks volcanic centres indicates their probable control by some major crustal fracture, which may have resulted in the formation of volcanic deposits along this line as far north as D 176. The distance of this sample site from the Auckland Islands and the angularity of the samples suggest that the volcanic rocks are unlikely to have come from the Auckland Islands, 110 km to the west.

Sample D 176c contains small conspicuous patches of intergranular zeolites within the groundmass. It has been suggested by Wright, who found similar zeolites in some of his Auckland Island samples (Wright, 1966, p. 266), that this may reflect a basanitic character. If this is so, and the arguments for a local source tenable, a close petrographic relation is indicated between the Auckland Islands and the volcanic centre responsible for formation of sample D 176.

The agglomerate contains fragments of basic volcanics in a calcite matrix. The calcite is considered to be a recrystallised, fine-grained limestone and the agglomerate originally a submarine deposit. The palagonitic character of glassy fragments within the agglomerate is in accord with a submarine eruption and deposition in a submarine environment.

In situ fragments of Miocene calcareous quartzose conglomerate (Appendices II and III) were dredged from station F 81 on the slope between the Auckland Islands Shelf and the Campbell Plateau north of the Auckland Islands (Fig 3). It consists of rounded pebbles of calcareous sandstone packed in a slightly sandy, well indurated and slightly glauconitic, foraminiferal limestone matrix. It is likely that the limestone was originally a foraminiferal ooze very similar to those being deposited today in the Snares Depression (pp. 23–6).

Small angular pebbles of friable foraminiferal and bryozoan limestone obtained at station D 101 on the Campbell Plateau near the foot of the Auckland Islands slope contain fresh, and apparently limonitised, glauconite and are thus similar to the present local foraminiferal oozes.

5. *Bounty Platform* (Table 5). Well indurated, moderately hard, foraminiferal limestone was obtained in 252 m off the southern edge of the Bounty Islands shelf at station F 122. As it is lithologically homogeneous, with subangular fragments, and similar in lithology to limestone presently forming in this locality the sample was probably locally derived. The foraminiferal fauna of this limestone indicates a Miocene - Pliocene age (Hornibrook, Appendix III).

Further north, greywacke crops out on the sea floor around the Bounty Islands (Cullen, 1967).

Small fragments of basic volcanics, thought to reflect the character of local sediments before nodule formation, were collected from the core of a manganese nodule at station F 132, south-west of the Antipodes. Volcanic fragments indicate local volcanism, not necessarily related to the Antipodes Islands volcanic centre (compare Appendix II).

6. *Snares Depression* (Table 5). Fresh, subrounded pebbles of soft, friable, foraminiferal and polyzoan limestone were dredged at station B 172.

AUTOCHTHONOUS ROCKS FROM THE NEW ZEALAND CONTINENTAL SHELF (Table 5)

In the immediate vicinity of the Snares Islands samples of gneissic biotite granite (D 155, D 100), muscovite granite (B 586), undifferentiated granite (B 587), and mica schist (D 155, D 100) were dredged. It appears that the Snares muscovite granite (Fleming, Reed, and Harris, 1953) extends west of the islands as far as B 586 but is replaced by a biotite granite to the south (D 155, D 100). The mica schist recorded from these islands definitely extends to the south as far as D 155 and D 100.

Further away from the islands polyzoan and shell-fragmental limestones, containing whole mollusc shells in a matrix of comminuted polyzoan and molluscan fragments, were obtained. A Pleistocene fauna has been obtained from D 132 and D 153 (Hornibrook and Maxwell, Appendix III).

FOSSILIFEROUS PHOSPHORITE NODULES

Phosphorite nodules from the Campbell Plateau are either phosphatised foraminiferal oozes, or accretionary phosphorite nodules containing small numbers of Foraminifera. The ages of Foraminifera from some nodules are as follows: F 127 (Pukaki Saddle), Upper Miocene–Pleistocene; D 147 (Snares Depression), Upper Eocene–Oligocene; D 134 (Snares Depression), Oligocene–Lower Miocene. These are thought to indicate the time of nodule formation.

The formation of the nodules (probably *in situ*) and their age are discussed in detail in a later section.

ALLOCHTHONOUS ROCKS (Table 5, Appendix I; Appendix IIa)

Several rock samples from the sea floor in this region contain fragments which, because they are lithologically anomalous (Appendix IIa), seem to have been transported from some distant source. The freshness and angularity of some samples indicate that they have not lain on the sea floor for any great length of time, another factor in favour of transportation.

Dominantly elongate, angular fragments (*see* D 5) may have been derived from joint blocks by glacial action and transported from Antarctica by drift ice. Samples of basic volcanics from stations D 138, D 160, and B 172, on the Campbell Plateau, may have been derived by kelp rafting from local volcanic islands. The source of granitic and metamorphic rocks at these and other sites is probably distant from this area, unless they were deposited during the Pleistocene erosion of local granitic and metamorphic terrain now hidden by later sediments.

The occurrence of hornblende granite at stations D 5 (on the Macquarie Ridge) and F 127 (on the Campbell Plateau) may indicate a common provenance.

REVIEW OF GEOPHYSICAL STUDIES IN THE NEW ZEALAND SUBANTARCTIC

Geophysical investigations in this area have been carried out mainly by the New Zealand D.S.I.R. Geophysics Division and Victoria University of Wellington. Magnetic gravity and seismic studies have been undertaken of parts of the Campbell Plateau, the Solander Trough, the Southwestern Pacific Basin, and the Macquarie Ridge north of 51° S.

SEISMIC STUDIES

Using surface-wave dispersion studies of earthquake records Adams (1962) determined an average crustal thickness of 17–23 km for the part of the Campbell Plateau that is shoaler than 2,000 m. This value is substantially different from the crustal thickness of 30–40 km calculated for New Zealand by Thompson and Evison (1962).

GRAVITY DATA

A positive Bouguer anomaly of 72 milligals is recorded from the Auckland Islands (Robertson, 1965). A similar value of 78 milligals, recorded from these islands by Stahl, was considered to reflect a crustal thickness of 29 km (Robertson, 1965). This represents merely the local crustal thickness, which, though it may be duplicated beneath the other islands on the plateau, is not representative of regional conditions, judged by the seismic evidence (*see above*).

A positive Bouguer gravity anomaly of 245 milligals has been observed on Macquarie Island (Robertson, 1965).

A large positive Bouguer gravity anomaly, reaching a maximum of 160 milligals near Dagg's Sound, Fiordland, trends north-east–south-west through Fiordland more or less parallel to and continuous with the Macquarie Ridge (Fig. 4).

MAGNETIC-FIELD STUDIES

Magnetic-field surveys over the Campbell Plateau (Adams and Christoffel, 1962) reveal a diverse magnetic character, the complexity of which does not permit inter-profile correlation. Anomalies are numerous, and range from small to large, and from narrow and local to broad and deep-seated.

The magnetic character of the Southwestern Pacific Basin is simple compared with that of the Campbell Plateau (Christoffel and Ross, 1965; Ross and Christoffel, in press). Large positive anomalies, with amplitudes of 200 gammas and wavelengths of 15–20 miles, form apparently linear features traceable across 10° of longitude. The ridge-trough anomaly pattern is disrupted by a

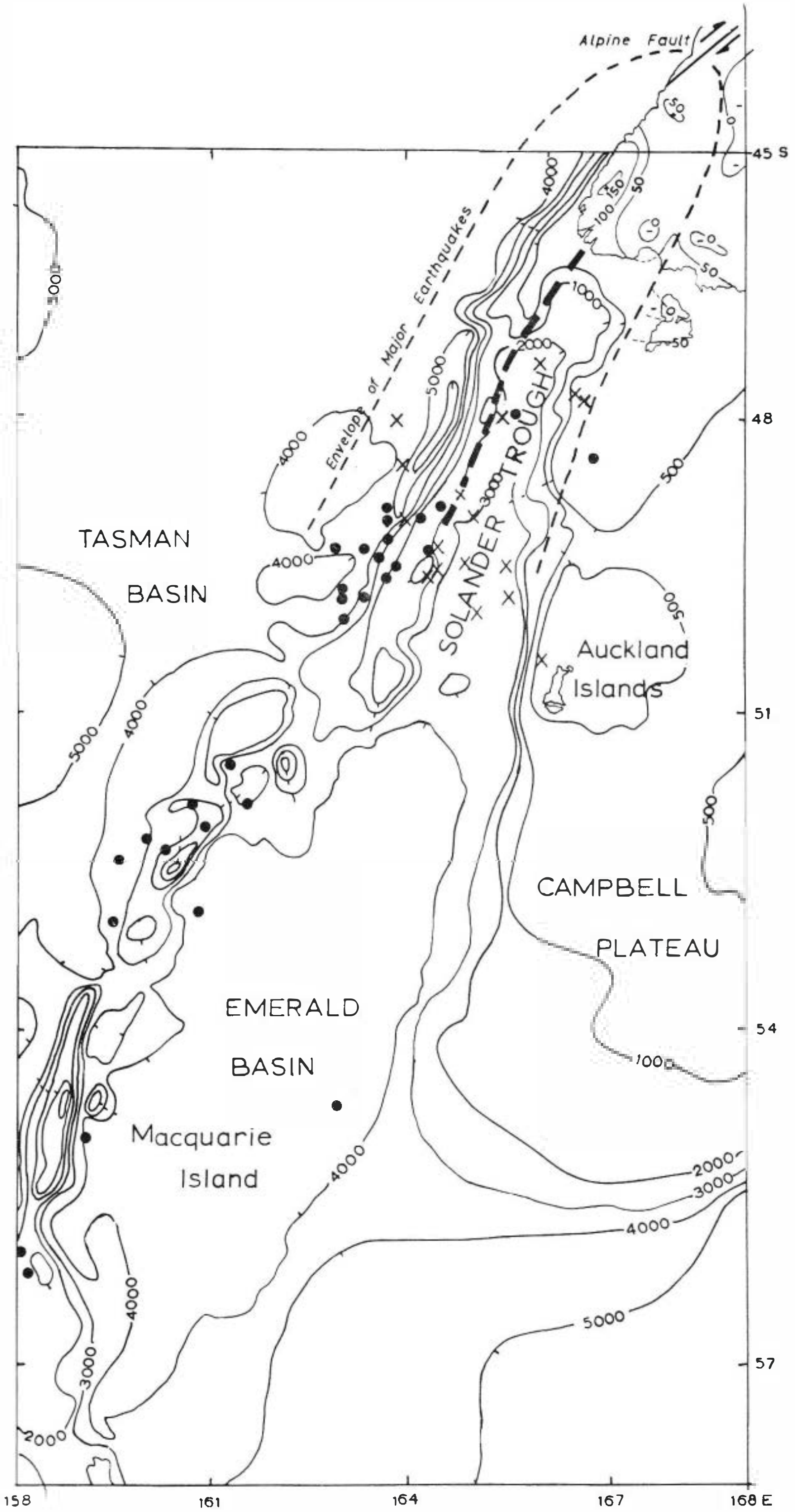

Fig. 4: Geophysical characteristics of the Macquarie Ridge and Solander Trough. Position of major magnetic anomaly (derived from Hatherton, in press); of earthquake epicentre distribution (from Cooke, 1966, and New Zealand Seismological Records, 1962–64; compare Table 7); of gravity anomalies (from Reilly, 1965).

left-lateral displacement zone trending south-south-east from around the Antipodes Islands toward the Pacific-Antarctic Ridge (Fig. 12).

At the foot of the Subantarctic Slope is a continental rise characterised by a large, broad, negative anomaly (Christoffel and Ross, 1965).

Magnetic characteristics of the Macquarie Ridge and Solander Trough, north of 51° S, have been studied by Hatherton (1967). A pronounced positive magnetic anomaly, 5–30 km wide and 500 gammas in amplitude, is closely associated with the ridge as far north as Preservation Inlet, where measurements cease. The form of the anomaly does not correlate with the ridge crest north of the Snares, where it follows the eastern flank (Fig. 4). Southward the anomaly becomes more complex and tends to correlate more closely with the crest of the ridge.

Magnetic-field profiles across the Solander Trough (Fig. 5) are generally smooth. A small magnetic high lies slightly west of the trough centre along much of its length but, at the latitude

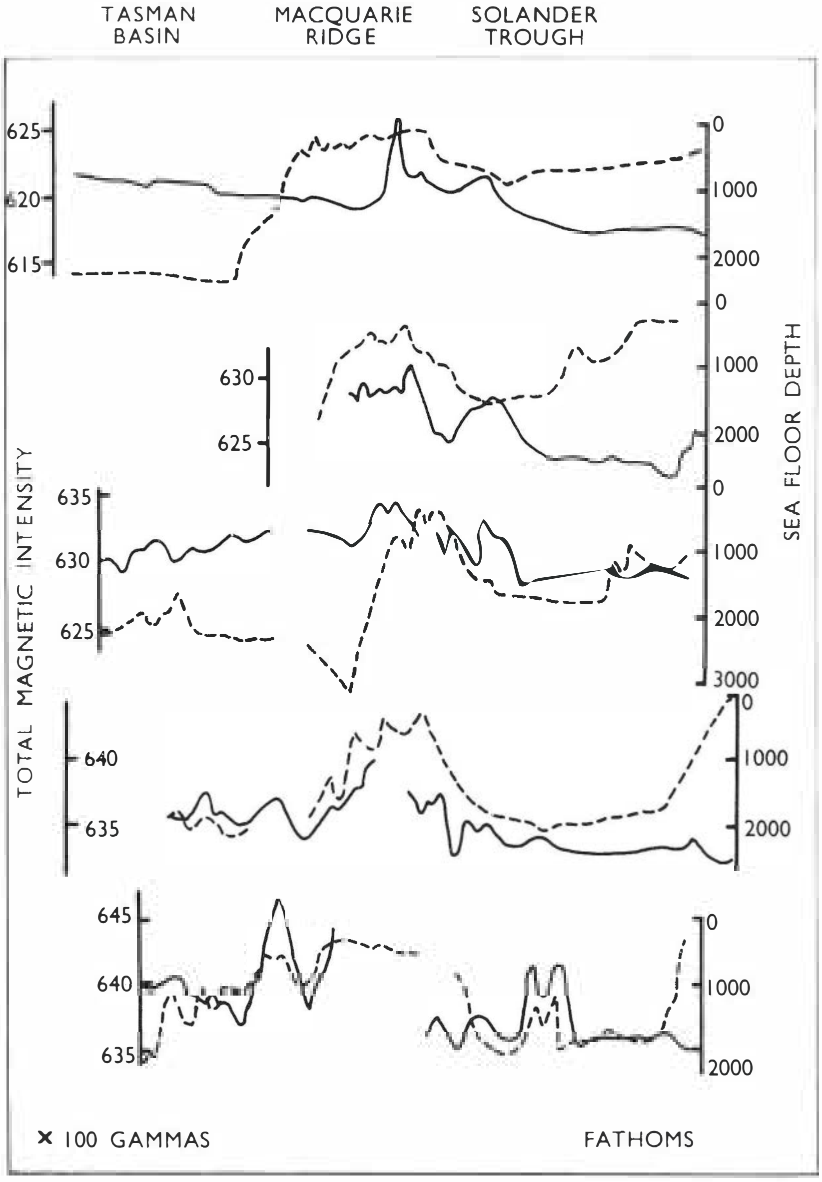

Fig. 5: Magnetic-field profiles across Macquarie Ridge and Solander Trough (from Hatherton, in press). Northernmost traverses occur at top of figure. The upper profile occurs north of the charted area; the southern four are recorded bathymetrically in Fig. 7a, b, c, d (Appendix IV). Magnetic profile = heavy line; bathymetric profile = dashed line.

of the Auckland Islands, gives way to a twin-peaked magnetic high corresponding with the twin peaks of a seamount rising from the trough floor (Fig. 5).

REGIONAL SEISMICITY

The association of shallow earthquakes with the Macquarie Ridge was first noted by Gutenberg and Richter (1954) and later by Cooke (1966). A plot of earthquake epicentres for this region (Fig. 4) has been derived from Cooke (1966) and by the writer from records in the N.Z. Seismological Observatory, 1962–64, hereafter referred to as N.Z. Seismological Records for 1962–64 (Table 7, Appendix I). Earthquake epicentres in the Solander Trough tend to be concentrated in the trough centre or along the flanks.

In southern New Zealand shallow earthquakes are concentrated in an "envelope of seismicity", a seismically active region margined eastward by the stable Campbell Plateau, Bounty Trough, and Chatham Rise, and westward by the Tasman Basin, and terminating northward at the southern end of the Alpine Fault (Fig. 4). The Macquarie Ridge - Solander Trough seismicity would seem to reflect the southward continuation of the "envelope of seismicity" of southern New Zealand (Fig. 4).

Data from New Zealand and other seismological centres indicate that the Campbell Plateau is a seismically stable unit (Eiby, 1958). One earthquake epicentre has been recorded at 50° 25′ S, 166° E on the Auckland Islands Shelf, a few miles north of the islands (Fig. 4).

RECENT SEDIMENTATION

(Sample distribution, Fig. 6A; sediment distribution, Fig. 6)

SHELF SEDIMENTS

Island shelves and the New Zealand continental shelf in this region are characterised by the accumulation of bryozoan and shell debris (Table 8, Appendix I; Fig. 6). Authigenic minerals are rare; they have been recorded from only one station (F 138) on the Campbell Island shelf.

The bryozoan and molluscan shell debris that dominate sediment on the continental shelf are of granular and coarse sand grade. Bryozoan and algal nodules, ranging in diameter from 1·5 to 0·25 in. occur quite commonly. Carbonate content of these sediments averages 90.1%.

On the Auckland Islands Shelf molluscan debris is more common, but bryozoans still contribute a large percentage of organic remains. Carbonate content averages 85·8% on the shelf above 400 m. Sediments on the Auckland Islands Shelf are slightly finer than those on the continental shelf, and medium grades are more common.

On the Campbell Island shelf the bryozoan content is substantially less, and coarse to medium molluscan shell debris is the dominant sediment component. Carbonate content averages 74·8% in depths less than 300 m.

A mixed assemblage of biogenic debris dredged from the Pukaki Bank in a depth of 82 m probably represents the remains of a rock-dwelling community.

Near the edge of the Bounty Islands shelf a sample of globigerina ooze with a carbonate content of 72·5% was obtained.

The amount of non-carbonate material in shelf sediments is low, indicating that terrigenous sediment plays very little part in shelf sedimentary processes at present. Why sediment from the Bounty Islands and Campbell Island regions should have the highest non-carbonate fraction is not fully understood. Such detrital components could well include relict sediments surviving at the levels of former low stands of sea level.

CAMPBELL PLATEAU

GENERAL DESCRIPTION (Table 8, Appendix I).

Globigerina ooze is the dominant sediment on the Campbell Plateau. Generally it is fine-grained, milky white, and homogeneous, consisting dominantly (50–60%) of Foraminifera, the remains of coccoliths, and other calcareous micro-organisms, and non-carbonate material—siliceous organisms and detrital and authigenic mineral matter.

Carbonate content averages 85·3%, ranging between 60·5% (F 117) and 99·5% (F 104). This is very similar to the world-wide range of 30·15% to 96·8%, average 64·47%, determined by Murray and Renard (1891). Variation in carbonate content is not related to depth within the depth range covered, nor is any geographical zonation of carbonate content recognised. Evidently terrigenous sediment plays a very limited role in sedimentation on the Campbell Plateau.

Quartz dominates the detrital mineral content of globigerina oozes on the plateau. Small amounts

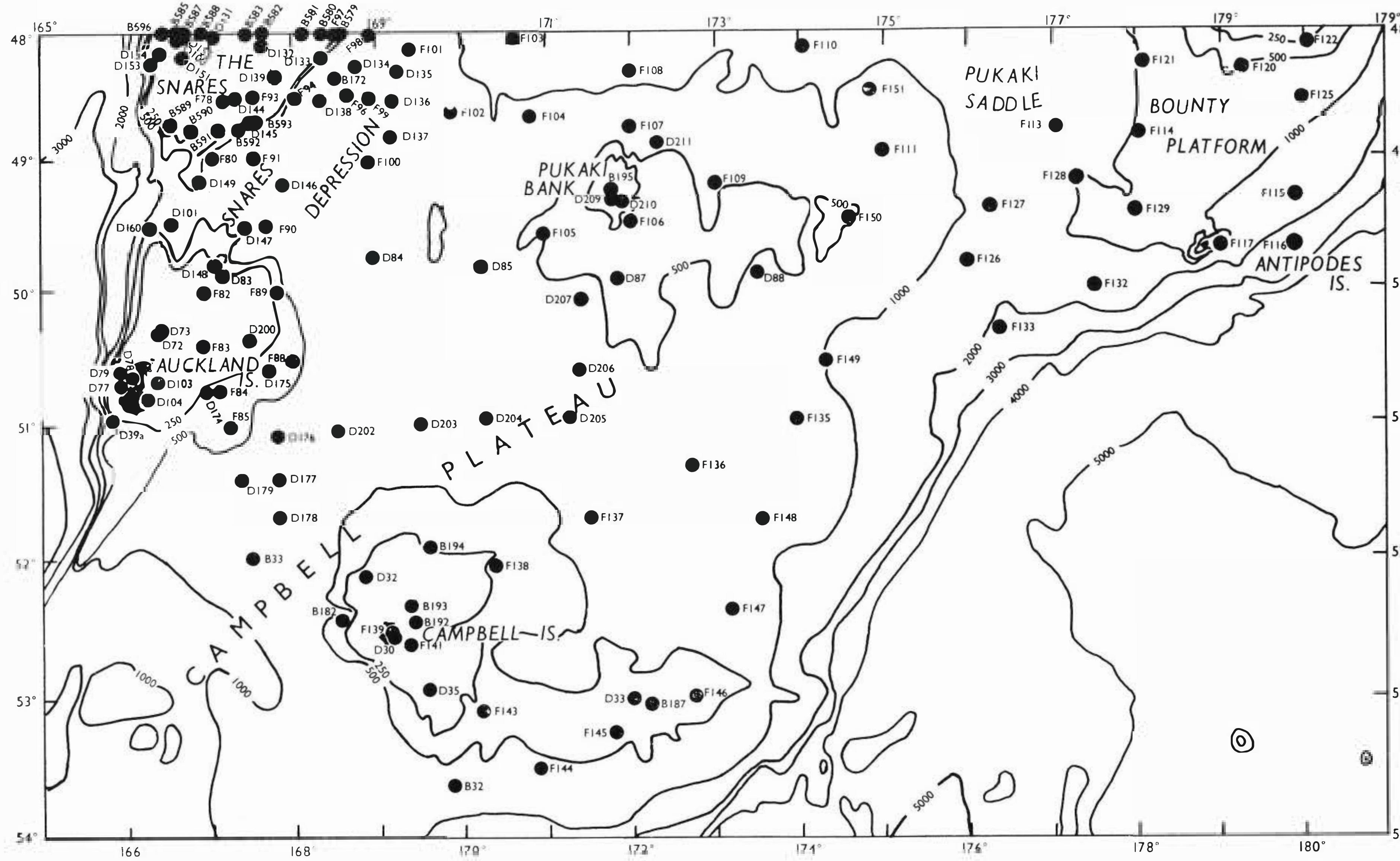

Fig. 6A: Sediment-sample sites.

of volcanic rock fragments and ferromagnesian minerals such as biotite are not uncommon. Identifiable skeletal remains of siliceous organisms are mainly those of radiolaria or sponges. Relative abundances of detrital minerals and siliceous organisms within the non-carbonate fraction are given in Table 9 (Appendix I).

It seems reasonable to infer that the Campbell Plateau is entirely mantled by globigerina oozes of the type found on the northern two-thirds of the plateau (Fig. 6). Lisitzin (1962) indicated that the sediments of the Campbell Plateau were mainly terrigenous in origin and (Lisitzin, 1962, fig. 1) graded marginally into foraminiferal oozes. This view is corrected by the substantial data now in our possession which show that the Campbell Plateau is an area of dominantly globigerina ooze.

TEXTURAL CHARACTERISTICS (Table 10)

Textural characteristics of selected sediment samples from the Campbell Plateau are determined in phi units (Inman, 1952) derived from cumulative particle distribution curves (Fig. 7). Median diameters of globigerina ooze range from 2·73 ϕ (D 175) to 4·60 ϕ (D 206) and lie in the fine sand to coarse silt grades. Dispersion ranges from 0·85 ϕ (F 111) to 2·45 ϕ (D 80) and is moderate to very poor. Skewness is positive in all the analysed samples except F 111.

Sample D 160 from the Snares Depression near the Aucklands Slope is a foraminiferal sand distinguished from the oozes by a lack of fine-grained material. It has a coarser median diameter (2·4 ϕ) and is better sorted than typical oozes (dispersion factor 0·7 ϕ). This sample reflects the winnowing action of ocean-bottom currents which have removed the finer fraction so that the dispersion factor is reduced. Foraminiferal sands also occur in the Pukaki Saddle (Fig. 6).

Analyses of non-carbonate fractions of globigerina oozes are rather difficult because of the small amount of such material in the samples available. The median diameter of the non-carbonate fraction is commonly finer than that of the overall sample, ranging from 2·58 ϕ (D 160) to greater than 4·12 ϕ (F 102). Samples F 111 and D 206 are characterised by a slightly coarser non-carbonate fraction, but this is unusual (Fig. 8).

Non-carbonate fractions have better dispersion than the total sample but it was not possible to obtain determinations of dispersion in half the samples (Table 10).

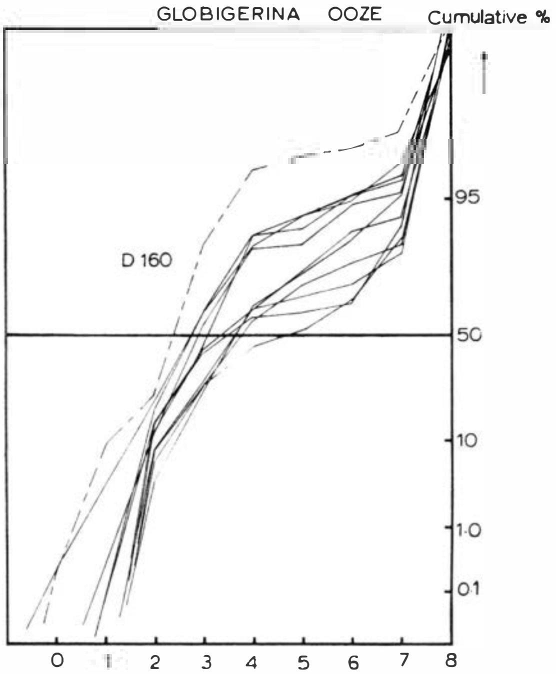

Fig. 7: Globigerina ooze; cumulative particle-distribution curves. D 160 (dashed line) is a foraminiferal sand. Sample numbers are given in Table 10.

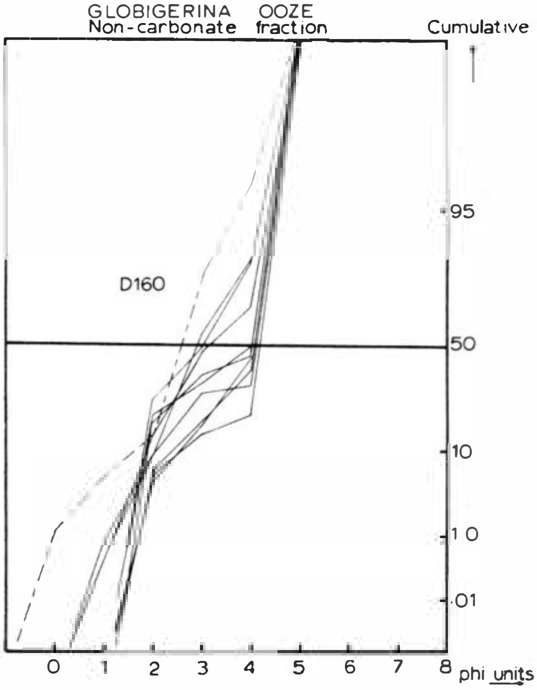

Fig. 8: Non-carbonate fraction of globigerina ooze; cumulative particle-distribution curves. D 160 is a foraminiferal sand (dashed line). Sample numbers are given in Table 10.

ACCUMULATION RATES OF CARBONATE SEDIMENTS

It is not feasible to determine a rate of sedimentation for the shelf areas as the rate of accumulation of bryozoan and shell sediments has not been studied thoroughly here or elsewhere. As these sediments occur on presumably wave-cut Pleistocene terraces at depths typical of such terraces elsewhere, it is inferred that the rate of accumulation of biogenic sediment is extremely low. Nevertheless, sufficient sediment has accumulated since the Pleistocene to cover local bedrock with a thin veneer.

If the rate of accumulation of globigerina ooze is 1 cm/1,000 years, as suggested for other regions (Mero, 1964), then some 650 m of ooze may have accumulated here since the Cretaceous. In view of the occurrence of early to mid-Tertiary phosphatised oozes on parts of the relatively flat plateau surface, this estimate may well be much too high.

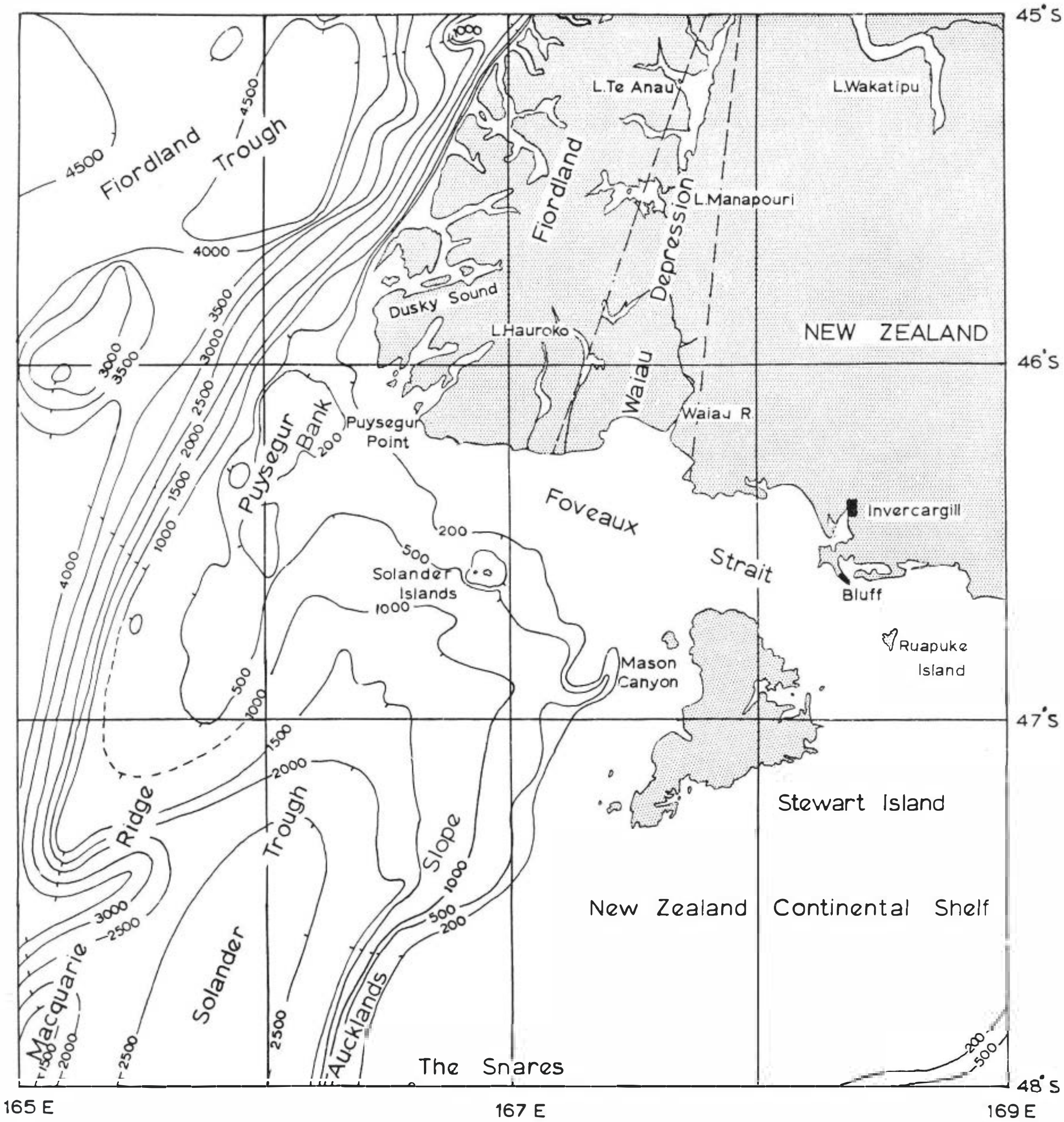

Fig. 9: Morphology of the northern sections of Macquarie Ridge and Solander Trough, between the charted area and New Zealand.

THE DEEP-SEA FLOOR

Sediment samples have not yet been obtained by the N.Z.O.I. from the deep-sea floor in this region. According to Lisitzin (1962) diatomaceous sediments occur south of the Campbell Plateau, and to west and east the deep-sea floor is characterised by the accumulation of red clay. Shelly bryozoan and foraminiferal sediments have been collected by the N.Z.O.I. from the shoaler parts of the Macquarie Ridge (Table 8, Appendix I).

A recognisable continental rise, abyssal plain, and abyssal hill province lies south of the Campbell Plateau and covers parts of the Southwestern Pacific Basin (Fig. 6). Both continental rise and abyssal plain are believed to be continental margin depositional features and may be characterised by slightly different sedimentary facies.

Sediments of the abyssal hill province are most probably formed by a steady shower of pelagic organic remains and suspended terrigenous particles. The roughness of topography in the abyssal hill province suggests a thin sediment cover. Sediments in these southern latitudes are generally diatomaceous (Lisitzin, 1962) and the deep-sea sediments of the abyssal hill province will probably be of this type.

Within the Solander Trough - Emerald Basin system two sedimentary environments are recognised (Fig. 6). The trough is partly formed by the coalescence of continental rises mantling the Auckland Slope and Macquarie Ridge. Its main character is its gentle southward gradient and a median channel that is probably a sediment movement channel. These are typical of continental rises, and it is inferred that the Solander Trough is floored by a continental rise formed of sediment derived chiefly from the New Zealand mainland (pp. 46–7). Evidence for supply of sediment from the Campbell Plateau is very limited although during the Pleistocene erosion of the Auckland Islands Shelf a large volume would have been available. Mason Canyon, off Stewart Island, may act as a main feeder for the Solander Trough (Fig. 9).

The flat floor of the Emerald Basin is an abyssal plain, which is probably supplied with sediment from New Zealand via the Solander Trough (p. 47). The Solander Trough (continental rise) and Emerald Basin (abyssal plain) are similar in physiography, and probably in origin, to the continental rise and abyssal plain at the foot of the Subantarctic Slope and are thus mapped as similar features (Fig. 6).

Little is known of the morphology of the Tasman Basin floor. The existence of a narrow trench between the Macquarie Ridge and the basin precludes the ridge and southern New Zealand as direct sources of sediments. It is suggested that pelagic deposits will be the main sediment type in this region. In accord with the findings of Lisitzin these sediments may be pelagic red clays (compare Lisitzin, 1962, fig. 1).

AUTHIGENIC MINERALS

GLAUCONITE

DESCRIPTION AND DISTRIBUTION

Green glauconite grains appear in several samples of globigerina ooze from the Campbell Plateau (Table 8, Appendix I). Small amounts of this glauconite are smooth surfaced, commonly botryoidal, dark green grains with syneresis cracks. These grains may have originated by glauconitisation of faecal pellets or ferromagnesian mineral grains.

Most glauconite occurs, however, as pale green internal casts of Foraminifera, reflecting development within micro-environments provided by the empty chambers of foraminifers. Internal casts of Foraminifera may be white, pale green, and pale brown. Complete colour gradations between them are evident, indicating a close relation. Pale green casts are probably the most glauconitic, closely resembling glauconitic foraminiferal casts from off the east coast of the North Island of New Zealand (compare Pantin, 1966). Brown casts may be caused by the presence of authigenic limonite, a mineral formed elsewhere in the New Zealand region in close association with glauconite. Pantin (1966) has indicated that similar chemical conditions control formation of these two minerals.

White casts may represent compacted clay minerals which have not yet been glauconitised. Where these occur glauconitisation must be fairly slow.

The relative abundance of detrital minerals, siliceous organisms, and the two main forms of glauconite have been qualitatively assessed in selected samples (Table 9, Appendix I). Those samples in which glauconite is visible are mapped as glauconitic globigerina oozes and, although the glauconite fraction rarely exceeds 2% of the total sample (visual estimate), zones of glauconitic ooze may be roughly delineated (Fig. 6).

One major zone of glauconitic globigerina ooze is the Snares Depression, beyond which glauconitic oozes stretch south along the slope bounding the Auckland Islands Shelf. Other major zones are the Pukaki Bank, the eastern end of the Campbell Island Rise, and the south-western part of the Bounty Platform.

ORIGIN OF GLAUCONITE ON THE CAMPBELL PLATEAU

Glauconite is an authigenic mineral; its presence indicates that sediment is deposited very slowly or not at all (Cloud, 1955). Typically glauconite is found on topographic highs such as the Chatham Rise (Norris, 1964) and the Lachlan Ridge (Pantin, 1966), regions where current activity is vigorous enough to inhibit sedimentation. Occurrences of glauconite on the Campbell Island Rise, the Pukaki Bank, and along the slope marginal to the Auckland Islands Shelf are thus quite normal since these rises are probably subject to fairly vigorous current activity because of their elevation. A more thorough investigation is necessary in the Snares Depression and on the edge of the Pukaki Saddle, where glauconite occurs in topographic depressions. The occurrence of winnowed foraminiferal sands in the Pukaki Saddle and in the Snares Depression (D 160) indicates the presence of currents stronger than those crossing the rest of the plateau. Other authigenic minerals are also found in these environments, which are traversed by channels in the sea floor perhaps cut by sediment-laden currents. Localisation of glauconitic deposits in depressions where sea-floor channels occur is taken to indicate locally high current velocities which inhibit sedimentation and allow glauconite formation (see also the section on phosphorite nodules).

Conditions essential for the formation of glauconite in micro-environments such as those provided by tests of Foraminifera have been extensively discussed by Pantin (1966) and are briefly reiterated below. Iron occurs both in ferric and ferrous states, indicating an environment of low Eh. Supplies of terrigenous and autochthonous organic matter to the area must also be very low, resulting in the retreat of iron-fixing bacteria or other organisms into available micro-environments, such as foram tests, where organic concentrations tend to be high. Vigorous bottom currents are commonly associated with the slow sedimentation which is a prerequisite for glauconite formation. However, these cannot be too vigorous as too strong a current will result in destruction of available micro-environments (Pantin, 1966).

It is concluded that the presence and development of glauconite on restricted parts of the Campbell Plateau reflect a low supply of terrigenous and autochthonous organic material, a relatively high bottom-current velocity, and slow rate of sedimentation. The Snares Depression and the Pukaki Saddle are probably the sites of fairly vigorous ocean-bottom currents. This hypothesis is substantiated for the Snares Depression, along which streams a northwardly directed arm of the Circumpolar Current and the southwardly directed Bounty - Campbell Gyral (compare Burling, 1961).

COMPARISON WITH NEW ZEALAND GLAUCONITE DEPOSITS

Norris (1964, p. 24) has pointed out that the present geologic and oceanographic setting of the Chatham Rise is a useful analogue in reconstruction of early Tertiary ecology and paleogeography of the New Zealand region. Many of the environmental parameters governing formation of rich Tertiary greensand deposits of New Zealand must have been similar to those operating on the Chatham Rise today. Campbell Plateau glauconitic globigerina oozes contain much less glauconite than those on the Chatham Rise. This type of poorly glauconitic ooze is, however, closely similar to parts of the early Tertiary McDonald Limestone, especially to the slightly glauconitic Flat Top Limestone that occurs in the Waitaki subdivision (Gage, 1957).

MANGANESE NODULES

DESCRIPTION AND DISTRIBUTION

Black, hydrous, manganese dioxide concretions are distributed on the deep-ocean floor as grains, nodules, slabs, rock coatings, and impregnations and as other less common forms (Mero, 1964). Well developed manganese nodules are found at station D 5 on the Macquarie Ridge and stations F 116, F 127, F 129, and F 132 on the Campbell Plateau (Table 8). Manganese-coated slabs of rock are found at station D 169 on the Macquarie Ridge and a small fragment of thin manganese crust, scraped probably from a submarine rock outcrop, was obtained on the Campbell Plateau at station F 133. Distribution of manganese nodules in the charted area is shown in Fig. 6. The main areas of concentration are the Macquarie Ridge and the Pukaki Saddle.

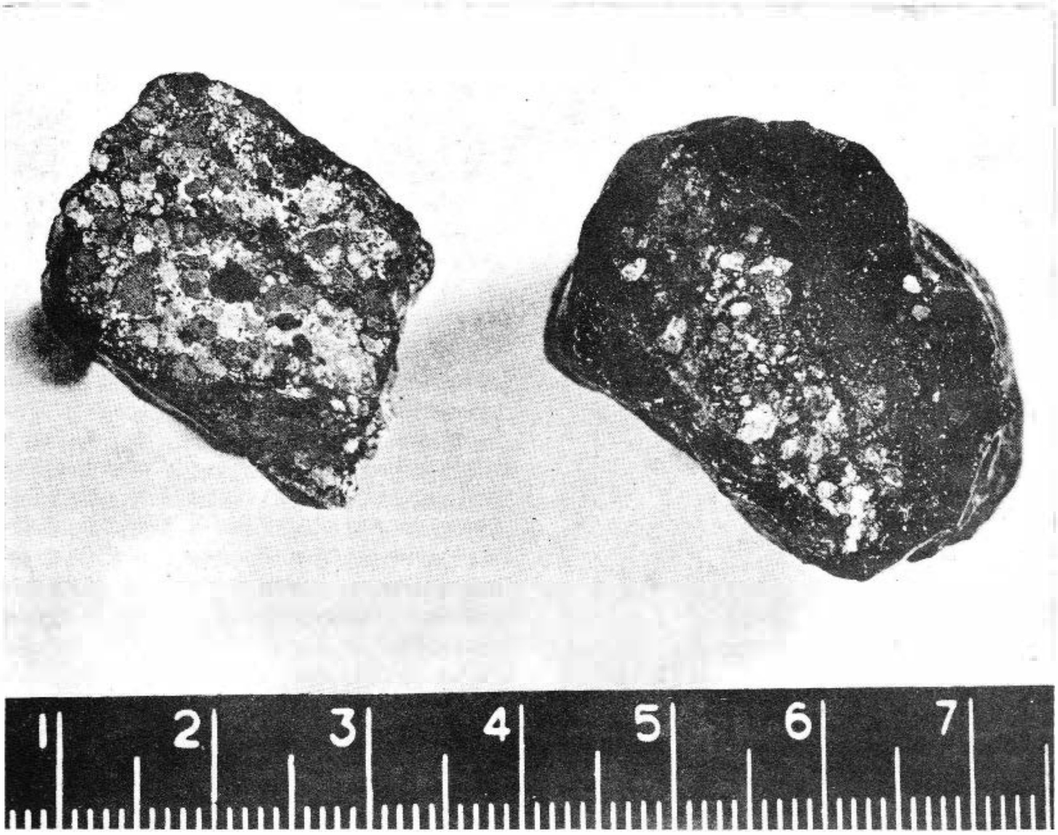

Fig. 10: Section through manganese nodule from station D 5, showing volcanic micro-agglomerate nucleus, thickness of manganiferous crust, and incipient layering within the crust. Scale is in cm.

Manganese nodules are earthy black, sometimes with a reddish or purplish tinge. Incipient layering in which black manganiferous bands are separated by narrow brownish bands is typical (Fig. 10). Brown layers may reflect zones of iron concentration or layers of fine-grained deep-sea sediment incorporated during a pause in manganese accretion (Mero, 1964). Manganiferous bands are up to 0·6 mm thick. Nodule size ranges from about 1 cm (F 116) to 6·5 × 5·5 × 4·5 cm (F 127). The larger nodules consist of a crust of manganese minerals around rock fragments. The manganese crust may be up to 2 cm (D 5) thick.

Examination of the nuclei of nodules from D 5, F 127, F 129, and F 132 provided interesting data on the development of manganese nodules in this region. At station D 5 the nuclei (Fig. 10) are of a slightly calcareous, ashy agglomerate of Pliocene age (p. 18). Evidently a bedded ash deposit was broken up by submarine erosion or tectonic activity before nodule formation. If it is assumed that the maximum of 2 cm of manganese crust has accumulated since the Pliocene, the rate of manganese accretion is 1 mm/100,000 years, slower than the 0·1 mm/1,000 years assumed by Mero (1964) for Pacific nodules.

At station F 127 manganese nodules and phosphorite nodules occur together. Sections through the manganese nodules show that they are formed around previously existing phosphorite nodules (Fig. 11). Although the manganiferous crust reaches 2 cm it is not possible to estimate the rate of formation of these nodules as the age of the phosphatised limestone nucleus is not known more accurately than upper Miocene to Pleistocene. Evidently, since the formation of phosphorite nodules, physico- or bio-chemical conditions favouring the formation of phosphorite have given way to those favouring deposition of manganese minerals.

Similarly, at station F 129 about 45% of the sample consists of manganese minerals which encrust and join together yellowish brown phosphorite nodules. A change in conditions governing authigenic mineral deposition is again evident.

Station F 132 yielded manganese nodules with a crust up to 2 cm thick around a nucleus of fragments of volcanic rocks, biogenic rocks, mineral

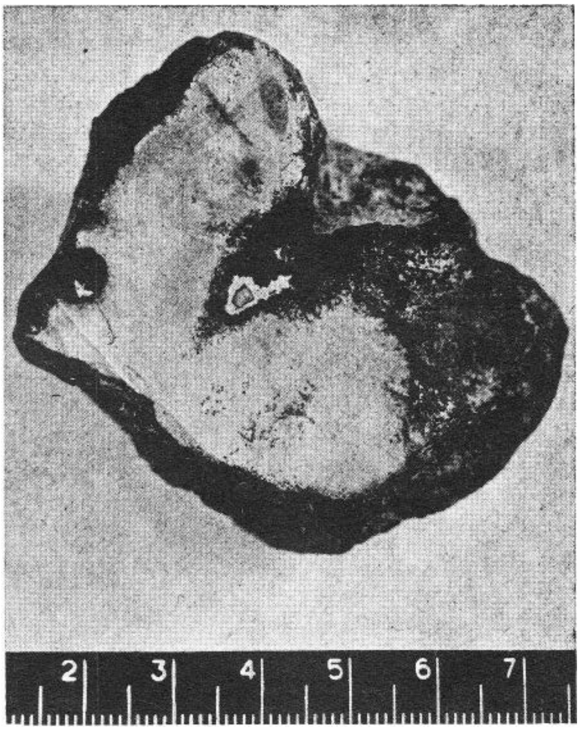

Fig. 11: Section through manganese nodule from station F 127, showing phosphatised foraminiferal ooze (phosphorite nodule) nucleus, covered by manganiferous crust reaching a thickness of 0·5 cm. Scale is in cm.

grains, and bryozoan skeletons (Appendix II) which probably represent the original deep-sea sediment. One extremely interesting feature of this sample is the replacement of bryozoan skeletal material by manganese minerals, apparently by some form of molecule by molecule substitution.

Nodules from station D 5 are smooth and ovoid (Fig. 10) with bryozoa and small coral growths on all sides, indicating that they have rolled on the sea floor. Roundness is well developed also in nodules from F 116, but phosphorite nodule nucleii from F 127 and F 129 have irregular contours.

CHEMICAL ANALYSIS

A sample of black, manganese mineral from station F 129 on the Campbell Plateau was sent for analysis to D.S.I.R. Chemistry Division. Results (Table 11) indicate that there is very little difference in composition between this and typical manganese nodules (Table 11). Actual compositional differences do not appear to reflect any significant difference in overall chemistry.

ORIGIN

Arrhenius and Bonatti (1965) have shown that manganese in manganese nodules may be derived either by direct precipitation from sea water or by precipitation from sea water saturated with volcanic effusives from submarine eruptions. They find that the cobalt content, co-precipitated with manganese, may be "used as a criterion for the solution history and source of the manganese". A high cobalt content (Mn/Co ratio less than 300) is taken to indicate rapid precipitation from liquids saturated with "volcanic solutions". If the ratio is greater than 300 slow precipitation from sea water is indicated. This theory is thought to be borne out by the occurrence of high Mn/Co ratios around the margins of the Pacific in non-volcanic areas (Arrhenius and Bonatti, 1965). Since sample F 129 has a Mn/Co ratio of 31·3 it is possible that its manganese was deposited from solution during a phase of local submarine volcanic activity. This may explain the localised occurrence of manganese nodules near the Antipodes Islands and on the Macquarie Ridge, areas of known volcanism. It also explains why manganese nodules are not found in the Snares Depression (a non-volcanic region) although this is an area of slow sediment accumulation in which precipitation of authigenic manganese minerals might otherwise occur. Association of manganese nodules with volcanicity also explains why manganese nodules are formed around mid-Tertiary phosphorite nodules, since volcanicity on the ridge and plateau is late Tertiary or Quaternary. A volcanic origin for manganese in these regions is also in accord with the theory of Bonatti and Nayudu (1965) that all manganese nodules are derived by precipitation from "volcanic solution".

PHOSPHORITE NODULES

DESCRIPTION AND DISTRIBUTION

Phosphorite is an authigenic mineral formed on the sea floor, where, typically, it occurs in nodule form. It contains 20–30% P_2O_5 and consists chiefly of collophane, a carbonate fluorapatite (Dietz, Emery, and Shepard, 1942).

Occurrences of phosphorite nodules off shore in the New Zealand region have been recorded previously by Reed and Hornibrook (1952) and by Norris (1964) on the Chatham Rise, a topographic high, the surface of which is likely to be an oxidising environment.

Phosphorite nodules are found on the Campbell Plateau in slight topographic depressions, in the Snares Depression, and on the Pukaki Saddle (Fig. 6; Table 8). These are zones in which sediment deposition is slow or lacking, as indicated also by the occurrences of glauconite in both areas and of manganese nodules in the Pukaki Saddle.

Age and Origin

Age of Foraminifera in nodules ranges from Upper Eocene to Lower Miocene in the Snares Depression, and from Upper Miocene to Pleistocene (a tentative age) in the Pukaki Saddle (Appendix III). Nodules from D 147 in the Snares Depression are accretionary, and in their nucleii Upper Eocene to Oligocene Foraminifera are recognised (Appendix III). In contrast, D 134, F 127, and F 122 are phosphatised foraminiferal oozes, the ages of which may not necessarily be the age of phosphatisation. Phosphatised oozes of different ages are here preserved on the sea floor in areas where sedimentation is slow or non-existent. Because these nodules are uncontaminated with younger material, it is reasonable to suppose that phosphatisation was either contemporaneous with formation of the ooze, or occurred shortly after.

Phosphorite nodules on the Chatham Rise are phosphatised Miocene foraminiferal oozes (Reed and Hornibrook, 1952). The mid-Tertiary ages of these off-shore phosphorite nodules are similar to early and mid-Tertiary "phosphatic nodules" and phosphatised limestone found in the Waitaki district of the South Island (Gage, 1957). In studying the off-shore distribution of phosphorite nodules it may be significant that only mid-Tertiary nodules are found in areas which are non-depositional at present.

Changes in environmental chemistry since formation of phosphorite nodules are indicated from stations F 127 and F 129, where manganese crusts hide the original phosphorite (Fig. 11), and at D 134 and F 90, where phosphorite nodules are coated with glauconite 5 mm thick.

Theories on the origin of phosphorite nodules have been reviewed by Emery (1960) and Mero (1964). Conditions favouring phosphorite formation are: (1) a probable oxidising environment; (2) the presence of cold nutrient-rich currents; (3) slow or negligible sedimentation (Emery, 1960). The last condition is believed to have been stable here at least since the early Tertiary. Geological evidence indicates mid-Tertiary development of rises on the Campbell Plateau, which may have caused environmental changes leading to the formation of phosphorite. However, the general mid-Tertiary age limit to nodule formation is more probably due to some change in chemical conditions affecting the New Zealand region as a whole.

Recent work by I. Devereux* on oxygen isotopes in Tertiary sedimentary rocks from New Zealand indicates mid-Tertiary temperature rises of about 6–8° C (pers. comm.). It is here suggested that physico- or biochemical conditions governing increases in phosphate precipitation may be sufficiently sensitive to temperature changes to have allowed phosphorite-nodule formation from the warmer waters of the mid-Tertiary. This hypothesis is paralleled by the experimental findings of Clark and Turner (1955) that rate of precipitation of phosphate increases with temperature.

Volcanic activity resulting in manganese-nodule formation prevented phosphorite-nodule formation in the Pukaki Saddle during the later Tertiary or Quaternary, but there is no evidence to show that phosphorite precipitation continued following the cessation of volcanism. In the Snares Depression, glauconite, not phosphorite, is the presently forming authigenic mineral. The hypothesis that phosphorite precipitation here is temperature controlled assumes that oxidising, nutrient-rich ocean currents have been common to this area throughout Tertiary and Recent times.

Chemical Analyses

Chemical analyses of phosphorite nodules were carried out by the D.S.I.R. Chemistry Division (Table 12). Nodules from the Campbell Plateau are similar in composition to those from the Chatham Rise but contain slightly less P_2O_5 than samples from the California Borderland (Table 12). Phosphatisation appears to be fairly uniform as a chemical process although one sample of foraminiferal limestone from F 122 has been only partly phosphatised and contains only 10% P_2O_5.

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*Institute of Nuclear Sciences.

GEOLOGY OF MAJOR STRUCTURES IN THE SUBANTARCTIC REGION

New Zealand Continental Shelf

The occurrence of biotite granite in the vicinity of the Snares Islands, which consist mainly of muscovite granite, expands our knowledge of the lithology of the southern continental shelf of New Zealand. Similar biotite granite also occurs at the Auckland and Bounty Islands.

The continental shelf is interpreted as a Pleistocene marine erosion surface. At present it is covered by sediments consisting of polyzoan and shell fragments. An indurated molluscan and polyzoan limestone occurs at some localities on the shelf. The age of limestone from D 132 has been determined as lower Pleistocene (Hornibrook and Maxwell, Appendix III). It is inferred that sediments similar to present-day sediments were deposited on the shelf during the Pleistocene and have since been indurated. Their occurrence as indurated rock outcrops may be due to the removal of Recent material by current action. A similar limestone dredged from D 153, on the Inner Aucklands Slope, was formed about this time, possibly during the Nukumaruan (Maxwell, Appendix III).

Campbell Plateau

The Campbell Plateau forms a large, upstanding part of the continental borderland of New Zealand. From seismic studies it is deduced that the plateau has a crust of intermediate thickness. Geological observations made on the islands rising from the plateau indicate that this crust is probably composed of sedimentary, igneous, and metamorphic rocks.

The predominantly smooth and generally level surface of the plateau is diversified by a number of broad, shallow rises and depressions, which may reflect something of the underlying geological structure (Fig. 12). Interpretation of these depends on the nature and formation of the plateau surface. This may be:

1. A primary feature caused by marine erosion and planation to base level as suggested by Fleming (1962);
2. The result of the filling up of depressions in an originally diverse topography with sediments supplied by turbidity currents from the New Zealand mainland; or
3. The result of prolonged pelagic sedimentation on a previously diverse topography.

Thus large-scale rises might be due to the warping of an erosion surface or to the existence of relict topographic highs separated by basins filled to a common level by some sedimentary process. These points will be closely considered below.

Like much of the New Zealand mainland, the Campbell Plateau was a land mass of very low relief during the late Cretaceous (Fleming, 1962). Peneplanation, together with marine erosion by the transgressive seas of the early Tertiary, may have entirely or nearly completed levelling of the plateau. Pelagic sediments, such as those forming the early Tertiary foraminiferal limestones of Campbell Island, are still accumulating on the plateau and may have smoothed out topographic irregularities.

Occurrences of phosphatised early Tertiary oozes on level parts of the plateau suggest that most of the levelling occurred during the pre-Tertiary and is not due to sedimentation. Pleistocene erosion of shelf regions probably did not contribute much sediment to the plateau.

Bounty Platform

If the Campbell Plateau is assumed to have been originally more or less level over the whole region the present 400 m discrepancy in level between the flat parts of the Bounty Platform and the western plateau surface must be explained in terms of faulting or folding. The western margin of the Bounty Platform is a regular feature with north-south trend. Along this margin the platform level drops toward the Pukaki Saddle in two horizontal "steps" (Appendix IV). The only suitable explanation for these "steps" is that they represent downfaulted sections of an originally continuous level (Fig. 13). Since faulting disrupts the plateau level, movement must be post-Cretaceous. The plateau surface west of the Pukaki Saddle is tilted down towards the saddle, possibly about a monoclinal fold axis, again with a north-south trend (Figs. 12 and 13). The possibility that this margin of the plateau is also fault controlled cannot be ruled out.

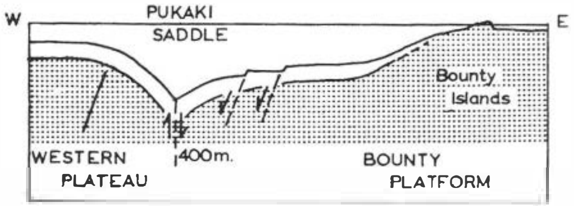


Fig. 13: Structural cross section (hypothetical) through Pukaki Saddle and the Bounty Platform, indicating probable fault lines (dashed) and fold axis (heavy line). Basement, stippled; cover rocks, blank.

A number of steep-sided, flat-topped elevations and depressions 25–50 miles north-west of the Antipodes Islands are interpreted as fault-controlled horsts and grabens. Since the horsts are flat topped and the grabens flat floored, it is likely that fault movement occurred after the formation of the flat plateau surface. Local occurrences of manganese nodules indicate submarine volcanicity, which could be associated with this block faulting. Owing to the paucity of echo-soundings from this area it is not possible to assign a trend to the faults within these zones. On each side of fault zones the depth of the platform is a fairly constant 980 m.

Within a 12-mile radius of the Antipodes Islands are numerous pinnacles, which appear to be subsidiary volcanic cones, a submarine extension of the complex of six or more small Quaternary or Recent volcanoes that form the island. Volcanic cones lying along the west coast of the main island are aligned with Bollon's Island, an outlying cone, and show a north-east trend. This probably represents the direction of a deep-seated major crustal fracture aligned with the Subantarctic Slope (Cullen, 1967).

The rise on which the Bounty Islands are sited modifies the regional level of the platform and thus may post-date Cretaceous planation. It was evidently extant at much the same level during the Pleistocene since it is "capped" by a shelf similar in depth to known Pleistocene shelves.

Miocene-Pliocene foraminiferal limestones from near the edge of the Bounty Islands shelf (Appendices II and III) indicate deposition here, during the Tertiary, of moderately deep-water pelagic sediments. Late Tertiary updoming of the Bounty Islands region is indicated, resulting in exposure of basement granite and greywacke. The limestone appears to mantle these basement rocks.

PUKAKI SADDLE

The Pukaki Saddle is a synform caused by marginal downwarping or faulting of the western plateau surface and marginal downfaulting of the Bounty Platform towards the saddle. Occurrence of Upper Miocene - Pleistocene authigenic phosphorite nodules and more recent manganese deposits within the Pukaki Saddle indicate that this is a non-depositional environment probably swept by strong currents. Assuming phosphorite precipitation to be closely related to the existence of currents, and currents to be controlled by topography, then evolution of the saddle may have been a mid to late Tertiary event, because the phosphorite nodules were formed at that time. Erosional agents may have helped to shape the saddle but are not thought to have caused it to form. Channels locally on the floor of the saddle indicate current movement, possibly of sediment-laden bottom currents.

PUKAKI RISE

The Pukaki Rise, a broad, east-west elongate feature from which rises the Pukaki Bank, lies along the northern edge of the plateau. The rise is probably a linear crustal warp, through which runs a generally definable axis of uplift (Fig. 12). By analogy with local volcanic centres Pukaki Bank volcanicity is probably late Tertiary. Uplift of the Pukaki Rise is probably genetically related to volcanicity and may therefore be late Tertiary. Small-scale block faulting is deduced from sub-bottom reflections recorded near the north-eastern end of the rise (Traverse 7a).

CAMPBELL ISLAND RISE

Like the Pukaki Rise, the Campbell Island Rise is a gentle linear feature with a broadly definable axis of uplift (Fig. 12). Campbell Island, a Tertiary volcanic centre, is sited on the dome-like western end of the rise. South-east of the island the rise has a median crest on which there are local elevations, up to depths of 260 m, which may be volcanic in origin.

The geological history of Campbell Island indicates sinking of this region, at the end of the Cretaceous, to a moderate depth at which accumulation of foraminiferal limestones could occur. Such a depth would probably be the same as that at present occupied by the greater part of the plateau. If the Miocene uplift of the island indicates wholesale uplift of the plateau a total "down-up" movement of 1,000–1,200 m is required during the early Tertiary. In view of the known stability of the plateau and the association of the Campbell Island Rise with the localised late Tertiary volcanism it is suggested that the concept of early Tertiary elevation of the whole plateau is unrealistic and that the Campbell Island Rise is probably the result of localised Miocene folding prior to late Tertiary volcanism.

Between the Campbell Island and Pukaki Rises the edge of the Campbell Plateau is downwarped about a broadly north-south axis. This appears to be continuous with the axis of warping along the western downwarped or faulted edge of the Pukaki Saddle (Fig. 12).

SNARES DEPRESSION

The general plateau level drops to about 650–730 m in the Snares Depression between the Auckland Islands Shelf and the Pukaki Rise

(Appendix IV). A deep median channel crosses the depression and extends north into the Bounty Trough (Fig. 14) where it may form part of the Bounty Trough channel system described by Krause (1966).

Glauconite and phosphorite nodules here are associated with recognised ocean currents, which may be vigorous enough to inhibit sedimentation. Since Eocene and Oligocene Foraminifera have been obtained from these phosphorite nodules, conditions may have been similar throughout most of the Tertiary and Recent. Movement of sediment-laden water or turbidity currents may be chiefly responsible for cutting the deep channel. Current winnowing of fine-grained material over parts of this region may cause formation of sediment-laden water and/or turbidity currents. Cutting of the channel at former sea levels is not feasible since there is no evidence for depression of sea level to the required depths, nor for post-Cretaceous elevation of the sea floor towards sea level. Krause (1966) reached similar conclusions as to the evolution of channels in the Bounty Trough (Fig. 14).

A system of poorly developed sea valleys along the northern plateau edge may be related to the channels in the Bounty Trough (Fig. 14).

Auckland Islands Shelf and Ridge

The Auckland Islands rise from the western edge of a broad shelf forming the northern end of the Auckland Islands Ridge. It has been suggested (Summerhayes, 1967) that the ridge is a late modification to the originally flat plateau surface. It is considered genetically related to the Tertiary volcanism of the Auckland Islands. Further paleontological evidence for the age of the ridge is derived from sample F 81 from the slope north of the Auckland Islands Shelf (Fig. 3). A calcareous conglomerate was formed during the Miocene (Appendix II) by calcareous sands slumping into slightly glauconitic foraminiferal ooze similar to that forming in the Snares Depression at present. The sands are derived from a granitic and metamorphic provenance (Appendix II). It is deduced that slumping occurred from shallower water on the then rising Auckland Islands Ridge. That the ridge was active during the Miocene is inferred from the Miocene sandstones deposited at the Auckland Islands (Fleming, in press). Exposure of deep-water limestone (F 81) on the flanks of the Auckland Islands Ridge suggests that early- to mid-Tertiary limestones are here mantling the upwarped granitic basement.

The shelf is cut on a very broad part of the ridge, which may continue northward to the New Zealand continental slope in much reduced form. The ridge has a broadly definable axis of uplift with a northerly trend reflected in the north-south alignment of the two volcanic domes of the Auckland Islands (Figs. 2 and 12).

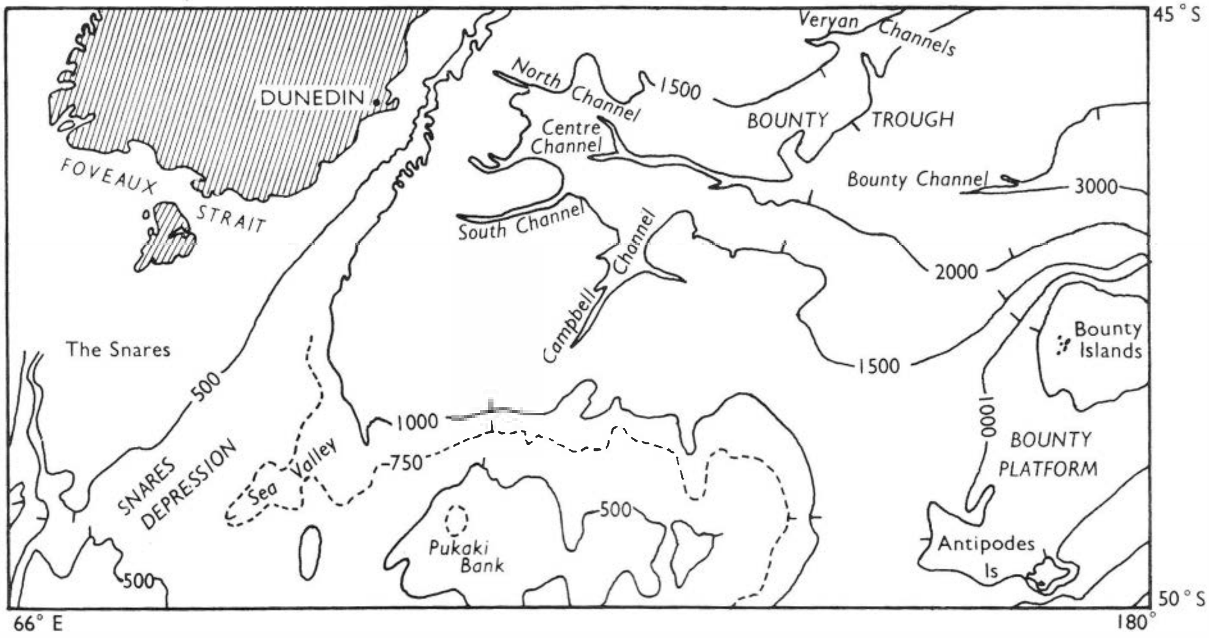


Fig. 14: Sea-floor channels and sea valleys east of New Zealand. Bathymetry of Bounty Trough from Krause (1966); of northern Campbell Plateau, from this work.

CATHEDRAL DEPRESSION

Alignment of the eastern wall of the Cathedral Depression and the Cathedral Banks may indicate continuity of a line of structural fracture or crustal discontinuity. The depression is a sharply bounded, narrow trough, the steep walls of which are probably controlled by normal faulting (Fig. 12). Since this feature cuts across an otherwise undisturbed plateau it is probably younger than late Cretaceous. The western wall of the depression is formed by the probably Tertiary Auckland Islands Ridge.

CATHEDRAL BANKS

The isolated and steep-sided elevations forming the banks are morphologically similar to pinnacles surrounding the Antipodes Islands and are probably submarine volcanic peaks. Since they follow a well defined north-north-east trend for 50 miles, they probably lie along a line of crustal weakness, possibly a continuation of the faults forming the eastern wall of the Cathedral Depression (Fig. 12). The plateau retains constant level on each side of this zone indicating the cessation of the normal fault movement evident further south. Cathedral Banks volcanism may be the same age as the Cathedral Depression and the Auckland Islands Ridge, late Tertiary, and thus is coeval with volcanic activity elsewhere on the plateau.

SOUTH-WESTERN PLATEAU SURFACE

The south-western plateau surface is tilted gently toward the Emerald Basin. Its outer edge is relieved by several flat-topped elevations (Fig. 12), which are probably tectonically controlled since they disturb the regional plateau level. These are interpreted as fault blocks.

SHELF AREA

A world-wide datum for the lowest stand of Pleistocene sea level is generally accepted at about 130 m (Shepard, 1963). Variation in the depth of terraces on the shelf around the Auckland Islands has been taken to indicate Pleistocene warping contemporaneous with terrace formation (Summerhayes, 1967). The shelf is warped about a north-south axis continuous with the Auckland Islands Ridge (Fig. 12).

Similar downwarping toward the Aucklands Slope is evident on the western edge of the continental shelf off the Snares. Here the western edge of the shelf is at a maximum recorded depth of 190 m, whereas the shelf edge elsewhere is at 150 m.

The shelf around Campbell Island is downtilted to the north-east about an axis along its south-western edge (Fig. 12). A maximum depth of 236 m is recorded north-east of the island, the shoalest depth being 183 m, north-west.

The shelf edge around the Bounty Islands varies between recorded extremes of 163 and 205 m. It is quite evidently warped about an approximately north-south axis corresponding with the anticlinal axis shown in Fig. 12.

Recorded depths for the shelf edge over the whole of this region lie deeper than Shepard's world-wide datum (1963).

On each of the shelves considered the shelf edge lies at variable depths indicating post-Pleistocene tilting or warping, so Shepard's datum is not likely to apply to present shelf-edge depths on the Campbell Plateau. The restricted parts of the Campbell Plateau on which the shelves are sited evidently were recently active. If Shepard's 130 m datum is accepted the sinking of the Campbell Plateau island shelves may reflect either sinking of the plateau as a whole or merely collapse of the rises on which the shelves are situated. In view of the seismic and probable isostatic stability of the plateau and the known localisation of volcanic activity to rises, it may be concluded that the sinking reflected on individual shelves is purely a local phenomenon restricted to the tectonically active ridges that carry the shelves. Depths similar to that cited by Shepard (1963) are typical of much of the east coast of New Zealand (N.Z.O.I. collections), although between Cape Palliser and Gable End Foreland the depth of the shelf edge is variable, and usually greater than 130 m (Pantin, 1963).

CRUSTAL CHARACTER OF THE CAMPBELL PLATEAU

The average thickness of the crust beneath the plateau, 17–23 km (Adams, 1962), contrasts with the average crustal thickness of 33 km (Worzel and Shurbet, 1955). Locally, beneath the Subantarctic Islands, the crust may be up to 29 km thick. It is substantially thicker than the Tasman Sea floor and northern and southern Pacific Basins, which are calculated to be 5–10 km thick (Officer, 1955). It is comparable in thickness to the Lord Howe Rise, 20 km (Officer, 1955), the 18–31 km of the California Borderland (Menard, 1964), and the 14·5 km of the Blake Plateau off the Florida coast (Hersey, Bunce, Wyrick, and Dietz, 1959). These features resemble the Campbell Plateau in extent, depth, and (especially the Blake Plateau) in character. A generalised picture of the distribution of crustal thickness in the New Zealand region is given in Fig. 15 (modified from Brodie, 1964, fig. 20).

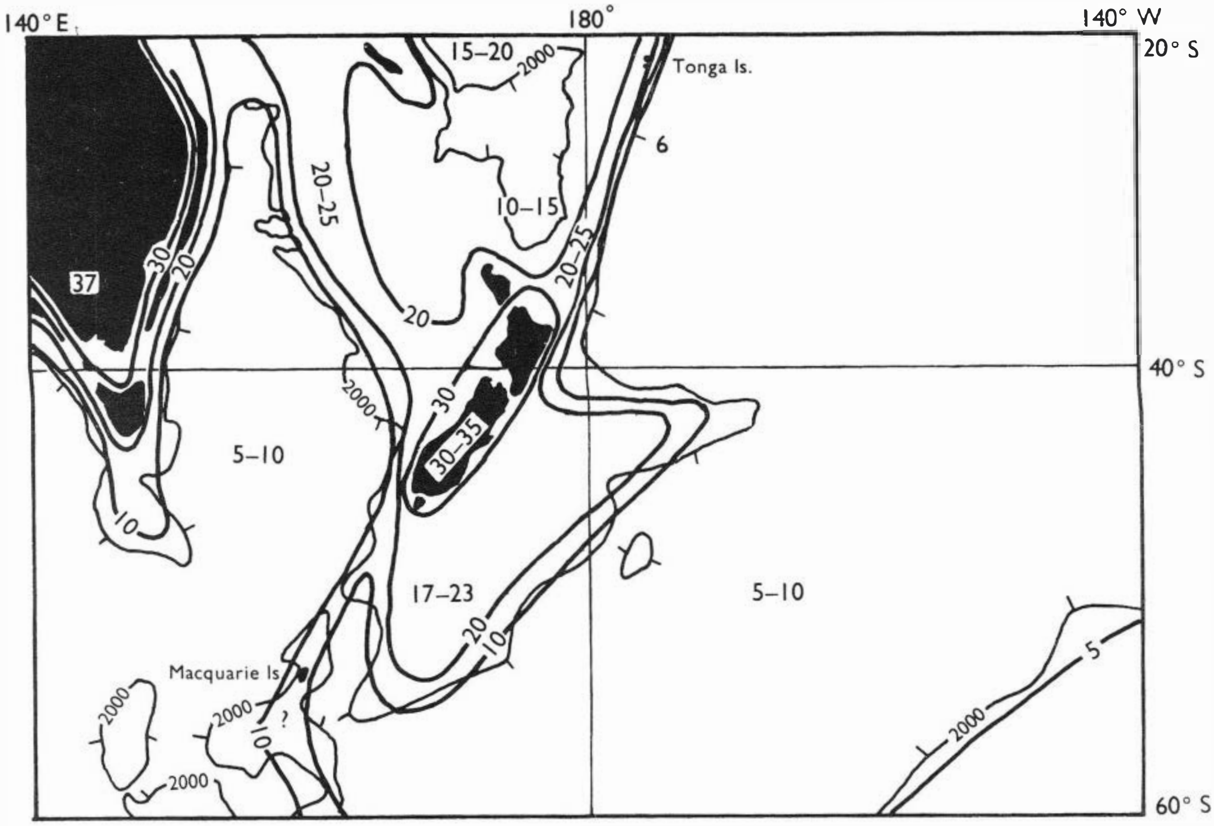


Fig. 15: Distribution of crustal thickness (km) in the south-west Pacific (modified from Brodie, 1964, fig. 20, and including a newly determined value of less than 5 km for the East Pacific Rise, derived from Menard, 1964).

Adams (1962) concluded that a thick series of indurated sediments and granitic rocks must be present within the confines of the Campbell Plateau. Magnetic data may be interpreted to indicate the intrusion of these rocks by basic igneous magma (Adams and Christoffel, 1962). The 17–23 km crustal thickness approaches the theoretical requirements for isostatic equilibrium between the Campbell Plateau and the surrounding deep-ocean floor (Adams, 1962).

VOLCANISM ON THE CAMPBELL PLATEAU

Alkaline volcanism was widespread during the Pliocene, when tuffs were deposited at Campbell Island, followed by the formation of a basalt and andesite cone and, at the Auckland Islands, an extrusive basalt complex. The fresh appearance of the basalt volcanic cones of the Antipodes Islands indicates that they are probably Quaternary and, since there is no shelf around the islands, they may be Recent.

By analogy with local centres, and because Pukaki and Cathedral Banks interrupt the probable "erosion level" forming the plateau surface, their volcanism was probably late Tertiary. Olivine basalts from Pukaki Bank and the Antipodes Islands have not been chemically analysed and it is not known whether they are alkaline. Basalts with alkaline affinities are found at station D 176 near Cathedral Banks; thus this centre may be assumed to be alkaline. Since Pukaki Bank and the Antipodes Islands occupy similar structural positions to other centres on the plateau, their volcanism too may be alkaline.

The south-east of the South Island of New Zealand is a major volcanic province distinct from the North Island volcanic centres though similarly of mid-Tertiary to Quaternary age (Benson, 1941; Grindley and Harrington, 1957). This province is characterised by olivine basalts with widespread alkaline affinities and by limited amounts of andesite and tholeiite (Gregg and Coombs, 1966). Selected analyses of basalts from Banks Peninsula (Gregg and Coombs, 1966) and Dunedin (Coombs, 1966) are given in Table 3.

Close geochemical similarities between the volcanic centres of the Campbell Plateau and those of the south-eastern part of the South Island (Table 3) suggest that their rocks form a single petrographic province, primarily alkaline and basaltic, separate and distinct from the Macquarie Tholeiite Province. The two provinces are of similar age but differ in volcanic character and structure.

The Subantarctic - South Island Alkali Province also includes the alkaline Chatham Islands volcanics. Despite their distance apart, these centres have similar petrographic character, a common location in more or less stable structural regions, and comparable age, which falls within the Kaikoura Orogeny.

Summary of the Structure and Geological History of the Campbell Plateau (Figs. 12 and 16)

1. The Campbell Plateau is a thin section of continental crust, isostatically and tectonically stable and probably aseismic.
2. Levelling of the plateau, now covered by pelagic sediments, was probably by Cretaceous erosion to a base level.
3. It is diversified by broad crustal upwards (rises) capped with related Tertiary volcanic centres.
4. Granitic and metamorphic basement rocks crop out at islands on the rises and on the sea floor around the Bounty Islands.
5. The rises appear to be mantled with moderately deep water, and mid- and early-Tertiary foraminiferal limestones.
6. Uplift followed sedimentation thus—

   The Bounty Islands uplift is probably mid to late Tertiary.

   The Campbell Island uplift was Miocene (Oliver, Finlay, and Fleming, 1950).

   Uplift occurred along the Auckland Islands Ridge probably during the Miocene.
7. Pukaki Saddle is a tectonic depression across which the Campbell Plateau surface is downfaulted 400 m to the east.
8. Snares Depression and the Pukaki Saddle have been areas of slow or negligible sedimentation throughout much of the Tertiary and Quaternary.
9. There appears to be a causal relation between active fault zones and the volcanicity of the Antipodes Islands and the Cathedral Banks, neither of which are sited on rises.
10. Submerged banks and emergent islands, with the exception of the Antipodes Islands and deep Cathedral Banks, are characterised by Pleistocene planation.
11. Post-Pleistocene warping and depression of shelves, in relation to a world-wide mean datum of 130 m (Shepard, 1963), may indicate wholesale sinking of the Campbell Plateau, but is more probably causally related to tectonic activity of the rises on which the shelves are sited.

It is concluded that the Campbell Plateau developed as follows. Following formation of an erosion surface in the late Cretaceous, the plateau sank to about its present level. Deposition of moderately deep-water foraminiferal oozes probably began at this time and has continued. Mid-Tertiary tectonic warping resulted in the formation of broad rises about well defined axes. This resulted in the exposure of the mantle of indurated sediment against their basement cores. The rises are areas of slow deposition where formation of authigenic minerals is encouraged. Sedimentation is probably more rapid away from rises, a factor

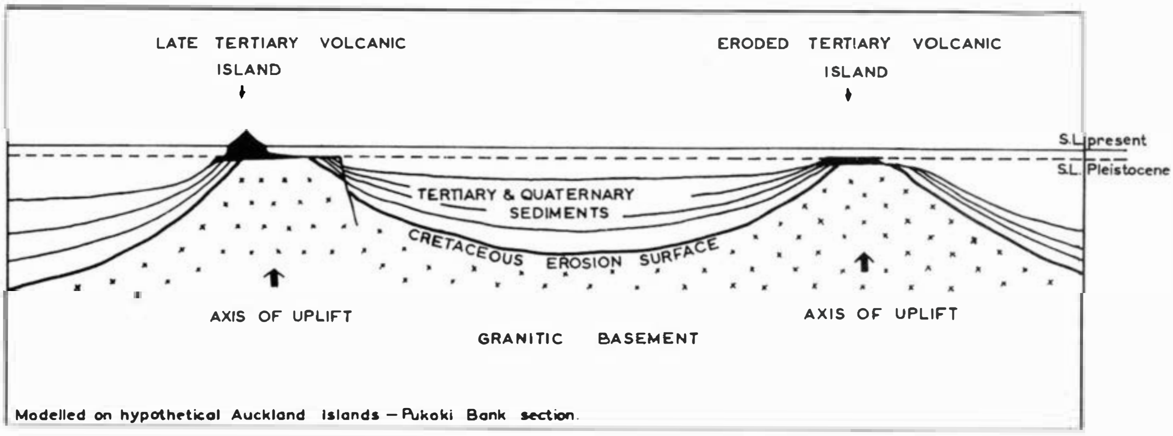


Fig. 16: Hypothetical crustal cross section of Campbell Plateau, indicating aspects of its probable geological history.

that may contribute to relative depression of the original plateau level between rises (Fig. 16). Uplift culminated in late Tertiary eruption at localised volcanic centres. That the rises have recently been active is indicated by warping of the Pleistocene erosion levels represented by the island shelves.

Fig. 16 is an idealised section across the plateau. The concept of fold development outlined here and in the above discussion has also been put forward for Cretaceo-Tertiary folding in New Zealand (Macpherson, 1946, fig. 2). Exposures of basement cores, for example in Hawke's Bay and north Canterbury, are mantled with Upper Cretaceous and Tertiary sediments in which numerous sedimentary disconformities are conspicuous, particularly on the flanks of domes or anticlines. The fold cores are regarded as " . . . arch-bends developed on a buried, subdued, mature-land cut on the basement Mesozoic and older rocks" (Macpherson, 1946, p. 15). However, although the plateau folds are similar in form to those developed in New Zealand at the same time, the trends of these two fold groups, with the exception of the Auckland Islands Ridge, are strongly divergent.

THE PLATEAU MARGINS

The steep 10–20° gradients of the Aucklands and Subantarctic Slopes are significantly higher than regional norms (compare Shepard, 1963; Menard, 1964). The straightness, steepness, and ruggedness of these features indicate probable structural control.

THE AUCKLANDS SLOPE

Aucklands Slope is the eastern margin of a seismically active rift depression, the Solander Trough. Normal faults probably cause mid-slope benching (compare Summerhayes, 1967) and the Aucklands Slope, rather than forming a single fault scarp, is probably the morphological expression of a narrow fracture zone (Figs. 12 and 17). Since it trends directly toward the New Zealand coast but is not manifest as a major fault on shore, it must be inferred that transcurrent movement plays little or no part in this fracture zone.

The junction between Aucklands Slope and the floors of the Solander Trough and Emerald Basin is commonly sharp. It is often associated with narrow marginal downwarps in the floors of these depressions, which probably indicates fairly recent downward movement along faults forming the slope (Summerhayes, 1967). Such downward movement along the edge of the Campbell Plateau is also suggested by the downwarping of the western edge of the Auckland Islands Shelf (Fig. 23) and, further north, the continental shelf.

SUBANTARCTIC SLOPE

Although neither as steep nor as straight as the Aucklands Slope, this feature is similarly benched and its morphology is controlled by normal faults. The junction between the slope and the flat Southwestern Pacific Basin is obscured by a gently sloping continental rise (see below). In comparison with the Aucklands Slope, which nearly everywhere lacks such a mantle and is known to be recently active, it must be concluded that the Subantarctic Slope has ceased to be an active fault feature. Like the Aucklands Slope, the Subantarctic Slope does not form a perfect fault-line scarp and is better regarded as a linear fracture zone, termed the Antipodes Fracture Zone by Cullen (1967). It is some 1,000 km long and cuts across the east-west-trending Chatham Rise and similar structures on the Campbell Plateau. Cullen (1967) considers that the fracture zone is a major transcurrent fault related in trend and probably in age to the Alpine Fault of New Zealand. This conclusion is discussed on p. 48.

Tensional characters are manifest along the slope, particularly in the blocky, faulted topography along the southern edge of the Campbell Plateau.

CONTINENTAL SLOPE

The north-west margin of Campbell Plateau is the New Zealand continental slope, a steep, straight feature for which a structural control is indicated (Fig. 12). This part of the continental slope forms the tail end of the continental slope off the South Island, New Zealand, which has been attributed to movement on the Waipounamu Fault (Cullen, 1967) extending north at least as far as Banks Peninsula (Fig. 24).

A 200-gamma magnetic anomaly occurs across the continental slope near the Snares. A 1-mile (1,600 m) drop in the level of magnetic basement beneath the Campbell Plateau is indicated by this anomaly. (Adams and Christoffel, 1962). It is reasonable to suppose that this displacement is caused by the Waipounamu Fault. Vertical displacement across the fault may be late Cretaceous or early Tertiary, the age of sinking of the Campbell Plateau, but is more likely to be mid Tertiary (*see later*). The fault has an evident normal component, but it has been suggested by Cullen that it is primarily a transcurrent fracture related to the Alpine Fault. Evidence for this conclusion is summarised on p. 48.

SUBMARINE CANYONS

Small submarine canyons cut the Inner Aucklands Slope just north of the Auckland Islands and are also developed on the inner slope south of the Snares. These canyons provide routes for the movement of turbidity currents between the shelf areas and the troughs and basins of the deep-sea floor (Fig. 6). It is possible that sediment moved in this fashion could be trapped by benches on the slope, but these are characteristically rugged features that do not have any depositional or erosional topography. Depositional features are found in the fault depressions separating the Endeavour Banks (Traverse 8d). Sediment supplied from the plateau covers the foot of the inner wall of each depression. At the foot of the opposite wall is a narrow channel, which probably leads out on to the Subantarctic Slope and thence to the Southwestern Pacific Basin. Sediment is supplied to these depressions probably by turbidity currents or through slumping. South of the Pukaki Saddle the Subantarctic Slope is cut by two broad depressions, which may be sea valleys through which sediment is supplied to the Southwestern Pacific Basin (Fig. 6).

South-east of the Auckland Islands the shelf edge is cut by small submarine canyons, which coalesce on the Campbell Plateau in the Carnley Sea Valley (Fig. 2). This is a poorly defined feature, directed south-west across the plateau, which probably connects with a narrow depression, believed to be a sea valley, running through the Cathedral Banks and heading towards the Cathedral Depression. This canyon and sea-valley system provides a route for the movement of sediment-laden turbidity currents from the Auckland Islands Shelf toward the plateau margin (Fig. 6). Sediment may ultimately be deposited on broad benches on the Aucklands Slope or in the Emerald Basin. The sea valley may be joined in the vicinity of the Cathedral Banks by canyons cutting the western shelf edge off Campbell Island.

CONTINENTAL RISE

Mantling the foot of the Subantarctic Slope is a continental rise (Appendix IV), the origin of which is assumed to be sedimentary. It is generally accepted that continental rises are formed by the coalescence of sediment fans formed at canyon outlets along the slope (Heezen, Hollister, and Ruddiman, 1966). Constant accumulation of sediment leads to development of a sediment wedge at the foot of the slope (Heezen *et al.*, 1966). Isostatic adjustment along the slope may result in normal faulting within the slope as a response to this sediment accumulation, and may explain the origin of step faults along the Subantarctic Slope. Another explanation is required for the step faults along the Aucklands Slope, where a continental rise is not developed.

An appreciably large negative magnetic anomaly beneath the continental rise south of the Campbell Plateau could be interpreted as reflecting a marginal trench filled with non-magnetic sediment (Christoffel and Ross, 1965). Krause (1966) also implies that a thick body of non-magnetic sediment overlies depressed basement in this region. The magnetic data seem to support the argument put forward for the depositional origin of the continental rise.

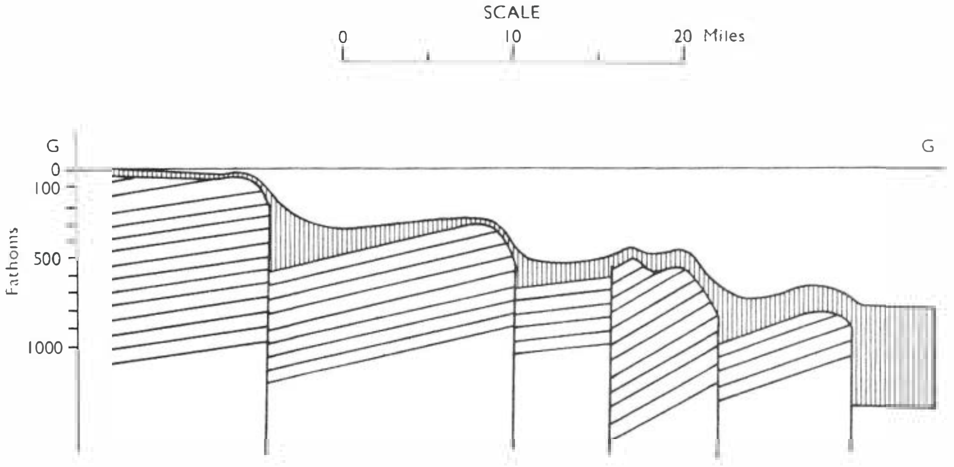


Fig. 17: Possible fault control of the morphology of the continental slope off the east coast of the North Island (from Pantin, 1963, fig. 12).

SOUTHWESTERN PACIFIC BASIN

ABYSSAL HILL PROVINCE

Distribution of depths approaches normal in the abyssal hill province of the Southwestern Pacific Basin (Fig. 18). Cumulative distribution of depth (Fig. 19) follows a straight line, indicating normal distribution between 4,700 and 5,700 m. In shallower depths a long, skewed tail persists to 3,000–3,100 m (Fig. 19).

If the skewed distribution between 4,400 and 3,000 m is temporarily ignored, it seems that we are dealing with a random distribution about a mean of effects produced by specific causes. Topography in the 5,700–4,700 m range is probably produced by a uniform controlling process, or series of processes, which could be volcanic (Menard, 1964) or tectonic (Krause and Menard, 1965). Superimposed on the regional abyssal hill topography are seamounts lying in the 4,400–3,000 m range, the origin of which is almost certainly volcanic.

ABYSSAL PLAINS

A flat abyssal plain lies between the abyssal hill province and the continental rise at the edge of the Southwestern Pacific Basin (Fig. 6, Traverses 9a-d). The suggested origins of abyssal plains have been carefully reviewed by Heezen and Laughton (1963), whose ideas are summarised below.

It is commonly found that abyssal plains (1) are situated between the continental rise and the abyssal hill province; (2) are characterised by the apparently anomalous occurrence of coarse sands containing shallow-water fossils; (3) have small-scale relief explicable only in terms of the action of bottom currents; and (4) occur only where there is no topographic bar to movement of currents from the direction of the continental rise. Thus Heezen and Laughton (1963) deduced that abyssal plains result from sediment deposition by turbidity currents spreading out on the deep-ocean floor adjacent to the mainland or other centres of erosion.

Routes for the movement of sediment from the Campbell Plateau to the Southwestern Pacific Basin have been described previously and are illustrated diagrammatically in Fig. 6. The lack of echo-sounding traverses along the plateau-margin slopes prevents adequate definition of submarine canyons and sea valleys on the plateau-margin slopes. Since a continental rise and abyssal plain are well developed at the foot of the Subantarctic Slope, sediment must have been supplied via several undiscovered routes.

DISPLACEMENT ZONE

The regional level of the Southwestern Pacific Basin is disturbed locally by a broad, elongate, northerly trending rise (Fig. 12). At one point along this feature is a seamount, probably volcanic in origin. The rise crest is parallel to and closely follows the line of displacement of a deep-ocean fracture zone (Fig. 12) identified by displacement of linear magnetic anomalies on the basin floor (Ross and Christoffel, in press). Displacement reaches nearly 200 miles near the Campbell Plateau but decreases to zero near the flanks of the Indian-Pacific Rise at 60° S (Fig. 24). Christoffel and Ross (1965) gave reasons for favouring, as a cause of the pattern of magnetic anomalies, either variation in magnetic susceptibility of a uniformly thick basement due to igneous activity or variation in the direction of remanent magnetisation.

MACQUARIE RIDGE

MORPHOLOGY AND STRUCTURE

Although the morphology of the northern part of the Macquarie Ridge has been briefly described (Brodie and Dawson, 1965) its structure has not previously been discussed. The ridge consists of a main ridge in depths of 1,750 m or less, a subsidiary ridge locally developed on the eastern, concave, side of the ridge, and a series of elongate narrow basins at the slope foot on the western or convex side of the ridge (Appendix IV; Fig. 12).

Steep flanks with sharply defined and roughly horizontal benches in intermediate depths are typical of the ridge, and imply a structural control, possibly by normal faulting. Sediment accumulation may have contributed to the level surface of some benches along the flanks.

A narrow, discontinuous, V-shaped median depression parallels the trend of the ridge, creating a "double crest" (Traverse 6b). Locally peaks rise from the ridge crest to as little as 100 m. These features may be cut by the median depression. Macquarie Island forms the emergent part of the eastern crest of the ridge, and is separated from a similarly shaped, flat-topped, submarine ridge to the west by the narrow median depression. This deep, narrow, linear feature is most probably a graben. Individual peaks along the ridge may be volcanic centres, which, since they are cut by the median depression, may be older than the rifting which gave rise to the depression.

The subsidiary ridge, elevated about 1,000 m above the Emerald Basin floor, is continuous with the main ridge north of 51° S. It is separated from the main ridge further south by a deep narrow trench reaching 5,700 m, some 1,200 m below the level of the Emerald Basin. Steep flanking slopes along the subsidiary ridge indicate a probable fault control of morphology (Fig. 12).

Trenches associated with the ridge form part of the overall structural complex and are also probably fault controlled.

In summary, the Macquarie Ridge consists of a series of elongate elevated blocks or horsts within a narrow, rigidly controlled structural zone (Fig. 12). The major horsts are separated by narrow fault depressions cutting across the ridge at right angles.

GEOPHYSICAL CHARACTER

Studies of the magnetic character of the Macquarie Ridge have been carried out by Hatherton (1967), and its seismicity has been studied by Cooke (1966). Distribution of earthquake epicentres along the ridge, which is seismically active, is shown in Fig. 4. Magnetic profiles discussed in the text are in Fig. 5.

A pronounced, narrow, positive magnetic anomaly closely follows the eastern flank of the ridge (Fig. 4). Should the anomaly be caused by the intrusion of basic or ultrabasic igneous material along an eastwardly dipping crustal fracture or plane of weakness it would conform with the interpretation of the eastern flank of the Macquarie Ridge as a fault structure controlled by eastwardly dipping normal faults. A genetic relation is implied between the anomaly and the faults forming the eastern flank of the ridge.

Seismic evidence indicates that the ridge is active (Cooke, 1966; New Zealand Seismological Records, 1962–64). In Fiordland there is a large, elongate, positive Bouguer gravity anomaly, reaching +160 miligals (Reilly, 1965), that is parallel to, and continuous in trend with, the Macquarie Ridge (Fig. 4). Where the Macquarie Ridge magnetic anomaly reaches the New Zealand coast it is aligned with the eastern flank of the Fiordland gravity anomaly (Hatherton, 1967). Fiordland is an elongate, upstanding, isostatically undercompensated massif, bounded eastward by the downfaulted Waiau Depression, which is parallel to the Fiordland massif (Fig. 9). Westward, off the Fiordland coast, isobaths descend steeply and abruptly to the Fiordland Trough in about 4,000 m. The trough is "a weakly developed but extensive feature, 30 miles across and 100 miles long", subparallel to the coast (Brodie, 1964).

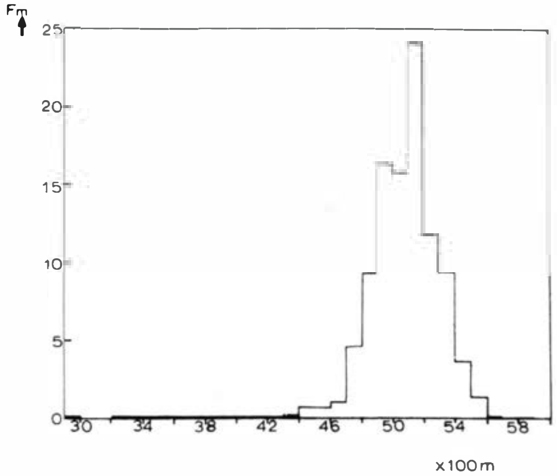


Fig 18: Depth distribution in the Southwestern Pacific Basin (compare Appendix IV for derivation of original values).

Hatherton (1967) observes that Fiordland forms a topographic and probably structural continuation of the Macquarie Ridge. Both features lie within an area of shallow seismicity terminated north at the Alpine Fault (Hatherton, 1967) and the ridge itself forms a prominently seismic feature further south (Cooke, 1966, and Fig. 4).

MACQUARIE PETROGRAPHIC PROVINCE

Normative analyses of the "calc-alkali" rocks (Mawson and Blake, 1943) of the Older and

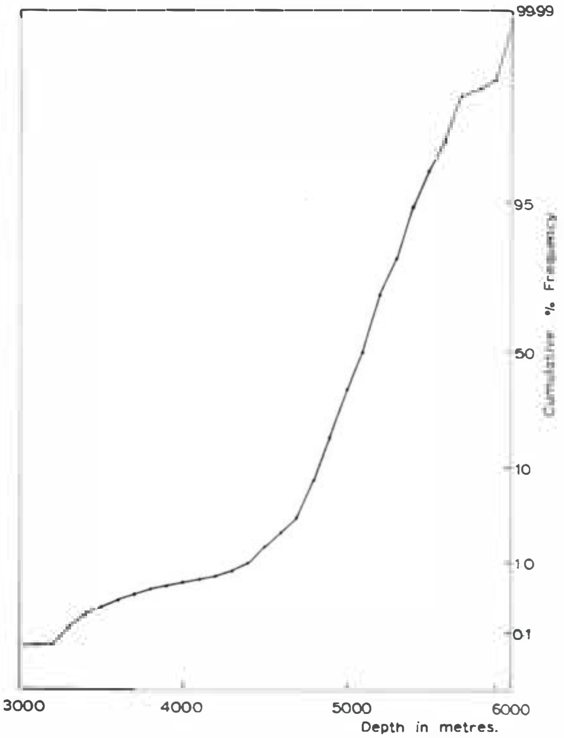


Fig. 19: Cumulative distribution of depth in the Southwestern Pacific Basin (derived from Fig. 18).

Younger Basic Series of Macquarie Island show hypersthene to be a common normative mineral. Nepheline occurs occasionally, and quartz in only one analysis. The common occurrence of normative hypersthene among both series may be taken, in conjunction with the general modal mineralogy, to indicate a broadly tholeiitic character (compare Yoder and Tilley, 1962). The plagioclase of the Macquarie Island Suite is more calcic than typical Hawaiian tholeiites, which have plagioclase An_{50}. The plagioclase of ridge samples (Appendix II) tends to have a greater range than those of the island, which are typically bytownitic. Otherwise there is very little difference in the mineralogical composition of gabbros, peridotites, dolerites, or basalts from ridge and island.

The occurrence of normative nepheline within both Older and Younger Basic Series, although in restricted amounts in a limited number of rocks, indicates that both series tend to produce alkali differentiates. Differentiation within the Younger Basic Series has proceeded to the extreme in producing very limited amounts of analcite basanites, tephrites, and soda-rich dolerites—all very alkaline rocks.

The gabbros and dolerites of E 236a, the gabbro of E 234, the gabbro and peridotite of D 9, and the gabbro and holocrystalline basalt of D 8 (Fig. 3) probably represent off-shore extension of the Older Basic Series of Macquarie Island. Basalts at D 8 and D 25, in which hornblende completely replaces previously existing pyroxenes, may also belong to the Older Basic Series, which has been subject to tectonic processes and some metamorphism. All the above samples show evidence for slight cataclastic deformation (Appendix II). Serpentinised peridotite at D 20, far to the north of Macquarie Island, indicates that the Older Basic Series may occur along the whole of the ridge.

Rocks of the Older Basic Group suffered uplift and subaerial erosion prior to the extrusion of Miocene basalts at Macquarie Island and are thus probably pre-Miocene.

The Younger Basic Series overlying the Older Basic Series unconformably at Macquarie Island (Mawson and Blake, 1943) are predominantly submarine extrusives or shallow crustal intrusives. The period of volcanicity of this Younger Basic Suite may be expanded from the Miocene eruptions of Macquarie Island to include the Pliocene eruption recorded from D 5, south of the Island (Appendix II). Glassy basalts, probably related to the Younger Basic Series, occur along the ridge at D 5, D 6, D 7, D 24, D 25, D 26, A 696, D 17, D 18, D 20, D 22, D 169, and D 159 (Fig. 3). Many of these are submarine extrusives.

A geochemical comparison may be made between rocks from the Macquarie Petrographic Province and other oceanic and circum-oceanic areas. Selected chemical analyses of rocks from Macquarie Ridge and Island are given in Table 4 together with analyses of tholeiitic basalt from Japan (Kuno, 1959), of Hawaiian tholeiites (Tilley, 1950), and oceanic tholeiites (Engel, Engel, and Havens, 1965). Alkali basalts from the Macquarie Province are also compared with an alkali basalt from Hawaii (Yoder and Tilley, 1962).

The Fe_2O_3/FeO ratios of Macquarie Island rocks are similar to other tholeiites but are rather high in the submarine samples D 17 and D 18. This may be due to slight post-emplacement alteration, although such alteration is not apparent in the modal mineralogy of these samples.

The Na/K ratios of rocks in the Macquarie Province show much greater conformity. They are strikingly different from oceanic tholeiites but similar to Hawaiian and Japanese tholeiites. Since the Na/K ratio is regarded as distinctive in oceanic tholeiites (Engel *et al.*, 1965) the Hawaiian, Japanese, and Macquarie tholeiites must be fundamentally similar, and different from the true ocean tholeiites of the mid-ocean rise system.

Macquarie Province tholeiitic rocks are less aluminous than typical oceanic tholeiites, are closely similar to the Japanese tholeiitic basalt, and slightly more aluminous than the Hawaiian tholeiites. Their MgO, Cao, K_2O, and Na_2O contents are similar to Japanese and Hawaiian rocks, but higher than in oceanic tholeiites.

Comparison of trace-element compositions (Table 13) between the oceanic tholeiite and the submarine basalts of the Macquarie Ridge indicates affinities of both alkali and tholeiitic basalts with the oceanic tholeiites. The Macquarie rocks, like oceanic tholeiites, have low Pb and Ba contents. They also have contents of Cu, Cr, Ni, Y, and Zr similar to oceanic tholeiites. The Sc content is more like that of oceanic alkali basalts (analyses derived from Engel *et al*., 1965, table 2, p. 721).

Comparing the alkali basalt D 7 of the Macquarie Province with an alkali basalt from Hualalai, Hawaii (analyses derived from Yoder and Tilley, 1962), there is little obvious difference in chemistry (Table 4). The Na/K ratio of the Macquarie sample is quite high and the Na and K contents are closely similar to the tholeiitic Macquarie basalt D 18. It can only be inferred that the alkali basalt is but slightly alkaline and also, possibly, that the tholeiitic basalt sample D 18 has alkaline affinities. The porphyritic bytownite basalt from Macquarie Island (Table 4) has a more typically alkaline Na/K ratio.

Since the Macquarie, Hawaii, and Japan tholeiites and tholeiitic basalts have compositions intermediate between oceanic tholeiites and alkali basalts and are associated with alkali differentiates and ultrabasics, it is reasonable to infer that they form part of a differentiation sequence culminating in formation of alkali rocks (compare Engel *et al.*, 1965). The composition of parent magma from which these rocks were derived may be similar to that of oceanic tholeiites (Engel *et al.*, 1965).

Macquarie Ridge, an Island Arc

The narrow, slightly arcuate Macquarie Ridge (Fig. 4) forms a distinct morphological, seismic, structural, and petrographic province extending from about 60° S to the New Zealand continental shelf near Puysegur Point.

In comparison with other pronounced oceanic ridge systems the Macquarie Ridge differs from (1) volcanic chains, such as the Hawaiian Archipelago, in having more basic volcanism and a horst-graben topography with no recognisable volcanic vents; and (2) mid-ocean ridges, particularly in being extremely narrow where these features are 1,000–4,000 km wide.

Gutenberg and Richter (1954) show that a well defined belt of seismicity extends south of New Zealand along the line of the Macquarie Ridge to 60° S where it meets the mid-ocean ridge system. Heezen and Ewing (1956) have argued that Macquarie Ridge is a branch of this mid-ocean ridge system. However, the structure of the ridge, as determined from morphology, is more closely related to typical circum-Pacific structures, the island arcs (Fig. 20). The presence here of an oceanic trench on the convex side of a well developed, long, narrow, arcuate, seismically active, and volcanic belt occurring near the boundary between continental and oceanic crust, is typical of island arcs as defined by Gutenberg and Richter (1954). The association of the arc with a transcurrent fault, the Alpine Fault, occurring at one end of the arc and on its concave side, is supposedly typical of island arcs (compare Menard, 1964). Control of topography by normal faulting, as here postulated for the Macquarie Ridge, is another character of island-arc - trench associations (Brodie and Hatherton, 1958; Menard, 1964; Ludwig *et al.*, 1966).

Incomplete development of a trench in association with the island arc is also typical of the New Hebrides and Solomon Island systems, where the major trench development occurs at each end of the arc and is discontinuous in the centre (Fisher and Hess, 1963) as in the Macquarie Ridge complex.

These authors have also explained the development of fractures at right angles to the ridge as reflecting tension cracks parallel to the supposed direction of compression associated with the arc (Fig. 21, from Fisher and Hess, 1963, fig. 10, p. 431). Development of such features would satisfactorily explain the fractures that split the Macquarie Ridge into a series of elongate horsts.

Hatherton (1967) postulates that Macquarie Ridge and the upstanding Fiordland massif may be related parts of a single structural complex. There is a pronounced change in level (Fig. 22) and geology between the Macquarie Ridge (about 1,300 m deep), the Puysegur Bank (about 180 m deep), and Fiordland (some 1,500 m above sea level). Between the ridge and the continental shelf

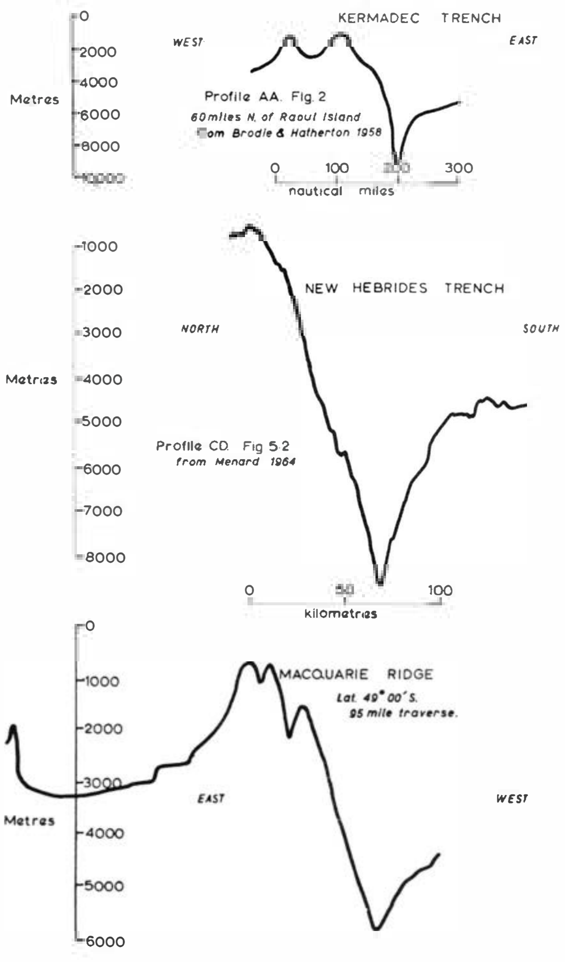


Fig. 20: Morphology of typical island-arc - trench associations, including Kermadec Ridge and New Hebrides Trench for comparison with Macquarie Ridge.

there is probably a major structural boundary, possibly reflecting the difference between oceanic crust to the south and continental crust to the north. However, the forces producing the Macquarie Island Arc obviously transgress this boundary, producing morphological and seismic continuation of off-shore characters in the continental crust of southern New Zealand.

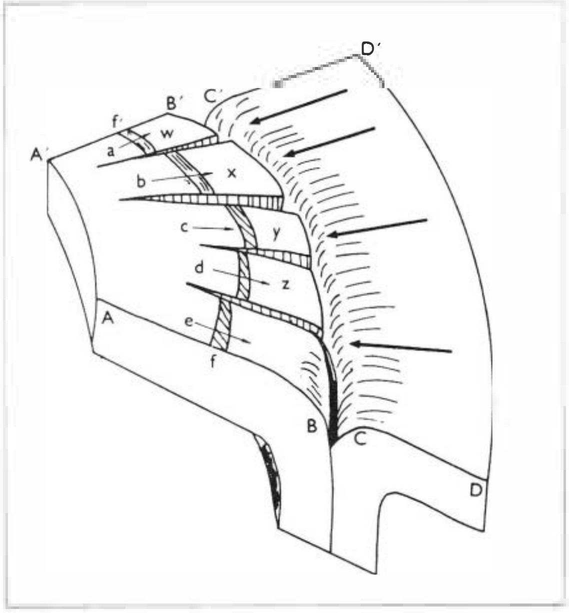


Fig. 21: Origin of transverse fractures in island-arc systems; a postulated mechanical situation, derived from Fisher and Hess, 1963. Movement A - B causes extension B - B′ allowing fissure development x, y, z. Movement D - C involves compression C - C′, precluding fissure formation. Relative movements of a, b, c, d may differ causing formation of strike-slip faults between adjacent blocks. This general situation may apply along Macquarie Ridge.

Carey (1958) defined that part of the Macquarie Ridge south of Macquarie Island as the Macquarie Nematath, a submarine ridge thought to mark the path of separation of crustal units originally much closer together (Carey, 1958, figs. 46a and b). At this time the extension of the ridge to the mainland shelf off Puysegur Point was not known. Since the present data seem best interpreted by regarding the ridge as part of the circum-Pacific belt, Carey's hypothesis needs re-examination.

It is here concluded that the Macquarie Ridge is an island-arc complex forming an integral part of the circum-Pacific belt as far, possibly, as the Balleny Islands.

RIDGE DEVELOPMENT

The ridge may have begun to develop during the Cretaceous (Mawson and Blake, 1943). After elevation of the ridge marine erosion exposed the pre-Miocene Older Basic Series at Macquarie Island. The submarine extrusive volcanism of the Younger Basic Series began in Miocene times when some subsidence had occurred at Macquarie Island. The ridge was also volcanically active in the Pliocene. Faulting of the median rift may be post-Miocene since the Miocene rocks of Macquarie Island are elevated as a fault-bounded horst on the eastern side of the rift. However, the Younger Basic Series basalts may have been extruded from earlier fissures along the line of this rift.

Inception of late Miocene faulting along the Hauroko Fault marginal to Fiordland (Wood, in press) indicates that evolution of the ridge may have commenced first near its southern end and reached New Zealand only during the late Miocene. Growth of the Solander Trough, however,

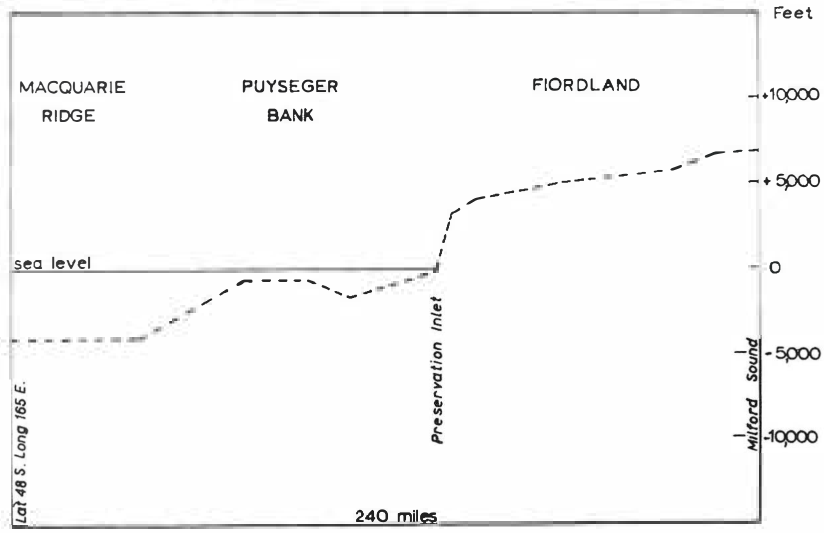


Fig. 22: Topographic section from Milford Sound through Fiordland and Puyseger Bank to the northern end of Macquarie Ridge.

commenced at least in the earliest Tertiary, judged by the sedimentary history of its northern projection, the Waiau Syncline. The trough is considered closely related to the Macquarie Ridge, and its age may give the closest estimate of the age of formation of the Macquarie Ridge (*see below*).

SOLANDER TROUGH

MORPHOLOGY, STRUCTURE, AND DEVELOPMENT

The Solander Trough is a narrow linear trough with a smooth floor sloping gently south. It extends from the vicinity of the Solander Islands off the South Island of New Zealand to the latitude of the Auckland Islands, where it gives way to the Emerald Basin (Fig. 12).

The eastern flanks of the trough are formed by the Aucklands Slope Fracture Zone, and the western by faults of the Macquarie Ridge complex. Thus the Solander Trough, a long strip of sea floor depressed between crustal blocks bounded by normal faults, is a rift depression (Fig. 23). At its head it bifurcates, one branch passing east, the other west of the Solander Islands (Fig. 9), and neither branch cuts across the continental shelf of New Zealand. Brodie (1958) pointed out the alignment of the Waiau Depression with the eastern branch of the Solander Trough. He considered this alignment to be tectonic in origin.

The Waiau Depression, thought by Park (1921) to be a rift valley, is a narrow tectonic depression extending from at least the northern tip of Lake Te Anau southward to the coast. It is bounded on the west by probably normal faults such as the Hauroko fault separating Fiordland rocks of Paleozoic and Mesozoic age from the Tertiary rocks of the Waiau Depression (Wood, in press). Members of the Tertiary sequence overlap unconformably eastward and may be interrupted by

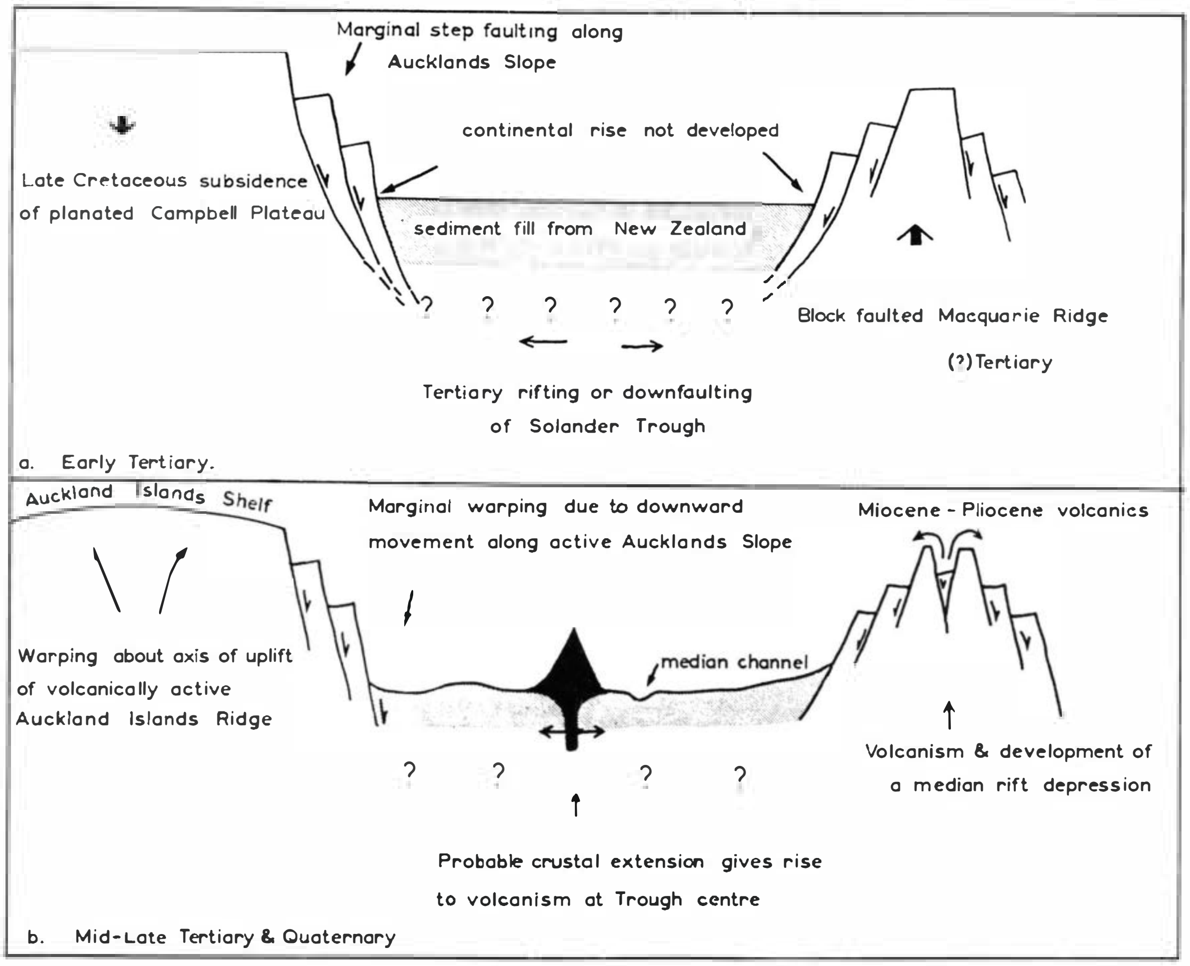


Fig. 23: Hypothetical Tertiary development of Solander Trough.

normal faulting (Wood, in press). North of the Tuatapere area studied by Wood the eastern boundary of the Waiau Depression is formed by a major fault skirting the western edge of the Takitimu mountains. Fault movements in the southern part of the area occurred mainly during the Waitotaran (Pliocene), but movement of the Hauroko Fault may have started in the Tongapurutuan (Upper Miocene). All fault movements along the margins of the Waiau Depression are considered to be Tertiary or Quaternary and related to the Kaikoura Orogeny (Wood, in press).

Linear downwarping in this area was first evident during the Oligocene when the Landon Series were deposited in a north-north-east - trending "geosyncline" on the site of the Waiau Depression (Wood, in press; Fleming, 1962, fig. 7).

The Waiau Depression is characterised by a narrow negative Bouguer gravity anomaly reaching −25 to −50 miligals and extending from the north of Lake Te Anau south to the sea (Fig. 4, from Reilly, 1965). The forms of gravity and isostatic anomalies (Fig. 4) indicate that the southern part of the depression has a basin form. Whether these anomalies continue southward is still conjectural. The negative gravity and isostatic anomalies of the Waiau Depression contrast strongly with the pronounced positive anomalies characterising the Fiordland region to the west (Fig. 4).

Along the axis of the Solander Trough are three seamounts (Fig. 12) and one volcanic island (Fig. 9). The largest seamount, some 50 miles west of the Auckland Islands, reaches 950 m and has a typical volcanic morphology—conical with steep flanking slopes. Smaller seamounts lie 20 miles north and south of it. The northernmost seamount has twin peaks and is characterised by a twin-peaked positive magnetic anomaly that exactly follows the topography (Fig. 5). The magnetic evidence favours a basic volcanic origin for this feature since it consists of material of magnetic susceptibility radically different from its surroundings. Solander Islands are composed chiefly of hornblende andesites and are probably Upper Pleistocene (Harrington and Wood, 1958; Table 2).

Magnetic profiles across the trough indicate a positive anomaly near the axis and along much of its length (Hatherton, 1967; Fig. 5).

Shallow earthquakes recorded from the centre and margins of the Solander Trough (Fig. 4, based on N.Z. Seismological Records 1962–64) indicate that the Solander Trough is an active crustal feature. Northward, the Waiau Depression lies in the centre of an active "envelope of seismicity" covering much of the southernmost part of New Zealand and extending over the Solander Trough and Macquarie Ridge (Fig. 4).

The Auckland Islands Ridge (Fig. 12), along the western margin of the Campbell Plateau, is thought to have been actively uplifted since the late Cretaceous erosion of the plateau, probably during the Miocene (p. 34). The ridge may be genetically related to the Solander Trough since both are Tertiary features with the same trend. The Tertiary volcanicity of the Auckland Islands and the Pleistocene warping of the shelf around the Auckland Islands indicate close association between the activity of the western margin of the Campbell Plateau and the sinking of the Solander Trough (Fig. 23).

Downwarping of the western edge of the Aucklands Islands Shelf and continental shelf near the Snares toward the Solander Trough may indicate recent downward movement on faults forming the Aucklands Slope, and may reflect either downward movement of the trough floor or relative downward movement of the western margin of the Campbell Plateau (Fig. 23; Summerhayes, 1967).

In summary, the Waiau Depression has been a depressed zone since the Oligocene. Marginal faults have been developed probably since the late Miocene, and definitely since the Pliocene. It is a presently active seismic feature. Topographically and structurally it is continuous with the Solander Trough, a seismically active, submarine fault depression. Incidence of volcanism along the axis of the trough may be taken as indicative of tension in the crust, and the presence of a linear magnetic anomaly along the axis of the trough also implies tensional rifting (Hatherton, 1967). I infer that the Solander Trough and Waiau Depression form related parts of a single geological, geophysical, and topographical complex. Since the Waiau Depression was an active geosyncline or synformal depression during the Oligocene the inception of the Solander Trough is assumed to be of this age.

NATURE OF THE TROUGH FLOOR

The trough floor slopes smoothly and gently southward. It is partly margined by low rises that obscure the junction between the flat trough floor and the steep flanking slopes. These rises are similar to the continental rise and are thought to be caused by deposition from turbidity currents or by slumping of sediment derived from the Campbell Plateau and adjacent shelf regions. Locally a continental rise is not developed, owing to either disruption of a previously existing rise by movement along faults margining the trough or a lack of deposition because of lack of a suitable sediment source in the immediate vicinity (compare Summerhayes, 1967).

Since a depositional continental rise forms much of the trough floor, smoothness of this feature is unlikely to reflect preservation of some former erosion surface perhaps formed at or near sea level. A well developed, narrow, median channel running down the length of the trough is similar in form to channels found in the Bounty Trough (Fig. 14). It could have been cut by movement of low-density turbidity currents (compare Menard, 1964) or by rivers at or near sea level. Preservation of a river-cut feature is unlikely, however, owing to sedimentation. The association between turbidity currents and channels cutting continental rises is a recognised phenomenon. There is every reason to suppose a genetic relation here between the channel and the smooth trough floor, and it is considered that turbidity currents provide the most suitable explanation for development of a channel in this environment.

Submarine canyons, through which turbidity currents may move, carrying sediments from the continental shelf down into the Solander Trough, are present around the edges of that feature. Off Stewart Island (Fig. 9) is the Mason Canyon (Cullen, 1965a), and other canyons occur on the shelf edge near the Snares and on the Auckland Islands Shelf edge (p. 39). Other depressions in the eastern flank of the Macquarie Ridge may be submarine canyons, but these are probably not "active" features since the ridge is probably too deep to be a present-day sediment source. Parts of the ridge were subject to marine erosion during the Pleistocene and canyons may have developed at this time. The general pattern of sediment movement in the Solander Trough is shown in Fig. 6.

EMERALD BASIN

THE BASIN FLOOR

The floor of the Emerald Basin is an abyssal plain (see p. 40) sloping gently down to south and west (Fig. 6). Abyssal hills rising from this plain are relict elevations of an initially rough sea floor. They are similar in depth and form to the abyssal hills of the Southwestern Pacific Basin and may be remnants of a similar ocean-basin floor now buried by sediments. If the original basin floor was as deep as the Southwestern Pacific Basin (5,100 m), about 500 m of sediment may have accumulated on the floor of the Emerald Basin.

A continental rise is developed locally at the foot of the Aucklands Slope. This is not a continuous feature and may be completely absent south of 53° S. Lack of a continental rise is explained if the supply of sediment to the area were mainly from the north along the Solander Trough. The Auckland Islands Shelf must have acted as a sediment source during the Pleistocene. Also, canyons on the Auckland and Campbell Islands shelf edges coalesce in the Carnley Sea Valley, which may supply sediment to the Aucklands Slope and thence to the Emerald Basin (Fig. 6). However, from the absence of a continental rise it is deduced that the Campbell Plateau has not been a major source of sediment for the Emerald Basin.

The median channel of the Solander Trough is thought to provide a route for the southward movement of sediment, particularly from the New Zealand continental shelf (Fig. 6), which may be the main source of sediment for the Emerald Basin.

STRUCTURE AND DEVELOPMENT

Emerald Basin may be: (1) a downfaulted block of continental crust; (2) a new ocean basin formed by the rifting apart of adjacent crustal blocks; (3) a previously existing ocean basin; or (4) a region of thinned continental crust formed on the drifting apart of adjacent crustal blocks.

The basin is very nearly as deep as other deep-ocean basins and is thus probably floored by oceanic crust, covered with a veneer of sediment (*see above*).

Relation of the oceanic Emerald Basin to the continental Campbell Plateau is obscure. Since the Emerald Basin has not been supplied with sediments from the Campbell Plateau it presumably originated during the late Cretaceous after the plateau ceased to be an effective sediment source.

It is a fault-bounded feature, continuous topographically, and presumably structurally, with the Solander Trough. As it is of similar age to that feature it may have been caused by the same system of tectonic forces. The boundary between Emerald Basin and Solander Trough may be a boundary between oceanic and thinned continental crust. Solander Trough is a tectonic depression partly filled with sediments, but the nature of the underlying crust is not known. The large group of submarine volcanoes at the southern end of the trough may indicate the presence of a major structural boundary. Here, the continental crust may be thinned or rifted apart causing volcanism. There is some geophysical evidence in Solander Trough for the existence of tensional cracks (Hatherton, 1967). Widening of these might ultimately result in the formation of a new ocean basin, as is occurring in the Red Sea and the Gulf of California (Girdler, 1962) where tensional processes have allowed large quantities of basic igneous material to well up through gaps in the overlying continental crust, achieve hydrostatic equilibrium at the level of deep-ocean basins, and form a new deep-ocean floor. Similarly the Emerald Basin may be a newly formed ocean basin of dilatational origin, the southern part of a major rift system.

Within the Waiau - Solander - Emerald Depression are negative Bouguer anomalies, shallow earthquakes, and a large positive marginal magnetic anomaly (Fig. 4), all characteristics of rift systems (compare Carey, 1958; Girdler, 1964), supporting their postulated dilatational origin. It is concluded that this major depressed zone is a single major tectonic complex, a rift system.

TASMAN BASIN

Available data suggest that the Tasman Basin is a typical deep-ocean basin (Standard, 1961).

STRUCTURAL SYNTHESIS

STRUCTURES IN THE NEW ZEALAND SUBANTARCTIC REGION

Major highs and deeps on the sea floor around New Zealand exhibit marked consistency of trend over large areas (Fig. 24) and are classified into two major structural provinces (after Brodie, 1952, 1958). Campbell Plateau is thought formerly to have been continuous with Lord Howe Rise, and Norfolk Ridge with Chatham Rise, these structures being displaced by dextral transcurrent movement across the Alpine Fault. These structures form a distinct North-western - Chatham structural province thought to be Mesozoic and older. Their primary east–west to north-west trend is reflected in the pre-Tertiary structure of New Zealand (compare Lillie, 1951; Wellman, 1956).

During the Tertiary, structures with predominantly north-east to north-north-east trend developed in New Zealand (Lillie, 1951; Wellman, 1956) and are directly relatable to off-shore Tertiary structures with similar trend, the Kermadec and Macquarie Ridges and the Havre and Solander Troughs. These, with most of New Zealand, lie within the narrow, well defined Kermadec structural province (Brodie, 1952, 1958).

CAMPBELL PLATEAU

The Campbell Plateau forms part of the North-western - Chatham structural province.

Dominating the central region of the plateau are the Pukaki and Campbell Island Rises (Fig. 12), Tertiary structures with east–west trend. In view of the broad contemporaneity of folding and the similarity in fold mechanism between these ridges and New Zealand Tertiary structures, it is concluded that a simple widespread geologic process may control the Tertiary history of both Kermadec and North-western - Chatham structural provinces despite their pronounced divergence of fold axes.

The plateau is margined southward by the Antipodes Fracture Zone and north-westward by the Waipounamu Fault, both with north-east, Kermadec trend. The Waipounamu Fault (Cullen, 1967) separates the plateau from the New Zealand continental shelf (Figs. 12 and 24). This fault, which extends north to near Banks Peninsula, forms a major structural boundary separating the Kermadec structural province from the off-shore Chatham province. Cullen observes that it sinistrally offsets the Southland Syncline, probably originally continuous off shore as the Bounty Trough. It has been shown here that Waipounamu Fault is truncated by the Aucklands Slope Fracture Zone and transcurrent movement along the fault must therefore be pre-early Tertiary, the age of the Fracture Zone. A case has also been put forward for relative elevation of New Zealand along the Waipounamu Fault during the mid-Tertiary (p. 38). Since the fault is not seismically active, movement has probably ceased.

Antipodes Fracture Zone (Cullen, 1967) truncates both the Campbell Plateau and the Chatham Rise (Fig. 24). It must therefore post-date these features of the North-western - Chatham province. Cullen (1967) suggests that Antipodes Fracture Zone is a mid-Mesozoic transcurrent fault related, in view of parallelism, to the Alpine Fault. More recent data suggest that the Antipodes Fracture may be a major rift margin fracture; this will be considered later.

Aucklands Slope, the western boundary of the Campbell Plateau, is a fracture zone marginal to a dilatational rift, the Waiau-Solander-Emerald Depression. The rift is Tertiary and has a north to north-north-east trend, and so forms an integral part of the Kermadec Province. Alignment of the

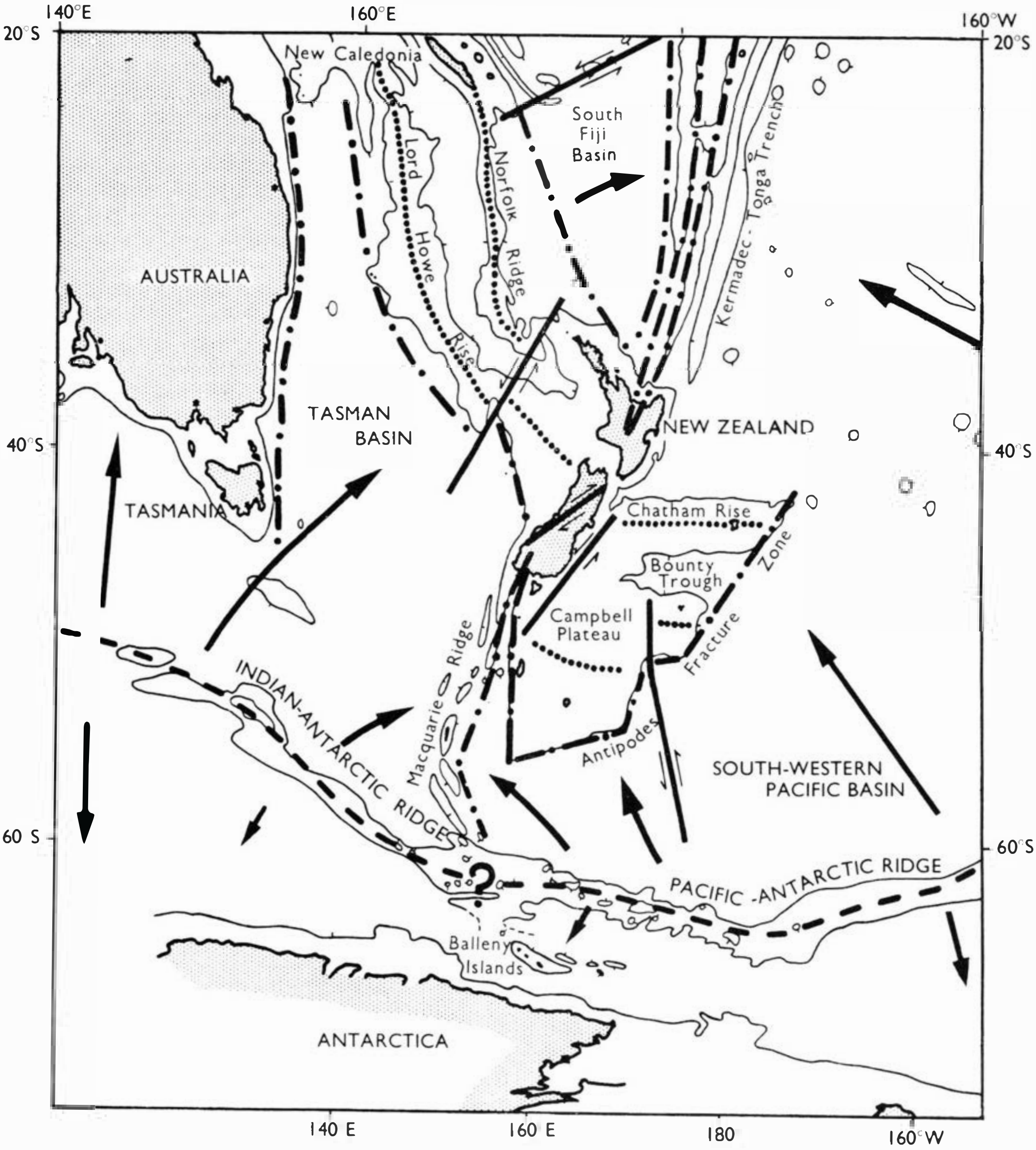


Fig. 24: Broadly defined morphology and structure of the south-western Pacific. Major fault-bounded rift features are shown by dashes + dots, and major ridge structures in North-western and Chatham Provinces by dotted lines; the approximate crest of the mid-ocean rise south of New Zealand is shown by a dashed line; strike-slip faults are shown by heavy lines; postulated directions of movement of crustal material away from the crest of the mid-ocean rise are indicated by arrows. This map is based partly on van Bemmelen, 1965; the base map is from Shuran, 1963.

twin volcanic domes of the Auckland Islands, the Auckland Islands Ridge, and the Aucklands Slope implies a genetic relation between rifting and the formation of volcanically active rises on the Campbell Plateau. Further south, other elevations occur along the plateau edge (Fig. 12) and may similarly be related in time to development of the Aucklands Slope Fracture Zone. Marked divergence in trend between the Auckland Islands Ridge and other rises on the plateau may be explained by the proximity of the Auckland Islands region to an evolving rift structure.

The north-north-east-trending Cathedral Depression - Cathedral Banks complex parallels the Waipounamu and Antipodes Fractures and may initially have been of similar age and fault character. There is no evidence for Tertiary transcurrent movement along this line but vertical movements have occurred to the south, allowing formation of Cathedral Depression flanking the Auckland Islands Ridge. Further evidence for tension along this fault line is given in the probably Tertiary volcanicity of the Cathedral Banks (p. 36). A similar association between late geological volcanicity and a possibly Mesozoic fracture is evident at the Antipodes Islands, probably Quaternary volcanoes along the edge of the Antipodes Fracture Zone.

The north-south-trending Pukaki Saddle is a tectonic depression (Fig. 13) across which the Campbell Plateau is downfaulted some 400 m to the east. The saddle lies along the line of a major sinistral displacement zone recognised on the floor of the Southwestern Pacific Basin (Fig. 12). Displacements across this zone reach 200 miles as the plateau is approached, and the Bounty Platform may have been moved northward along a fault within the Pukaki Saddle forming part of this displacement zone (Fig. 24). The Bounty Platform appears to have been shifted northward by about 60 miles, this displacement resulting in the pronounced kink in the Subantarctic Slope near the Antipodes Islands.

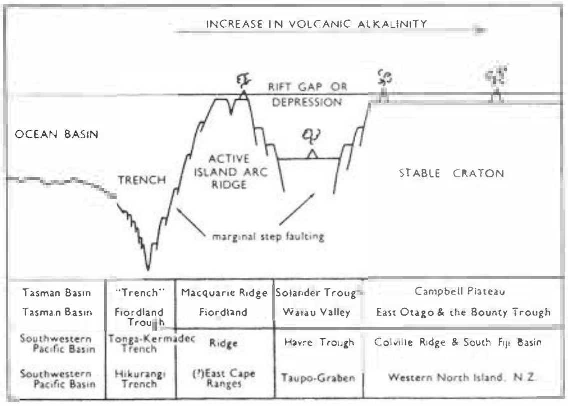
MACQUARIE RIDGE, SOLANDER TROUGH, AND EMERALD BASIN

Elements of the north to north-north-east-trending Kermadec structural province south of New Zealand include Macquarie Ridge, Solander Trough, and Emerald Basin, all probably of Tertiary age.

Macquarie Ridge, an island arc, is obviously at a different stage of development from the Kermadec - Tonga island-arc system which forms the continuation of the Kermadec structural province north of New Zealand (Fig. 24). The ridge (Fig. 12) is a horst-graben complex characterised by parallel and transverse tensional faulting. Development of a trench on the convex side of the Macquarie Ridge is incipient and localised, resulting in discontinuous linear depressions in about 6,000 m (Fig. 12). One of these is the Fiordland Trough, a moderately deep, elongate depression off the Fiordland coast (Brodie, 1964).

The fault-bounded Solander Trough and Emerald Basin apparently form a complex dilatational rift structure continuous on the New Zealand land mass in the Waiau Depression (Figs. 9 and 24).

In structural setting the Waiau - Solander - Emerald Depression occupies a position directly comparable with the Havre Trough north of New Zealand and its probable southward structural analogue on the mainland, the Taupo Graben. Havre Trough (Figs. 24 and 25) is some 800 miles long, about 70 miles wide, and reaches a depth of 3,657 m (2,000 fm). It is considered to be a broad seaward prolongation of the tectonically controlled thermal area of the North Island of New Zealand (Brodie and Hatherton, 1958). Andesites in the Taupo Graben and at White Island are in an analogous structural position to the hornblende andesites of Solander Island, a relationship which is considered hardly fortuitous. The ages of these volcanoes are also closely similar, Quaternary andesites in the North Island (Grindley, Harrington, and Wood, 1959) being the same age as those of Solander Island. The Havre Trough, the Solander Trough, and their terrestrial continuations lie on the concave side of an island arc between the arc and regions of relative crustal stability—the Campbell Plateau south of New Zealand, and western North Island, New Zealand, and the South Fiji Basin north of New Zealand (compare Fig. 25).



Tasman Basin	"Trench"	Macquarie Ridge	Solander Troug	Campbell Plateau
Tasman Basin	Fiordland Trough	Fiordland	Waiau Valley	East Otago & the Bounty Trough
Southwestern Pacific Basin	Tonga-Kermadec Trench	Ridge	Havre Trough	Colville Ridge & South Fiji Basin
Southwestern Pacific Basin	Hikurangi Trench	(?)East Cape Ranges	Taupo-Graben	Western North Island, N.Z.

Fig. 25: Generalised crustal section through island arcs in the New Zealand region.

A generalised crustal cross section (Fig. 25) summarises the characteristics of the island arcs of the Kermadec structural province north and south of New Zealand. The main structural elements are:

(i) a deep-ocean basin on the convex side of the major structures;
(ii) a deep, narrow "trench" at the edge of the ocean basin;
(iii) a narrow, steep-sided, fault-bounded, prominent ridge forming the main "island arc";
(iv) a narrow, elongate, rift-like depression, on the concave side of the arc, characterised by andesite volcanism;
(v) a stable continental or oceanic crustal segment or foreland, on the side of the depression opposite the arc, characterised by alkali volcanism (in the North Island and South Island of New Zealand and on the Campbell Plateau).

PETROGRAPHIC PROVINCES

There is a close relation between structure and volcanism in the New Zealand Subantarctic. The alkali volcanism of the Campbell Plateau occurs at isolated centres on a seismically stable, thinned section of continental crust. Volcanism on the Macquarie Ridge is dominantly tholeiitic, although small amounts of alkaline differentiates are produced. Volcanic samples from the ridge are mainly from peaks which may be volcanic cones. The median rift along parts of the ridge could also be the site of volcanic fissure eruptions. The structural setting of the Macquarie Tholeiite Province is that of a circum-Pacific island arc.

At least three volcanic centres are developed within the Solander Trough. The one sampled centre is Solander Island where hornblende andesites occur. It is postulated that the Solander Trough - Emerald Basin depression forms a third petrographic province, characterised in continental regions by hornblende andesite volcanism as at the Solander Islands. A trend towards increasing alkalinity away from the trench of an island arc towards stable regions on the concave side of the arc is recognised here as in Japan (Kuno, 1959) and the North Island of New Zealand (Cole, 1967).

Evidently there is a close interrelation between the mid-Tertiary and Quaternary volcanicity and structure of Campbell Plateau, Solander Trough, and Macquarie Ridge, although these are separate structural units and apparently distinct petrographic provinces. The three provinces probably represent the different effects of the same tectonic events on areas of differing crustal and subcrustal structure.

PRE-TERTIARY GEOLOGICAL HISTORY OF THE SOUTH-WESTERN PACIFIC

The formerly continuous structure comprising Campbell Plateau and Lord Howe Rise (Brodie, 1952) has an off-shore history which, as continental crust, covers at least 190 million years, the age of the Bounty Island granite. A long geosynclinal history is indicated for the pre-Permian rocks of New Zealand, and by implication, for the rocks of the Campbell Plateau - Lord Howe Rise. This extensive orogenic belt contains rocks thought to be pre-Cambrian (Grindley, Harrington, and Wood, 1959) for some of which a probable late Precambrian age has been determined using radioactive-dating techniques (Aronson, 1965).

The history of the Southwestern Pacific, north of New Zealand, is characterised by the development of a series of island arcs which are youngest north-eastwards away from Australia (Hess and Maxwell, 1953). The youngest and furthest from Australia is the present Tonga - Kermadec system. Carey (1958) and van Bemmelen (1965) consider these orogenic belts or island arcs to form a broad dilatational zone in which orogenic ridges are migrating away from Australia. Ridges are separated from each other by broad rift gaps such as the Tasman Basin and the Fiji Basin. Movement of the opening ends of these rifts has occurred along large strike-slip faults such as the Alpine Fault and the apparent strike-slip fault (Fig. 24) joining the New Hebrides arc to the Tonga - Kermadec arc (Hess and Maxwell, 1953).

Standard (1961) and Officer (1955) consider that the orogenic belts of this region have been built up from primary oceanic crust by successive orogenies and that individual belts are still in the areas where they formed. These hypotheses do not take transcurrent movement along faults such as the Alpine Fault into consideration. Also, since the geosynclinal sediments of the New Zealand Precambrian and Paleozoic were evidently derived from some nearby land mass these elements of continental crust must have been closer to Australia in the past, this being the nearest ancient continental massif.

Close Paleozoic faunal affinities between New Zealand and Australia suggested the existence of "Tasmantis", a continental massif occupying the Tasman Basin, along the shore line of which faunal migration could occur (Gill, 1952–53; Glaessner, 1952). A massive conversion of continental "Tasmantis" into the oceanic Tasman Basin floor is implied in this hypothesis. Basification of the crust after the manner proposed by Beloussov and Ruditch (1961) might satisfy this requirement but is not thought necessary since the Lord Howe Rise - Campbell Plateau ridge system linking Australia to New Zealand fulfils most of the requirements of the postulated "Tasmantis".

Van Bemmelen (1965) and Wright (1966) postulate that the north-eastward drift of New Zealand and its attendant ridge system, together with such continental fragments as Fiji, was accomplished as a response to the development of the Indian Ocean mid-ocean ridge system, and possibly achieved by convection currents (compare Fig. 24).

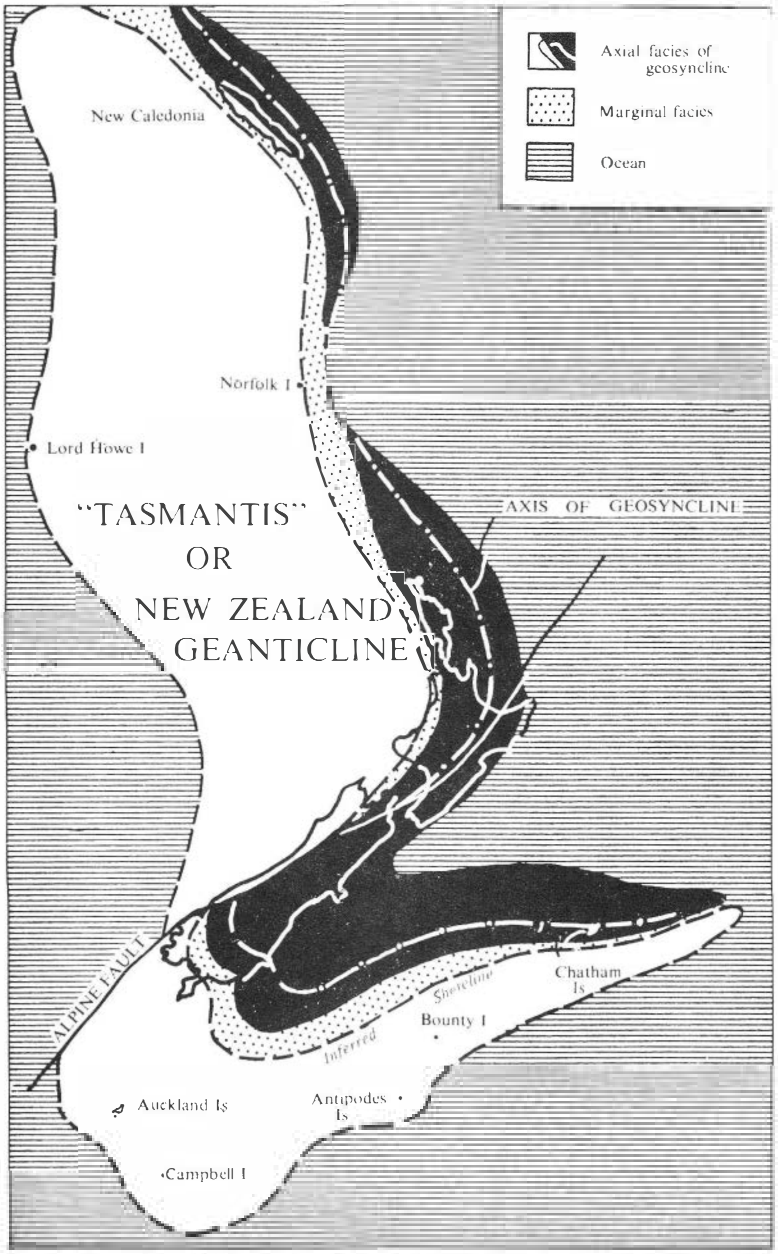


Fig. 26: The New Zealand Geosyncline during the Triassic (from Fleming, 1962, fig. 3) illustrating the probable position of a western foreland comprising Lord Howe Rise, Campbell Plateau.

The belt of crystalline and sedimentary rocks forming the Lord Howe Rise - Campbell Plateau is thought to have acted as a foreland for the New Zealand Geosyncline (Fig. 26, based on Fleming, 1962). Norfolk Ridge and the Chatham Rise may be composed of the uplifted and folded sediments of this geosyncline. A brief outline of the evolution of the geosyncline, given below, is based on Fleming (1962).

The New Zealand Geosyncline apparently began to develop in the Carboniferous and was well defined as a locus of rapid sinking and deposition during the Permian. During its formation, fossiliferous rocks of marginal facies were deposited in shallow water along the western margin of the geosyncline. These are distinguishable from an axial facies of greywackes consisting of altered, folded, extremely thick, and poorly fossiliferous sediments (Wellman, 1956). During the Triassic the western foreland rose rapidly, supplying abundant sediment to the geosyncline. There is evidence for contemporaneous volcanicity on the mainland and granite emplacement in west Nelson. During the early Jurassic granite was intruded at the Bounty Islands. The geosyncline persisted through the Jurassic but became very restricted during the early Cretaceous as a result of the Rangitata Orogeny, during which geosynclinal sediments were folded and uplifted. The late Cretaceous of New Zealand was a time of widespread transgression over a land of low relief. Fresh- to brackish-water deposits of this age are found on Campbell Island, and the Camp Cove conglomerate on the Auckland Islands may be contemporary.

A fossil mid-ocean rise has been identified in the western Pacific, north of the Southwestern Pacific island-arc system (Hess, 1962). This feature, the Darwin Rise (Menard, 1964, 1965), had a north-west–south-east axial trend, parallel to the line of the New Zealand Geosyncline (Fig. 27), with which it thus has a spatial relation. The Darwin Rise went through a pronounced development phase 100 million years ago (Menard, 1964) at the same time as the Rangitata Orogeny (as dated by Aronson using radioactive-dating tech-

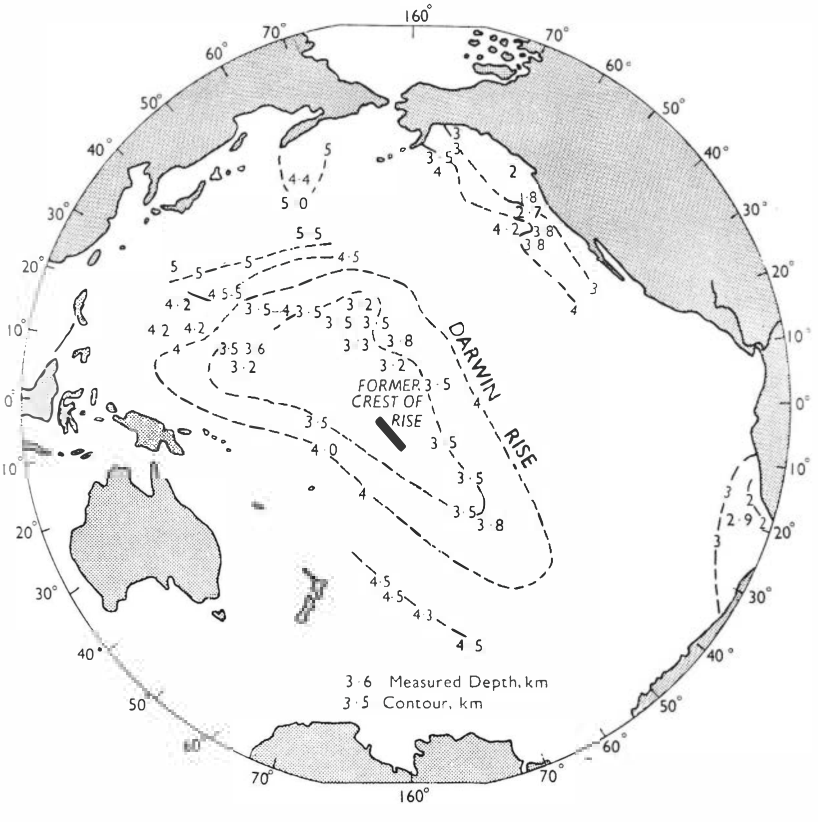


Fig. 27: Paleobathymetry of the Mesozoic Darwin Rise (from Menard, 1964, fig. 6.15) indicating its prominent north-west trend, parallel to the Mesozoic New Zealand Geosyncline.

niques, 1965). Darwin Rise had a pre-Cretaceous history of about another 100 million years according to Menard (1964).

In summary, it is evident that the stresses which gave rise to the development of a broadly north-west-trending ridge system, the Campbell Plateau - Lord Howe Rise and the New Zealand Geosyncline north of it, were primarily directed north-east–south-west (Fig. 24). Disruption of Gondwanaland by the development of the Indian Ocean mega-undation is thought to be Permo-Triassic and to have been followed by the north-easterly drift of New Zealand and associated continental fragments (van Bemmelen, 1965). At this time, the New Zealand Geosyncline began to develop (Fleming, 1962). During the mid-Mesozoic the Darwin Rise developed in the western Pacific. The Rangitata Orogeny which marked the end of the Mesozoic New Zealand Geosyncline was a late-Mesozoic event contemporaneous with a pronounced developmental stage of the Darwin Rise (Menard, 1964), a mid-ocean feature parallel to the geosyncline. Evolution of the New Zealand Geosyncline may have been a function of its position between two developing mid-ocean ridge features, each thought to be the centre of outward movement in terms of crustal drift.

Tertiary and Quaternary

During the Bortonian and Oligocene conformable foraminiferal ooze was deposited at Campbell Island in fairly deep water away from coastal influences (Fleming, 1962). Recent research (p. 31) has shown that foraminiferal ooze was deposited over much of the Campbell Plateau throughout the Tertiary. This type of foraminiferal ooze is also typical of parts of the South Island of New Zealand. The Eocene Amuri Limestone (Fleming, 1962) and the widespread Oligocene limestones of the type found at Oamaru (Gage, 1957) and other parts of the South Island (Fleming, 1962) indicate that both New Zealand and the off-shore Campbell Plateau were fairly stable areas probably submerged to similar depths and far removed from coastal regions during the early Tertiary.

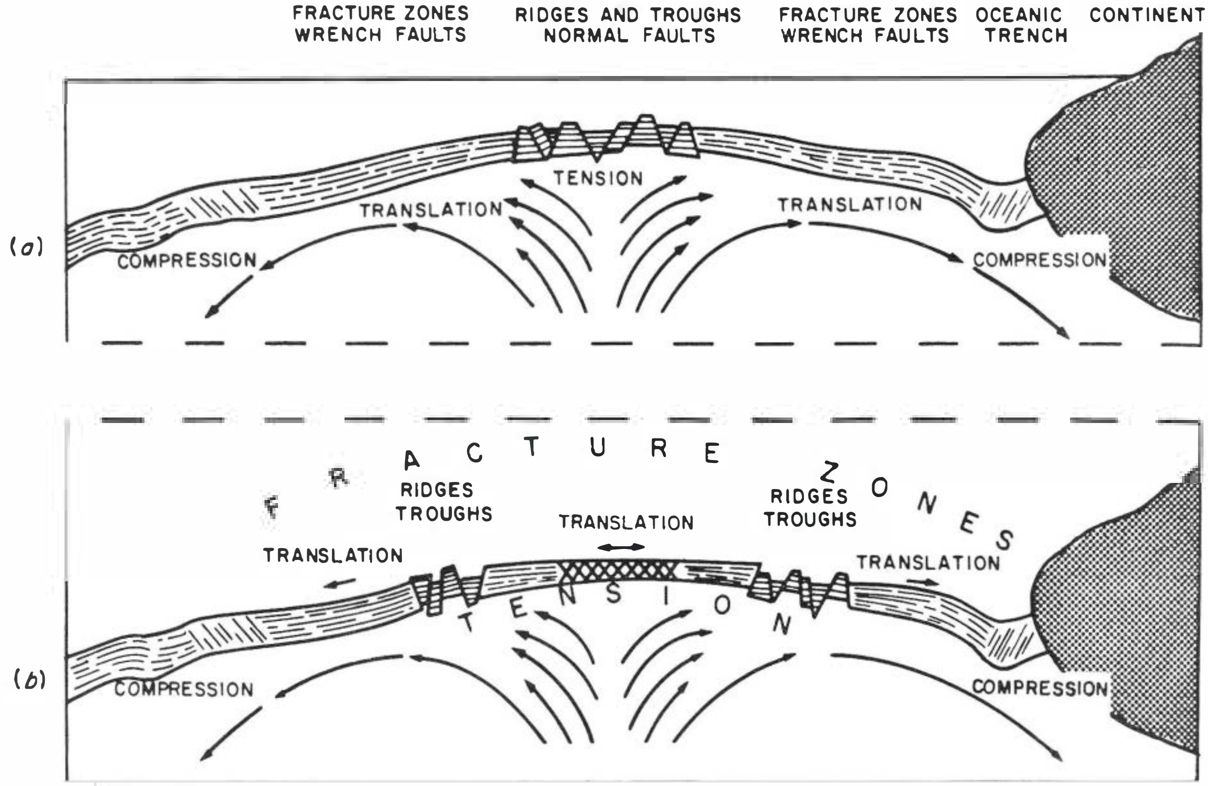


Fig. 28: Illustration of the convection-current hypothesis for the origin of the topography and structure of mid-ocean rises (from Menard, 1964, fig. 6.20).

During the Miocene, uplift and emergence of much of New Zealand began. Movement became more intense during the Pliocene towards the climax of the Kaikoura Orogeny. Off shore, on the Campbell Plateau, there is little indication of orogeny, apart from the development of broad rises on the Campbell Plateau culminating in Pliocene volcanism. Whereas the Campbell Plateau was geologically unstable during the early Mesozoic it formed a fairly stable tectonic unit from the end of the Cretaceous to the present. In contrast, the geology of the South Island of New Zealand, which was similar to that of the plateau in the early Tertiary, is one of mid-Tertiary - Quaternary tectonism. The edge of the continental shelf off the east coast of the South Island, the border of the Kermadec structural province, is formed by the Waipounamu Fault (Cullen, 1967). It is here suggested that a large amount of vertical movement occurred across this fault, possibly during the Kaikoura Orogeny after the early Miocene, as a result of uplift of the South Island. Uplift stopped the widespread formation of foraminiferal limestones over the South Island. A late Tertiary peneplain, formed possibly during the late Miocene, is recognised in southern New Zealand (Benson, 1941) and indicates a stable period between mid-Tertiary deformation and the intense movements of the Pliocene.

Developments in New Zealand are closely related to the Tertiary evolution of other structures in the Kermadec province, notably the Solander Trough - Macquarie Ridge (from the Oligocene onwards) and the Kermadec Ridge - Trench system, a presently active island arc of Tertiary age.

Menard observed that in the early Tertiary a mid-ocean ridge system was initiated in the east Pacific. This structure, which may be the locus of upwelling convection currents, appears to form the locus of outward movement of crustal blocks (Menard, 1964, 1965; Fig. 28). The western margin of the East Pacific Rise or, in other words, the edges of oceanic or continental crustal blocks which formerly lay along the crest of the rise but have now been displaced westward, may be recognised in the Pacific Ocean (Wilson, 1965). Wilson (1965) suggests that the Subantarctic Slope is part of the western margin of the East Pacific Rise and may have moved westward away from the rise crest, a view supported by van Bemmelen (1965). The concept of Tertiary drift in the eastern Pacific helps to explain regions of tensional block faulting along the southern margin of the Campbell Plateau. Parallelism of linear magnetic anomalies to the Subantarctic Slope on the floor of the Southwestern Pacific Basin and their displacement in a narrow zone crossing the basin from the mid-ocean rise to the Campbell Plateau suggest that crustal movement is taking place north-westwards away from the rise crest (compare Ross and Christoffel, in press). Although the Antipodes Fracture Zone forming the Subantarctic Slope may have been initiated in the Mesozoic as a transcurrent fracture, as suggested by Cullen (1967), it is apparent that the fracture zone now forms the edge of a large dilatational ocean basin—the Southwestern Pacific Basin—and it may have been initiated as a rift fracture.

South of the Campbell Plateau (Fig. 24) the East Pacific Rise curves west into the Pacific - Antarctic Ridge, continuous south of Australia with the Indian - Antarctic Ridge (Ewing and Heezen, 1956). This ridge system south of New Zealand is referred to as the Indian - Pacific Ridge. The fracture zone discovered by Ross and Christoffel and referred to above terminates near the crest of the Indian - Pacific Ridge.

Displacement across this sinistral fracture zone is about 200 miles near the Campbell Plateau, decreasing to almost zero near the crest of the

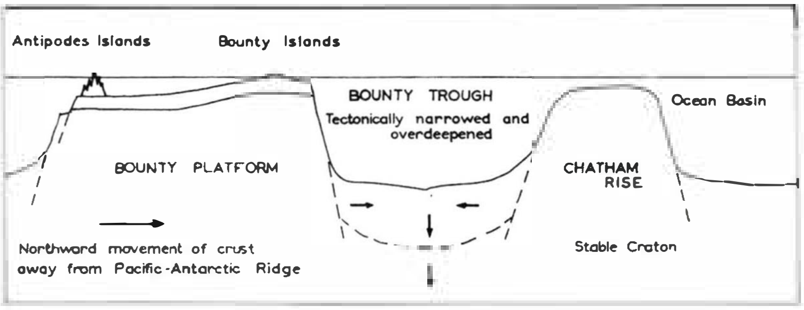


Fig. 29: Hypothetical crustal section across Campbell Plateau, Bounty Trough, Chatham Rise, indicating the possible mechanism of overdeepening of the eastern end of the Bounty Trough between the stable Chatham Rise and the northwardly moving Bounty Platform.

Indian - Pacific Ridge. This relative displacement accords with Menard's views that adjacent crustal blocks are moving away from the rise crest with different rates of movement. The Bounty Platform is displaced some 60 miles north along faults forming an extension of the fracture in the Pukaki Saddle (Fig. 24). Northward there is no sign of a continuation of this fracture affecting the Chatham Rise, and it is assumed that northward movement of the Bounty Platform has been taken up by downfolding and overdeepening of the eastern end of the Bounty Trough (Fig. 29).

The present distribution of island arcs and oceanic trenches, particularly in the Southwestern Pacific, may be closely related to formation of the mid-ocean ridge system (compare Menard, 1964). Parallelism of the Tonga - Kermadec Ridge with the East Pacific Rise is considered hardly fortuitous. It may be inferred that the deep crustal force causing the westward movement of crustal blocks away from the East Pacific Rise is causally related to the formation of island arcs and trenches in the Kermadec Structural Province.

Macquarie Ridge trends at 90° to the Indian - Pacific mid-ocean ridge section. Less pronounced seismicity and lack of deep-focus earthquakes beneath the Macquarie Ridge suggest that it is at a different stage of development from the Kermadec Ridge. This may be explained by the late geological development of the Indian - Pacific Ridge. The East Pacific Rise began to form during the Tertiary in the North Pacific (Menard, 1964). Its southern branch, the Pacific - Antarctic Ridge, continuous via the Indian - Antarctic Ridge with the Indian Ocean mid-ocean ridge, may be a later Tertiary development. This Indian - Pacific Ridge appears to disrupt a formerly continuous island-arc system, the Macquarie - Balleny Islands Ridge, which was probably formed parallel to the north–south-oriented section of the East Pacific Rise during the early Tertiary (compare van Bemmelen, 1965, fig. 6).

CONCLUSIONS

Two very different episodes of crustal deformation, contrasting in the direction of their stress patterns, have in turn dominated the New Zealand region. Firstly, Permian evolution of the mid-ocean rise in the Indian Ocean resulted in drift of New Zealand away from Australia. Later, during the Mesozoic, development of the Darwin Rise in the western Pacific led to crustal movement directed south-west toward New Zealand. The New Zealand Geosyncline was formed at the Pacific margin of a segment of north-eastwardly drifting continental crust, between the Indian Ocean mid-ocean rise and the Darwin Rise. During the Cretaceous Rangitata Orogeny, New Zealand geosynclinal sediments were deformed and uplifted. At this time the Darwin Rise went through a pronounced stage of active development.

Secondly, the present configuration of the Tonga - Kermadec island-arc system, of which New Zealand is part, is predominantly Tertiary, and probably arose from stresses caused by evolution of the Tertiary East Pacific Rise, a part of the mid-ocean ridge system to which it is parallel. Tertiary drift of New Zealand away from this feature is inferred and a tensional origin is indicated for the Antipodes Fracture Zone, probably a rift margin fracture.

Considerable overlap of volcanic activity is evident during the Tertiary - Quaternary period on belts with "Kermadec" and "North-western - Chatham" trends (in terms of the structural provinces of Brodie, 1958). Tertiary volcanicity occurs at Norfolk Island and in Northland, New Zealand, and on the Pukaki Bank and Campbell Island Rise—all "North-western - Chatham" structures. Similarly, volcanicity occurs along the Kermadec and Macquarie Ridges, in the Solander Trough, and along the Taupo Graben—all "Kermadec" structures. It has been suggested (Wright, 1966a) that the two major convective systems beneath the East Pacific Rise and the Indian Ocean Ridge, outlined above, were operative at the same time over parts of the Tertiary and Quaternary. Wright suggests that it was the interaction between these nearly opposed forces which may have caused shearing along the Alpine Fault. This seems to be a simple, adequate attempt to explain the occurrence of the fault, which would similarly explain the occurrence of strike-slip faults parallel to the Alpine Fault elsewhere in the Southwestern Pacific Basin (compare Fig. 24).

The theory of interrelation of movement patterns centred on mid-ocean rises provides a broad structural setting for New Zealand and related off-shore structures in the Southwestern Pacific Region. It involves the widely held assumption that mid-ocean ridges are the loci of outward movement of crustal blocks, possibly by a convective cell mechanism, and that island-arc systems are sited over down-turning convection currents.

Van Bemmelen (1965) infers that development of the Indian - Pacific Rise would result in anti-clockwise rotation of the New Zealand Region. Such rotation had previously been suggested by Carey (1958) in an attempt to explain the origin of oroclinal bends in the folds of the South Island, New Zealand. The oroclinal bend or recurved arc is typical also of the present island-arc system which is directed east from Tonga to New Zealand and west from New Zealand to Macquarie Island (Fig. 24). These recurved structures, typical of New Zealand, may be caused by interaction between the two major convection cells of the region (Fig. 24).

The disposition of fragments of the New Zealand Plateau relative to Antarctica and Australia, prior to drift, is thought similar to that shown in Fig. 30 (based on Carey, 1958, fig. 46c). This configuration is achieved by reversing movement on the Alpine Fault and unbending the New Zealand Orocline (Carey, 1958, figs. 47a, 47b, and 47c, p. 306). Northward drift of Australia and the opening of the Tasman Basin were probably mainly Mesozoic events. Disruption of the West Antarctica - New Zealand link occurred during the Tertiary as a result of the inception of the Pacific - Antarctic segment of the East Pacific Rise (compare Fig. 24).

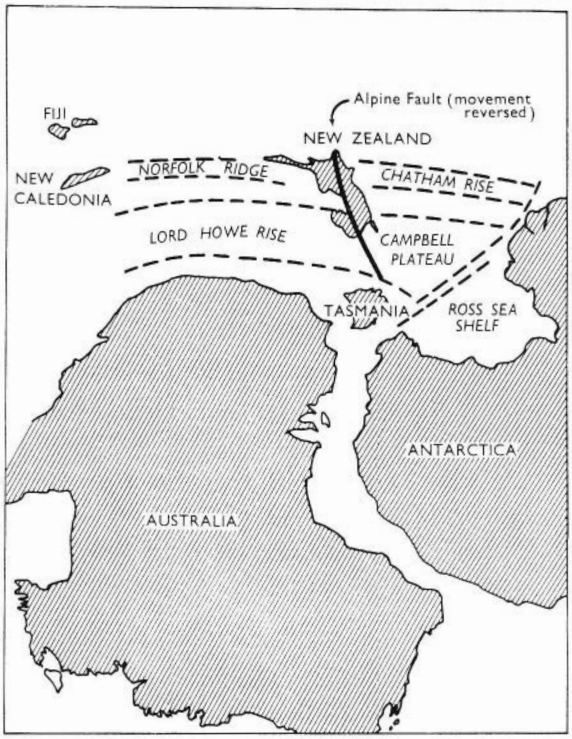


Fig. 30: The New Zealand region as part of Gondwanaland; a reconstruction involving reversal of the 300-mile displacement of the Alpine Fault, and unbending the New Zealand orocline (based on Carey, 1958, fig. 46c).

APPENDIX I

TABLES

Table 1. Chief Sources of Bathymetric Data Used in Chart Compilation

N.Z.O.I. Track No.	Vessel	Date	Cruise Name or General Area
T 1	HMNZS *Tutira*	1949	Subantarctic Islands
T 3	HMNZS *Pukaki*	1950	Campbell and Auckland Islands
T 5	HMS *Veryan Bay*	1950	Campbell Island
T 33	HDMS *Galathea*	1952	Campbell Island
T 142	HMNZS *Kiwi*	1954	Campbell and Auckland Islands
T 206	RNZFA *Tui*	1956	Campbell Island
T 218–9	HMNZS *Pukaki*	1956	Ross Sea voyage
T 223	HMNZS *Hawea*	1956	Ross Sea voyage
T 229	RNZFA *Tui*	1957	Cruises III and IV
T 258	HMNZS *Kaniere*	1958	Southern ocean
T 280	HMNZS *Endeavour*	1959	Ice edge
T 284	HMNZS *Endeavour*	1960	Ross Sea
T 292	HMNZS *Lachlan*	1956	Subantarctic region
T 319–20	USS *Wilhoite*	1960	Subantarctic region
T 334	MV *Taranui*	1962	Subantarctic region
T 336	HMNZS *Rotoiti*	1962	Macquarie Ridge
T 409	RV *Vema*	1960	Subantarctic region
T 419	MV *Taranui*	1962	Bounty Platform
T 425–7	HMNZS *Endeavour*	1962	Supply trip
T 434–8	HMNZS *Endeavour*	1963	Macquarie benthos
T 441	HMNZS *Endeavour*	1963	Macquarie benthos
T 445	RNZFA *Tui*	1963	New Zealand southern waters
T 460	MV *Taranui*	1963	Campbell benthos
T 494–501	HMNZS *Endeavour*	1964	Macquarie gap
T 506	RV *Argo*	1964	Subantarctic region
T 508–11	HMNZS *Endeavour*	1963–4	Supply trip
T 512	USS *Glacier*	1964	Supply trip
T 526	RRS *Discovery II*	1936–8	Subantarctic region
T 528	MV *Bear*	1933–5	Subantarctic region
T 549	HMNZS *Endeavour*	1964	Supply trip
T 553	*Umitaka-Maru*	1965	Subantarctic region
T 556–8	HMNZS *Endeavour*	1965	Campbell Plateau
T 560	USS *Eltanin*	1965	Shakedown cruise
T 561	USS *Eltanin*	1964–5	Macquarie Ridge
T 565–6	HMNZS *Endeavour*	1965	Supply trip
T 527	RRS *Discovery II*	1950	Subantarctic region
T 613	HMNZS *Endeavour*	1966	Supply trip

The United States Oceanographic Office supplied the following depth records from the area, all of which were used in this study. The N.Z.O.I. traverse numbers for these records are T 343, 344, 353, 363, 364, 369, 370, 378, 383, 401, 408, 410, 411, 413, 414.

Table 2. Stratigraphic Successions in the Subantarctic Islands of the New Zealand Region

Geologic Period	(a) The Snares	(b) Auckland Islands	(c) Campbell Island	(d) Antipodes Islands	(e) Bounty Islands	(f) Macquarie Island	(g) Solander Island
QUATERNARY				(?) Basalts			Hornblende Andesite
TERTIARY							
Pliocene		Basalts	Basalts				
Miocene		Sandstones				Basalt + lst.	
Oligocene			Limestone				
Eocene							
Paleocene							
MESOZOIC							
Cretaceous		(?)Conglomerate	Conglom. + mst.			(?)Gabbro	
Jurassic					Biotite granite		
Triassic		(?)Trachytes	(?)Gabbro			(?)Basalt + seds.	
PALEOZOIC	(?)Granite (?)Schist	(?)Gabbro (?)Granite (?)Gneiss + Schist	(?)Schist				

(a) Fleming, Reed, and Harris, 1953; (b) Speight and Finlayson, 1909; Fleming, 1959 and in press; (c) Oliver, Finlay, and Fleming, 1950; (d) Postulated by present author; (e) Wasserburg, Craig, Menard, Engel, and Engel, 1963; (f) Mawson and Blake, 1943; (g) Harrington and Wood, 1958.

TABLE 3. Chemical Composition of Selected Samples from the South Island – Campbell Plateau Petrographic Province

	1	2	3	4	5	6	7	8	9	10
SiO_2	46·01	45·48	45·77	53·40	52·36	49·44	47·41	49·16	46·18	47·14
Al_2O_3	16·34	15·56	12·98	15·94	14·20	11·38	16·18	15·27	16·82	19·35
Fe_2O_3	3·63	4·25	3·17	9·14	5·80	2·41	12·92	8·64	2·08	5·61
FeO	8·81	8·24	8·86	3·59	6·68	12·53	8·37	7·22	13·64	10·82
MgO	5·19	6·97	11·00	3·37	3·09	12·26	1·16	8·34	5·68	3·98
CaO	8·62	9·35	9·67	8·57	7·40	7·30	4·03	5·62	4·81	5·42
Na_2O	4·07	3·78	2·33	2·18	5·42	2·91	3·49	2·72	5·02	2·38
K_2O	1·53	1·40	0·83	1·19	2·06	1·19	2·15	0·61	1·83	1·60
TiO_2	2·86	2·94	2·05	2·60	2·40	0·72	0·31	0·48	0·51	0·68
P_2O_5	0·92	0·83	0·40	..	..	0·12	Tr.	Tr.	1·05	0·14
MnO	0·18	0·15	0·20	..	..	0·32	0·27	0·29	0·27	0·18
CO_2	0·03	0·10	0·10	..	..	Tr.	0·21	0·71	0·31	0·26
H_2O+	1·26	0·39	1·40	1·51	1·57	Tr.	1·38	0·98	0·82	0·96
H_2O-	0·34	0·59	0·98			0·12	1·42	0·44	0·89	0·84

Tr. = trace

1. Akaroa Group Basalt, Banks Peninsula (Gregg and Coombs, 1966).
2. Diamond Harbour Group Basalt, Banks Peninsula (Gregg and Coombs, 1966).
3. Olivine Dolerite, Dunedin Volcanic Complex (Coombs, 1966).
4. Basalt, Mt. Honey, Campbell Island (Marshall, 1909).
5. Basalt, Beeman Hill, Campbell Island (Marshall, 1909).
6. Gabbro, Auckland Islands (Speight and Finlayson, 1909).
7. Older Basic Series Dolerite, Auckland Islands (Speight and Finlayson, 1909).
8. Older Basic Series Olivine Dolerite, Auckland Islands (Speight and Finlayson, 1909).
9. Younger Basic Series Basalt, Auckland Islands (Speight and Finlayson, 1909).
10. Younger Basic Series Dolerite, Auckland Islands (Speight and Finlayson, 1909).

TABLE 4. Chemical Analyses of Selected Samples from the Macquarie Tholeiite Province and Other Tholeiite Localities

	1	2	3	4	5	6	7	8	9	10
SiO_2	46·75	47·56	46·57	47·30	46·70	48·90	50·71	46·53	49·34	49·70
Al_2O_3	15·95	15·50	22·93	15·30	15·80	19·00	12·89	14·31	17·04	15·30
Fe_2O_3	3·84	1·88	0·87	3·40	4·30	3·90	1·85	3·16	1·99	2·80
FeO	4·49	6·82	4·43	5·20	4·00	3·30	9·55	9·81	6·82	9·40
MgO	9·33	7·77	6·02	9·00	10·00	5·10	7·68	9·54	7·19	7·70
CaO	10·91	10·92	12·09	9·30	10·30	12·80	10·81	10·32	11·72	12·00
Na_2O	2·25	2·53	2·60	3·60	2·70	3·30	2·15	2·85	2·73	1·30
K_2O	0·33	0·27	0·50	0·40	0·30	0·40	0·45	0·84	0·16	0·30
TiO_2	1·55	1·41	0·60	1·13	1·14	1·22	3·08	2·28	1·49	0·60
P_2O_5	0·43	0·20	0·08	0·12	0·17	0·18	0·29	0·28	0·16	0·10
MnO	0·15	0·13	0·04	0·14	0·10	0·10	0·12	0·18	0·17	0·30
CO_2	..	1·63	..	..	..	..	..	..	..	..
H_2O+	2·59	2·87	2·85	2·90	2·20	0·70	0·28	0·08	0·69	0·30
H_2O-	1·40	0·68	0·15	1·80	2·00	0·50	0·10	..	0·58	0·30
Fe_2O_3/FeO	0·86	0·28	0·20	0·65	1·08	1·18	0·19	0·32	0·29	0·30
Na : K	6·09	8·36	4·64	8·03	8·03	7·56	4·26	3·03	16·00	3·87

1. Olivine dolerite, Macquarie Island (Mawson and Blake, 1943, VI, p. 120).
2. Aphyric basalt, Macquarie Island (Mawson and Blake, 1943, II, p. 131).
3. Bytownite basalt, Macquarie Island (Mawson and Blake, 1943, III, p. 131).
4. Olivine basalt (D 7). Analyst: J. Ritchie; Chemistry Division, D.S.I.R., Gracefield.
5. Olivine basalt (D 17). Analyst: J. Ritchie, Chemistry Division, D.S.I.R., Gracefield.
6. Olivine basalt (D 18). Analyst: J. Ritchie, Chemistry Division, D.S.I.R., Gracefield.
7. Average of 28 analysed lavas from Hawaii (Tilley, 1950, table I, col. I, p. 41).
8. Alkali basalt (Yoder and Tilley, 1962, table 2, p. 361–362).
9. Average oceanic tholeiitic basalt (Engel *et al.*, 1965. AD 2, table 2, p. 721).
10. Tholeiitic basalt, Japan (Kuno, 1959, table 3, p. 48).

TABLE 5. Occurrence of Different Lithologies

NZOI Sta. No.	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P	Q	R	S	T	
B 583																		x			New Zealand Continental Shelf
B 587									x												
B 589																			x		
D 100									x		x										
D 132																	x				
D 153																			x		
D 155									x		x										
B 586									x												
D 39A	x	x					x		0									x			Auckland Islands Shelf and Slope
D 76	x																				
D 80	x											0									
D 103			x					x													
D 170	x																				
D 172	x																				
D 173																		x			
D 198	x																				
F 81																x					
B 172								0	0								x				Snares Depression and Western Campbell Plateau
D 101										0							x				
D 138								0													
D 160								0										0			
D 176	x						x			0											
F 80													0								
B 183	x																				Campbell Island Shelf and Rise
F 139	x																				
F 144																				0	
B 195	x						x	x													Pukaki Bank
D 209							x														
D 210	x		x																		
F 106								x									x				
A 696			x	x				x													Macquarie Ridge
C 734			x					x													
D 5	x						x	0	0	0	0		0	0	0	0					
D 6	x		x				x														
D 7	x																				
D 8			x		x	x															
D 9					x	x															
D 17	x																				
D 18	x							x													
D 20			x			x															
D 22	x																				
D 24			x																		
D 25			x																		
D 26	x		x																		
D 159	x							x										x			
D 169	x							x	0	0					0						
E 234					x			x													
E 236				x	x																
F 127									0	0						0	0				Pukaki Saddle
F 122																	x				Bounty Platform
F 129								0	0	0											
F 132								x									x				

1. Basic Igneous Rocks

A = Olivine basalt
B = Olivine dolerite
C = Basalt
D = Dolerite
E = Gabbro
F = Peridotite
G = Basic volcanic agglomerate
H = Undifferentiated basic volcanics

2. Acid Igneous, Sedimentary, and Metamorphic Rocks

I = Granite
J = Gneiss
K = Schist
L = Undifferentiated acid metamorphics
M = Sandstone
N = Greywacke
O = Mudstone
P = Conglomerate
T = Pumice

3. Calcareous Rocks

Q = Foraminiferal limestone
R = Polyzoan limestone
S = Molluscan limestone

x = in situ or locally derived rock fragments
0 = exotic, probably transported, rock fragments

TABLE 6. Chemical and Normative Analyses of Rocks From Macquarie Ridge

CHEMICAL ANALYSES	D6	D7	D17	D18
SiO_2	48·60	47·30	46·70	48·90
Al_2O_3	16·90	15·30	15·80	19·00
Fe_2O_3	1·60	3·40	4·30	3·90
FeO	6·80	5·20	4·00	3·30
MgO	7·80	9·00	10·00	5·10
CaO	10·90	9·30	10·30	12·80
Na_2O	2·80	3·60	2·70	3·30
K_2O	0·05	0·40	0·30	0·40
TiO_2	0·90	1·13	1·14	1·22
P_2O_5	0·19	0·12	0·17	0·18
MnO	0·14	0·14	0·10	0·10
H_2O+ 110 °C	2·45	2·90	2·20	0·70
H_2O- 110 °C	0·15	1·80	2·00	0·50
Na : K	..	8·03	8·03	7·36
Fe_2O_3 : FeO	0·23	0·65	1·08	1·18
NORMATIVE ANALYSES				
Ilmenite	1·67	2·13	2·28	2·28
Apatite	0·34	0·34	0·34	0·34
Magnetite	2·32	5·10	6·50	5·57
Orthoclase	0·28	2·22	1·67	2·22
Albite	23·58	30·39	23·58	27·77
Nepheline	..	0·57	..	..
Anorthite	33·92	24·74	30·58	36·42
Diopside (Wo)	8·40	9·05	8·70	11·25
(En)	5·25	6·67	7·05	9·10
(Fs)	2·64	1·48	0·59	0·79
Hypersthene (En)	3·95	..	3·90	0·65
(Fs)	7·26	..	5·41	0·92
Olivine (Fo)	3·04	11·34	3·92	0·84
(Fa)	6·12	2·84	6·12	1·43

TABLE 7. Earthquake Epicentres in the New Zealand Subantarctic Area

Date		Lat.(S)	Long.(E)	Type	Magnitude
1962					
Jan	19	48° 45′	164° 45′	N	5·3 D
Sep	29	49° 00′	165° 00′	N	5·0 C
1963					
Apr	26	48° 00′	164° 00′	N	4.7 D
Aug	14	47° 15′	165° 00′	S?	5·3 D
Aug	23	50° 25′	166° 00′	N	4·9 D
Oct	14	48° 30′	164° 00′	N	4·4 D
1964					
Jan	11	49° 00′	165° 00′	N	? D
Mar	16	47° 42′	166° 48′	S	4·4 D
Mar	17	49° 30′	165° 00′	N	4·8 D
Mar	26	47° 45′	166° 30′	S	4·1 D
Apr	2	47° 30′	166° 00′	N	5·0 D
May	23	50° 00′	165° 00′	N	4·9 D
Jun	3	49° 30′	164° 30′	N	4·1 D
Jul	30	49° 30′	165° 30′	N	5·0 D
Sep	15	49° 00′	164° 00′	N	5·0 D
Sep	23	49° 00′	165° 00′	N	5·0 D
Oct	2	49° 48′	165° 30′	N	4·8 D
Oct	2	49° 30′	165° 30′	N	4·7 D
Oct	6	48° 00′	165° 30′	N	5·0 D
Oct	9	49° 36′	164° 18′	N	5·0 D
Oct	11	49° 18′	164° 30′	N	4·7 D
Oct	17	50° 00′	165° 00′	N	4·9 D
Oct	24	49° 30′	165° 00′	N	4·9 D

Values derived from N.Z. Seismological Records, 1962–64.

N = normal focus
S = shallow focus
C = 10–15 miles radius of error.
D = greater than 15 miles radius of error.

TABLE 8. Sediment Sample Types and Localities

NZOI Sta.	Lat. (°′S)	Long. (°′E)	Depth (m)	$CaCO_3$ (%)	Sediment Type	Authigenic Minerals
MACQUARIE RIDGE SAMPLES						
D 5	56 40·6	158 45·5	1280	..	not analysed	M
D 6	55 29	158 31·5	415	..	not analysed	
D 8	54 52	158 39	141	..	v.c. shell sand	
D 18	52 31	160 31	128	..	c.m. bryoz. and shell debris	
D 22	50 38	163 57	755	..	shelly foraminiferal sand	
D 159	49 01	164 30	741	82	c. bryoz. and shell sand	
D 169	50 47	163 57·5	896	..	foraminiferal sand	M
AUCKLAND ISLANDS SHELF AND SLOPE (ABOVE 600 m)						
D 39a	50 58	165 45	549	*90·4	c. shell, bryoz. and coral debris	
D 72	50 19·8	166 24	163	82·5	m. shell debris	
D 73	50 18	166 23·5	177	84·5	c.m. shell debris	
D 77	50 44·7	165 52·5	168	*54·1	slightly muddy fine sand	
D 78	50 39·6	166 01·8	132	*35·7	slightly muddy fine sand	
D 79	50 37	165 53	183	..	not analysed	
D 103	50 40	166 19·5	71	..	c.m. bryoz. and shell debris	
D 104	50 49·2	166 15·6	95	*99	v.c. bryoz. and shell debris	
D 148	49 48	167 02·5	146	95	c. shell debris	
D 173	50 53	166 32	141	90	c. bryoz. and shell debris	
D 174	50 46·5	166 56	474	83·5	globigerina ooze	G
D 175	50 36·5	167 41	426	81·5	globigerina ooze	G
D 200	50 22	167 28	113	92	v.c. bryoz. debris	
F 82	50 01	166 54	137	92	v.c. bryoz. debris	
F 83	50 26	166 54	117	93	c.m. bryoz. and shell debris	
F 84	50 46	167 07	417	79	globigerina ooze	G
F 85	51 02	167 09	474	83·5	globigerina ooze	G
F 88	50 31·5	168 00	439	87·5	globigerina ooze	G
F 89	50 01	167 49	371–90	91·2	v.c. bryoz. and shell debris	

TABLE 8. Sediment Sample Types and Localities—*continued*

NZOI Sta.	Lat. (°′S)	Long. (°′E)	Depth (m)	$CaCO_3$ (%)	Sediment Type	Authigenic Minerals
CAMPBELL ISLAND SHELF AND RISE (above 600 m)						
B 182	52 28·5	168 32	238	72	sandy foraminiferal ooze	
B 187	53 02·6	172 14	393	90	globigerina ooze	
B 192	52 25·2	169 21	192	72	slightly pebbly c.m. shelly sand	
B 193	52 21	169 21·5	192	76	slightly muddy fine shell sand	
B 194	51 53	169 37	265	78	shelly fine sand	
D 30	52 33·5	169 13	40	*50	sandy pebbly mud	
D 32	52 08	168 50	188	88	c.m. shell debris	
D 33	53 00	172 00	408	88·5	globigerina ooze	
D 35	52 56·4	169 33	188	. .	c.m. shell debris	
F 138	52 03	170 23	193–87	81	globigerina ooze	G
F 141	52 38	169 23	176	*82	v.c. pebbly shell sand	
F 143	53 05·5	170 13	380	83	foraminiferal sand	G
F 145	53 14	171 48	435	87·5	globigerina ooze	G
F 146	53 00	172 45	435	87·5	globigerina ooze	G
PUKAKI BANK AND RISE (above 500 m.)						
D 210	49 21	171 53	353	. .	globigerina ooze	G
F 105	49 34·5	170 57	499	81·5	globigerina ooze	G
F 106	49 30	172 00	371	77	globigerina ooze	G
F 150	48 28	174 35	500	87·5	globigerina ooze	
B 195	49 14·5	171 45	82	88·5	biogenic skeletal debris	
D 87	49 56	171 50	483	81·5	globigerina ooze	G
BOUNTY ISLANDS SLOPE (above 500 m.)						
F 120	48 18	179 16	494–512	90	globigerina ooze	
F 122	48 06	179 57W	252	72·5	sandy globigerina ooze	
CAMPBELL PLATEAU (below c. 500 m)						
B 32	53 38	169 52	799	89		
B 33	52 00	167 30	748	82·5	globigerina ooze	
B 172 to	48 18·5 48 22·9	168 32 168 30	732	88·5	shelly foraminiferal sand	G
D 84	49 46·5	168 56	649	85	globigerina ooze	
D 85	49 50	170 13	611	. .	globigerina ooze	
D 88	49 53	173 30	519	. .	globigerina ooze	
D 134	48 16	168 43·5	668	75	foraminiferal sand	GP
D 135	48 17·5	169 12	631	86	globigerina ooze	G
D 136	48 33·5	169 10	713	86	globigerina ooze	
D 137	48 50·5	169 07	668	86	globigerina ooze	
D 138	48 32	168 19·5	668	89·5	globigerina ooze	G
D 146	49 11	167 51·5	667	86·5	globigerina ooze	G
D 147	49 31	167 25	574	85·5	globigerina ooze	GP
D 160	49 31·5	166 15·5	722	85·5	foraminiferal sand	GP
D 176	51 06	167 48·5	582	*93·5	c.m. shell debris	
D 177	51 25	167 49·5	640	87	globigerina ooze	
D 178	51 43	167 50	629	87·5	globigerina ooze	
D 179	51 25·5	167 21	611	91	globigerina ooze	
D 202	51 03·5	168 33	611	86	globigerina ooze	
D 203	51 00	169 29·5	556	86	globigerina ooze	
D 204	50 58·5	170 16	565	. .	globigerina ooze	
D 205	50 57·5	171 16	529	85·5	globigerina ooze	
D 206	50 36	171 23·5	529	. .	globigerina ooze	
D 207	50 04	171 23	510	81	globigerina ooze	
D 211	48 53	172 17·5	519	89	globigerina ooze	
F 80	49 00	167 01	621	82	c.m. shell debris	
F 90	49 30·5	167 40	601	88	foraminiferal sand	GP
F 91	49 00	167 30	676	84	globigerina ooze	G
F 94	48 31	168 01	604	. .	foraminiferal sand	
F 96	48 30	168 38	697	. .	globigerina ooze	G
F 98	48 01	168 55	612	. .	globigerina ooze	
F 99	48 32	168 54·5	706	84·5	globigerina ooze	G
F 100	49 02	168 53·5	733–46	. .	globigerina ooze	
F 101	48 08	169 23	589	. .	globigerina ooze	G
F 102	48 39	169 51	810	88	globigerina ooze	
F 103	48 03	170 38	1280	86·5	globigerina ooze	
F 104	48 40	170 48·5	814–788	99·5*	globigerina ooze	
F 107	48 45	172 00	658	90·5	globigerina ooze	
F 108	48 19	171 59	1108	90·5	globigerina ooze	

TABLE 8. Sediment Sample Types and Localities—*continued*

NZOI Sta.	Lat. (°′S)	Long. (°′E)	Depth (m)	$CaCO_3$ (%)	Sediment Type	Authigenic Minerals
CAMPBELL PLATEAU (below c. 500 m.)—*continued*						
F 109	49 11	173 00	501	88·5	globigerina ooze	
F 110	48 07	174 02	1167	90·5	globigerina ooze	
F 111	48 57	174 58·5	704	91	globigerina ooze	
F 113	48 56	177 02	1372	89·5	globigerina ooze	
F 114	48 58	178 02	1072	90	globigerina ooze	
F 115	49 18·5	179 52	1518	84	foraminiferal sand	G
F 116	49 40	179 53	2286	87	foraminiferal sand	M
F 117	49 40	179 00	1202–93	60·5	pebbly foraminiferal sand	
F 121	48 14·5	178 06	975	82	globigerina ooze	
F 125	48 32	177 59	1063	..	globigerina ooze	
F 126	49 48	176 01	1256	94·5	foraminiferal sand	
F 127	49 22	176 16	1280	..	foraminiferal sand	MP
F 128	49 09	177 18	978	92·5	globigerina ooze	G
F 129	49 24	177 59	978	88	globigerina ooze	GMP
F 132	49 59	177 32	1335	..	foraminiferal sand	M
F 133	50 19	176 23	2377	..	manganese crust	M
F 135	50 58	173 57	832	..	globigerina ooze	
F 136	51 20	172 42	547	91	globigerina ooze	
F 137	51 42	171 31	519	87·5	globigerina ooze	
F 144	53 29	170 56	596	85·5	globigerina ooze	G
F 147	52 21	173 09	611	91	globigerina ooze	
F 148	51 43	173 32	677	92	globigerina ooze	G
F 149	50 31	174 19	1026	91	globigerina ooze	
F 151	48 32	174 50	814	91	globigerina ooze	
CONTINENTAL SHELF AND SLOPE (above 500 m)						
B 579	48 00	168 34	145	..	3″ diam. bryoz. nodules + bryoz. debris	
B 580	48 00	168 20	140	*95	v.c. bryoz. and shell debris	
B 581	48 00	168 06	138	..	1″ diam. bryoz. nods. + bryoz. and shell debris	
B 582	48 00	167 38	142·5	..	1–1·5″ diam. bryoz. nods. + bryoz. and shell debris	
B 583	48 00	167 20	143·5	..	+ 0·5″ diam. bryoz. nods. + bryoz. and shell debris	
B 585	48 00·1	166 35·2	80·5	..	v.c. bryoz. debris.	
B 586	48 00	166 26·5	300	*95	c.v.c. bryoz. and shell debris	
B 587	48 00·2	166 39	155	*88	1″ bryoz. nodules	
B 588	48 00	166 53	81	*94·4	v.c. bryoz. and shell debris	
B 589	48 44	166 30	188	*92·5	c.v.c. bryoz. and shell debris	
B 590	48 46	166 49	159	..	v.c. bryoz. and shell debris	
B 591	48 46	167 05	142	..	0·5″ bryoz. nodules and shell debris	
B 592	48 46	167 19	152	94	c.v.c. bryoz. and shell debris	
B 593	48 43	167 32	161	..	1–0·5″ bryoz. nodules and shell debris	
D 100	48 02	166 36	157	89	v.c. bryoz. and shell debris	
D 131	48 02	167 03	128	..	v.c. bryoz. and shell debris	
D 132	48 06	167 36·5	130	91	v.c. bryoz. and shell debris	
D 133	48 11·5	168 21	137	..	v.c. bryoz. and shell debris	
D 139	48 20·5	167 46·5	137	85	c. shell debris	
D 144	48 31	167 17	128	91	c.v.c. bryoz. and shell debris	
D 145	48 42	167 27	356	93·5	0·5″ bryoz. nods. and bryoz. and shell debris	
D 149	49 10·5	166 51	448	73	v.c. bryoz. and shell debris	
D 151	48 12	166 38	146	91	fine shell sand	
D 153	48 15·5	166 16	347–256	94	fine shell sand	
D 155	48 02	166 38	450 ft	82·5	v.c. bryoz. and shell debris	
D 156	48 01·5	166 35	250–350 ft	..	bryoz. nodules and bryoz. debris	
F 78	48 32	167 09	135	91	v.c. bryoz. debris	
F 93	48 31	167 30	137	91	v.c. bryoz. debris	
F 97	48 00	168 32	130	92	v.c. bryoz. debris	

Visual estimates of Wentworth grades are denoted by—
c.v.c. = coarse to very coarse; v.c. = very coarse; c. = coarse; c.m. = coarse to medium; m. = medium.

Authigenic minerals are denoted by—
G = glauconite; M = manganese; P = phosphorite.

Abbreviations are—
diam. = diameter; bryoz. = bryozoan; nods. = nodules; ft = feet.

Carbonate Analyses by R. Titcombe.

*Analyses by Miss L. M. O'Cain.

TABLE 9. Qualitative Analyses of Non-carbonate Fractions from Selected Samples of Globigerina Ooze

N.Z.O.I. Sta.	Detrital Minerals	Siliceous Organisms	Dark Green Glauconite	Foraminifer Casts
B 172	C	MC	R	MA
D 138	C	R	C	C
D 146	MA	R	R	C
D 147	R	R	MC	MA
D 160	MA	R	R	C
D 174	R	R	MC	MA
D 175	R	R	MC	MA
F 84	C	R	R	C
F 85	MC	R	—	A
F 88	C	MC	—	MA
F 90	R	R	MC	MA
F 91	MA	C	R	C
F 99	MA	R	R	MA
F 105	MA	R	—	MA
F 115	R	R	C	C
F 122	A	—	R	—
F 128	A	R	—	MC
F 129	MC	—	—	A
F 138	MC	R	MC	MA
F 143	MA	R	—	MA
F 144	MA	C	—	C
F 145	C	MC	—	C
F 146	MA	C	—	MC
F 148	MA	R	—	MC

A = 75–100%. MA = 50–75%. C = 25–50%.
MC = 5–25%. R = 0–5%.

TABLE 10. Globigerina Ooze; Sediment Parameters

N.Z.O.I. Sta.	Wt % Non-carbonate Fraction	Md Total Sediment	Md Non-carbonate Fraction	σ Total Sediment	σ Non-carbonate Fraction	α Total Sediment
F 102	7·22	3·49	>4·12	2·28	—	+0·38
F 106	14·21	2·90	2·92	0·90	0·84	−0·16
F 107	5·06	2·75	3·15	0·91	1·03	−0·11
F 111	4·36	3·08	3·03	0·85	1·05	−1·06
F 113	4·60	3·62	4·01	1·47	1·15	−0·23
F 137	9·07	3·65	>4·04	1·45	—	−0·34
D 84	8·89	3·33	4·04	2·45	1·14	−0·49
D 175	18·50	2·73	—	1·05	—	−0·11
D 202	9·36	3·75	>4·08	1·96	—	−0·40
D 206	12·81	4·60	>4·20	2·12	—	—
D 160	13·75	2·40	2·58	0·70	0·57	−0·20

Sample D 160 is a foraminiferal sand: all the other samples are globigerina oozes. All measurements are in phi units derived from cumulative distribution curves (Figs. 7 and 8).

Md = median diameter
σ = dispersion
α = skewness (positive or negative).

TABLE 11. Chemical Composition of Manganese Nodules

	F 129	Other Areas
CHEMICAL ANALYSES		
MnO_2	29·7	31·7
Fe_2O_3	20·2	24·3
SPECTROGRAPHIC ANALYSES		
Si	1·5	9·4
Ca	1·0	1·9
Al	0·6	2·9
Co	0·6	0·35
Ni	0·25	0·99
Ti	1·0	0·67
Cr	0·002	0·001
Sr	*c.*0·1	0·081
Ba	0·1	0·18
V	0·07	0·054
Zr	0·03	0·063
Mg	0·7	1·7
Mo	0·015	0·052
Na	*c.*0·1	2·6
B	0·04	0·029
Be	0·001	..
Pb	0·1	0·09
Sn	0·01	..
La	0·03	0·016
Yb	0·01	0·003
Y	0·01	0·033

Not detected in F 129
Cu, In, Tl, Nb, Zn, Bi, W, Ag, Sb, As, Cd, Eu, Gd, Sc, Sm, Er, Nd, Tm, Tb, Lu, Pt.

Analysis of F 129 by D.S.I.R. Chemistry Division, Gracefield.

Average analyses of MnO_2 and Fe_2O_3 from other areas, derived from Mero, 1964, table 27, p. 179.

Average analyses of minor constituents of Pacific manganese nodules (other areas), derived from Mero, 1964, table 28, p. 180.

TABLE 12. Partial Chemical Analyses of Phosphorite Nodules

N.Z.O.I. Sta.	P_2O_5	F
D 147	20·60	2·90
D 134	23·10	3·10
F 127	26·20	3·90
F 90	19·40	2·60
F 122	10·00	1·30
X	21·80	2·56
Y	29·56	3·31

Composition is recorded in wt %.

X = Discovery II St. 2733, on Chatham Rise (Reed and Hornibrook, 1952).

Y = Sample from Forty Mile Bank, California Borderland (Dietz, Emery, and Shepard, 1942).

TABLE 13. Trace Element Analyses (p.p.m.)

Element	Oceanic Tholeiite	D 17	D 18	Alkali Basalt	D 7
B	..	15	30	..	20
Cu	77	100	70	36	150
Cr	297	500	500	67	500
Ni	97	300	100	51	250
Co	32	70	100	25	100
V	292	200	250	252	250
Y	43	30	30	54	30
Sc	61	30	40	26	30
Zr	95	100	150	333	100

Mo, Sn, Pb, La, Ba were below the detection limits of 1, 2, 5, 50, and 100 respectively in Macquarie Ridge Rocks.

D 17: Tholeiitic basalt, Macquarie Ridge. Analysts—D.S.I.R. Chemistry Division.

D 18: Tholeiitic basalt, Macquarie Ridge. Analysts—D.S.I.R. Chemistry Division.

D 7: Alkali basalt, Macquarie Ridge. Analysts–D.S.I.R. Chemistry Division.

Oceanic tholeiite: Av. comp. from table 2, Engel *et al.*, 1965.

Alkali basalt: Av. comp. from table 2, Engel *et al.*, 1965.

APPENDIX II

DESCRIPTIONS OF ROCK SAMPLES

AUTOCHTHONOUS ROCKS

Macquarie Ridge
Auckland Islands Shelf
Campbell Island Shelf
Pukaki Bank
The Slope Surrounding the Auckland Islands Shelf
Campbell Plateau
New Zealand Continental Shelf

TRANSPORTED ROCKS

INTRODUCTION

Visual estimates of bulk composition have been made for each sample. Suitable examples of different lithologies were selected for petrographic analysis, and slides were prepared by the writer in the Oceanographic Institute laboratory. Rocks from the Macquarie Ridge were mainly basic volcanics, many of which were basalts. Some of their distinctive textural characters, noted occurrences of glass and palagonite, and brief details of sample composition and mineralogy are listed below.

Some textural and mineralogical characters of basalts from the Macquarie Ridge—

Variolitic Structure: D 26, D 25, D 24, D 20, A 696.
Flow texture: D 5, D 169.
Vescicular: C 734, C 17, D 18, D 22, D 159.
Amygdaloidal: D 26, D 25, D 17, D 159, A 696.
Glassy: D 5, D 7, D 24, D 25, D 26, D 17, D 18, D 20, D 22, D 159, D 169, C 734.
Palagonitic: D 5, D 17, D 18, D 20, D 22, D 24, D 169, C 734, A 696.

MACQUARIE RIDGE

D 5 56° 40·6′ S, 158° 45·5′ E. 1,280 m

40% Mn nodules surrounding ashy volcanic agglomerate.

30% angular fragments of interbedded ash and agglomerate.

Agglomerate. Fragments are fresh plagioclase crystals, occasional discrete olivines, fine-grained weathered glassy basalts, olivine basalts, and other undifferentiated volcanics. Plagioclase is typically An_{60-80}. Foraminifera from the ash matrix of the agglomerate indicate submarine deposition. Results of paleontological investigation are given in Appendix III.

D 6 55° 29′ S, 158° 31·5′ E. 415 m

100% angular or subangular basic volcanic fragments.

Olivine Basalt. Porphyritic with plagioclase (An_{60-70}) forming dominant phenocrysts + subordinate amounts of augite and olivine. Hypersthene (5%) and magnetite (1%) are minor constituents. Pyroxenes are partly replaced by hornblende. Sample veined by clinozoisite.

Weathered Basalt. Plagioclase phenocrysts determined as An_{32} form dominant phenocrysts. Augite is identified in equal amounts with plagioclase in groundmass, but is rare as phenocryst. Hypersthene and magnetite are about 5% total constituents. Sample cut by clinozoisite veins.

Basic Agglomerate. Consists of basalt and glassy basalt fragments in a volcanic-ash matrix containing high proportions of devitrified volcanic glass. One plagioclase determination made on a basalt sample indicated a composition of An_{40-50} but this is not reliable since based on a single observation. Fragments of clinozoisite veins, together with slight granulation of coarser rock fragments, indicate some mild cataclasis.

D 7 55° 11·4′ S, 158° 43′ E. 241 m

100% angular olivine basalt fragments 1–20 cm across.

Olivine Basalt. Angularity of samples and occurrence of freshly broken surface indicates sample probably *in situ.* Plagioclase 34·4%; augite 27·6%; olivine 10·67%; glass 22·43%; amygdales 3·4%; (?) analcime 1·5%. Plagioclase (An_{50-60}) has been extensively altered. Olivine occurs as completely serpentinised or chloritised phenocrysts. Augite is found solely as small groundmass crystals. Glass, brown and devitrified, contains a high proportion of magnetite specks. (?) Analcime occurs rarely, either in intergranular spaces or occasionally as discrete anhedral crystals. Veining is observed. Chemical and normative analyses are given in Table 6.

E 236 54° 59·7′ S, 158° 36·4′ E. 175 m

80% subangular, weathered gabbro and dolerite fragments, 0·5–9·0 cm across.

Gabbro. Plagioclase 43%; augite 47%; hypersthene 10%. Fresh plagioclases (An_{60-70}) are cracked, with wavy extinction indicative of cataclasis. Titaniferous augite and hypersthene are slightly chloritised. Magnetite is an accessory. Subparallel lineation of feldspar and pyroxenes is probably a primary texture.

Dolerite. Plagioclase 45%; augite 25%; hornblende 21%; hypersthene 2%; magnetite 7%. Plagioclase is An_{40-60}. Pyroxenes are partly replaced by hornblende (not a primary mineral). Magnetite is either intergranular or dendritically intergrown with partly replaced pyroxenes. Partial granulation of major minerals indicates mild cataclasis.

E 234 54° 55·5′ S, 157° 47·5′ E. 220 m

65% fresh angular fragments of gabbro up to 16 cm across.

35% angular, weathered fragments of sheared fine-grained basic volcanics.

Gabbro. Plagioclase 37·5%; augite 40·6%; hypersthene 12·5%; magnetite 9·4%. Plagioclase (An_{50-60}), augite, and hypersthene are all fresh.

D 9 54° 52′ S, 158° 50′ E. 113 m

90% fairly fresh angular fragments of serpentinised peridotite, 0·5–17 cm across.

10% fairly fresh subangular gabbro fragments up to 8 cm across.

Peridotite. Olivine 71%; enstatite 25%; augite 4%. Olivine is nearly completely serpentinised. Enstatite is also partly replaced, probably by a serpentine mineral.

Gabbro. Plagioclase 48%; augite 23%; hornblende 10%; magnetite 19%. Plagioclase (An_{45-55}) is marginally granulated and larger crystals have bent twin lamellae. Titaniferous augite is marginally granulated and partly replaced by brown hornblende. Granulation indicates cataclasis which has caused shear planes to form.

D 8 54° 52′ S, 158° 39′ E. 141 m

60% fairly fresh, subangular fragments of partly sheared and veined gabbro, samples of which may have gneissic texture. One sample contains a black peridotite xenolith. All fragments are between 1·5 and 10 cm long.

30% fairly fresh, subangular altered basalt fragments, 1–8 cm long.

10% fairly weathered angular pyroxenite fragments, 1–7 cm long.

Gabbro. Plagioclase 60%; augite 25%; hypersthene 14·5%; magnetite 0·5%. Plagioclase (An_{50-60}) exhibits strain features. Pyroxenes are marginally chloritised. Chloritisation is best developed along shear planes. Veining slight.

Basalt, D 8B and D 8C. Plagioclase c. 50%; hornblende c. 50%; magnetite between 1% (D 8B) and 5% (D 8C). Plagioclase (An_{50-60}) may be altered (D 8B) or fresh (D 8C). Fibrous green hornblende pseudomorphs pyroxenes. Bending of hornblende aggregates and fracturing of samples indicate mild cataclasis.

Pyroxenite D 8D. Enstatite 50%; brown hornblende c. 50%. Sample extensively sheared, fractured, and veined. Individual enstatites are marginally granulated and replaced by brown hornblende. This mineral also appears to completely replace and pseudomorph a second mineral, possibly a clinopyroxene.

D 26 54° 40′ S, 158° 49′ E. 68 m

100% fresh rounded fragments of basalt and olivine basalt 2–6·5 cm long.

Basalt D 26A. Plagioclase 56·5%; augite 30%; magnetite 13·5%. Plagioclase extensively altered. Reddish colour of sample could be due to extensive iron staining or some form of impregnation with manganese minerals.

Basalt D 26B. As above. Contains extensively altered plagioclase, one phenocryst of which appears to be An_{32} (universal stage determination by A. Duncan). Calcite occurs in veins, and partly replaces plagioclase.

Basalt D 26C. Plagioclase (An_{54}) occurs as phenocrysts or as radiating sheaf-like aggregates in a devitrified glassy mesostasis containing specks of magnetite and pyroxenes.

Olivine Basalt D 26D. Plagioclase 57·8%; augite 20%; olivine 1·4%; magnetite 5·8%; glass 11·4%; amygdales 3·6%. Plagioclase phenocrysts are An_{40-50}, groundmass crystals are An_{50-60}, and phenocrysts are zoned accordingly. Olivine occurs as rare phenocrysts. Phenocrysts are set in devitrified glassy mesostasis containing numerous magnetite specks.

D 25 54° 40′ S, 158° 49′ E. 55 m

100% subangular and subrounded fragments of basalt, 2–3 cm across, + one cobble, 12 cm × 9 cm.

Basalt D 25A. Plagioclase (An_{50-60}) phenocrysts form aggregates. Plagioclase lathes in glassy mesostasis have variolitic texture. Basalt D 25B is holo-crystalline, consisting of plagioclase (An_{60-70}) and green hornblendes, which pseudomorph pyroxene minerals. Magnetite is an accessory.

A 696 54° 37·7′ S, 158° 57′ E. 433 m

85% rounded and subrounded basic volcanics, 0·5–6·5 cm across.

15% subangular and subrounded weathered amygdaloidal basalt fragments, 0·5–2·0 cm across. Differential weathering makes the white amygdales stand out about 1 cm from the surface of these samples.

Dolerite A 696B. Plagioclase (An_{60-70}) constitutes more than 50%. Augite and hypersthene, in equal amounts, are both partly replaced and pseudomorphed by fibrous green-brown hornblende. Magnetite is a prominent accessory.

Basalt A 696C. Augite and plagioclase microlites occur in radiating sheaf-like aggregates. Palagonite partly replaces devitrified volcanic glass. Amygdales contain zeolites in radial aggregates.

D 24 54° 29·6′ S, 158° 59·5′ E. 459 m

A few angular to subangular fragments of basalt, reaching a maximum diameter of 3·4 cm.

Basalt. Plagioclase and augite microlites form variolitic texture in a devitrified glass mesostasis, much of which is palagonitised. Clinozoisite veins cut the sample.

C 734 53° 55′ S, 158° 55′ E. 360 m

100% subangular and subrounded fine-grained basic volcanics, 2 mm–3 cm across.

Basalt. Plagioclase, augite, and hypersthene microlites are identified in glassy, palagonitised mesostasis containing up to 5% magnetite.

D 17 52° 31′ S, 160° 31′ E. 124 m

100% weathered, angular, sheared and veined olivine basalt fragments, 2 mm–10·5 cm across.

Olivine Basalt. Olivine occurs as phenocrysts, forming 2% of the sample. It is rare in the glassy mesostasis, which contains plagioclase, titaniferous augite, and magnetite. Slight palagonitisation of glass is evident. Some vesicles contain calcite and zeolites. Chemical and normative analyses are given in Table 6.

D 18 52° 31′ S, 160° 31′ E. 128 m

80% fairly weathered, angular, olivine basalt fragments, ranging from granule size to a block 11 × 5·5 cm across.

10% poorly indurated basic volcanic fragments in an ashy matrix.

Olivine Basalt. Plagioclase 44·7%; olivine 2·2%; hypersthene 9·2%; augite 21·8%; magnetite 3·3%; glass 17%; vesicles 0·8%. Plagioclase (An_{50-60}) is 90% of the phenocrysts. Olivine is rare both as phenocrysts and in the groundmass, which consists of plagioclase and augite microlites with a little hypersthene set in devitrified volcanic glass diversified by magnetite specks. Partial palagonitisation of glass is evident. Chemical and normative analyses are given in Table 6.

D 169 50° 48′ S, 163° 57·5′ E. 896 m

30% one large angular and generally shapeless fragment of Mn-encrusted, weathered basic volcanic rock, about 12 cm across. Its ropy surface texture indicates that it may have been part of the surface of a lava flow from which it has been broken by dredging. Several smaller fragments of this material were also collected.

40% flattened and subrounded pebbles of fine-grained basic volcanics, 0·5–7 cm in diameter. Flattening of these samples is controlled by poorly developed shear planes. All samples have a thin film-like coating of Mn oxides.

Olivine Basalt D 169B. Plagioclase 55%; olivine 20%; augite 10%; glass 10%; magnetite 5%. Plagioclase (An_{50-60}) occurs as phenocrysts and in the groundmass. Pale brown augite, replaced olivines, and magnetite occur in the groundmass with devitrified volcanic glass which is partly replaced by palagonite.

Olivine Basalt D 169E. Mineralogy similar to above sample but exhibiting flow lamination of phenocrysts and groundmass crystals. One phenocryst, probably hypersthene, is observed.

D 22 50° 38′ S, 163° 57′ E. 755 m

Very small sample consisting of thin, angular, freshly-broken flakes of fresh palagonitic olivine basalt containing 2% olivine and up to 10% magnetite.

D 20 49° 39·8′ S, 164° 02·2′ E. 126 m

75% fairly weathered subangular pebbles of basalt, 2–5·5 cm across.

25% serpentinised peridotite pebbles, 2–4 cm across.

Basalt D 20H. Poorly formed, radiating, sheaf-like aggregates of plagioclase, augite, and rare hypersthene microlites form variolitic texture. Hypersthene is partly replaced by magnetite aggregates. Palagonite occurs in intercrystal spaces.

Enstatite Peridotite D 20J. Olivine, almost completely serpentinised, constitutes about 50%. Yellow fibrous minerals (possibly also serpentine minerals) partly replace enstatite.

D 159 49° 01′ S, 164° 30′ E. 741 m

25% one poorly indurated subangular pebble (8·5 × 4·5 cm) consisting of polyzoan and foraminiferal limestone containing small fragments of basic volcanics, up to 2 mm across.

50% two large subrounded pebbles of fine-grained basic volcanics, about 8·5 × 4·5 cm, having a ropy texture indicating that they may have been part of the surface of a lava flow.

25% small angular and subangular fine-grained basic volcanic pebbles, up to 1·5 cm across, some of which are firmly cemented to fragments of the limestone.

Glassy Olivine Basalt. Altered olivine, about 2% of the sample, occurs as phenocrysts and in the groundmass, which is formed by plagioclase, augite, magnetite, and devitrified glass.

Results of paleontological investigation of the limestone are in Appendix III.

AUCKLAND ISLANDS SHELF

D 39A. 50° 58′ S, 165° 45′ E. 549 m

60% subangular, fairly weathered fragments of olivine basalt and dolerite, 1·5–11 cm across.

30% weathered subangular fragments of polyzoan and shell fragmental limestone, 3–21 cm across, containing sparse granules of basic volcanic rocks.

5% a single weathered subangular fragment of volcanic agglomerate, 5.6 cm across, containing small fragments of fine-grained basic volcanic rocks, up to 2·2 cm long.

Olivine Basalt D 39aB. Sparse olivines occur as fresh phenocrysts and in the groundmass, which is mainly colourless augite intergrown with plagioclase (An_{40}). Brown, devitrified volcanic glass occurs in intercrystal spaces, has been partly replaced by yellow palagonite, and contains concentrations of magnetite. Some vesicles are partly filled with calcite.

Olivine Dolerite D 39aC. Plagioclase 50%; olivine 10%; augite 17%; magnetite 15%; anygdales 8%. Euhedral olivine has been completely replaced by a pale greenish serpentine mineral, and exhibits a brown reaction rim which may reflect the development of iddingsite. Magnetite is concentrated around olivine crystals. Augite rarely occurs in crystals greater than 0·5 mm long, and is found mainly in the intergranular groundmass. Plagioclase occurs as well developed subparallel lathes.

D 170 50° 54·5′ S, 165° 42·5′ E. 335 m

100% fairly fresh, subangular fragments of black olivine basalt up to 3·5 cm long. One in 25 samples is vesicular.

Olivine Basalt. Serpentinised olivine, constituting 10% of the sample, forms most of the phenocrysts. Plagioclase is rare as a phenocryst, being mainly found in equal proportions to augite in the groundmass. Euhedral magnetite makes up about 10% of the total minerals. Amygdales appear to be filled with calcite and probably zeolites.

D 76 50° 53·7′ S, 165° 54′ E. 168 m

Small sample composed entirely of weathered, subangular fragments of olivine basalt, 1–4 cm.

Olivine Basalt. Olivine, completely altered to a pale orange-brown mineral which may be iddingsite, occurs commonly as phenocrysts and in the groundmass, and forms about 5% of the sample. The groundmass consists of small replaced olivines, unaltered colourless augite, plagioclase microlites, and well formed magnetite crystals which constitute about 10% of the total.

D 172 51° 00′ S, 166° 03′ E. 179 m

Very small sample of freshly broken fragments of olivine basalt, 1–3 cm long.

Olivine Basalt. Fresh euhedral olivine is the sole phenocryst mineral. Partly developed iron-rich alteration rims are present. A glassy mesostasis contains small olivine, plagioclase, and augite microlites and accessory magnetite. Devitrified volcanic glass has been partly replaced by yellow palagonite.

D 80 50° 31·55′ S, 165° 59′ E. 104 m

90% one pebble (2·1 × 2·8 cm) and one large, freshly broken, angular fragment (7 × 5·5 cm) of olivine basalt.

10% an exotic fragment, described in Appendix IIa.

Olivine Basalt. Plagioclase (An_{50-60}) forms more than 60% of the sample. It is intergrown with small olivine and augite microlites and magnetite in the groundmass. Slight granulation of minerals indicates mild cataclasis.

D 198 50° 24′ S, 166° 14′ E. 141 m

100% fresh subangular fragments of olivine basalt, ranging from 1·5 to 7 cm on average, although one larger boulder measures 15 × 11 × 6 cm.

Olivine Basalt D 198A. Olivine 20%; augite 36%; plagioclase 16%; magnetite 18%; glass 10%. Fresh euhedral olivine, as phenocrysts and in the groundmass, may show slight marginal alteration to serpentine. The groundmass consists of pale brown augite microlites and magnetite, which is ophitic toward plagioclase in some cases. Altered plagioclase microphenocrysts contain, marginally, numerous small augites and magnetite crystals. The plagioclase commonly appears to be an interserstitial mineral. Pale yellow palagonite partly replaces brown devitrified volcanic glass in interserstitial spaces.

D 103 50° 40′ S, 166° 19·5′ E. 71 m

100% fresh to partly weathered, subrounded, dark grey pebbles of basic volcanics, 2–12 cm.

Basalt. Plagioclase (An_{50-60}) is the dominant mineral, occurring as large euhedral phenocrysts, and within the groundmass. Phenocrysts are normally zoned with slightly sodic outer rims, and tend to occur in aggregates. Small augite phenocrysts are also zoned. The groundmass consists of subparallel plagioclase lathes and augite microlites with up to 6% of magnetite.

D 173 50° 53′ S, 166° 32′ E. 141 m

100% moderately hard, yellow-brown, angular and freshly broken fragments of polyzoan and shell limestone, up to 8·8 cm across.

CAMPBELL ISLAND SHELF

F 139 52° 33·5′ S, 169° 09′ E. 141 m

100% fresh, dark grey, rounded pebbles of olivine basalt up to 5 × 3·5 × 1·7 cm (Perseverance Harbour, Campbell Island).

Olivine Basalt F 139B. Plagioclase (An_{50}) occurs as rare phenocrysts, and in the groundmass. Olivine is tentatively identified from the rhombohedral and prismatic form of small altered ferromagnesian minerals. The brown replacing mineral may be iddingsite. Groundmass constituents are plagioclase laths, augite microlites (unaltered), magnetite, and devitrified volcanic glass.

Olivine Basalt F 139A. Zoned plagioclase (An_{30-40}) forms phenocrysts with calcic rims in accord with the composition of groundmass plagioclase (An_{50}). Serpentinised olivine (10%), magnetite (10%), plagioclase (60%), and minor amounts of small euhedral augite are present.

B 183 52° 34·4′ S, 168° 49′ E. 210 m

One large subrounded fragment (6 × 6 × 4·5 cm) and one very small fragment of fairly fresh olivine basalt.

Olivine Basalt. Plagioclase (An_{70}) is the commonest phenocryst, although olivine crystals replaced by brown iddingsite are common. Augite occurs rarely as a phenocryst. Groundmass minerals are mainly plagioclase and altered olivine with smaller amounts of unaltered augite and magnetite.

PUKAKI BANK

B 195 49° 14·5′ S, 171° 45′ E. (1) 82 m, (2) 157 m

100% angular to subrounded, fairly fresh, black pebbles of olivine basalt and undifferentiated basic volcanics, all less than 1 cm in diameter.

Olivine Basalt B 195 (1). The sample consists almost entirely of plagioclase lathes in a devitrified glass mesostasis. The glass contains numerous small magnetite crystals and probable clinopyroxenes. This sample includes a section through the contact between olivine basalt and an ashy, indurated, foraminiferal limestone containing altered volcanic fragments and discrete foraminifera. It is inferred that the basalt was extruded into the sea floor in a foraminiferal ooze environment.

Basic Volcanic Agglomerate B 195 (2). A glassy and palagonitic ash matrix surrounds discrete fragments of glassy olivine basalts in which the glass is still fresh. Subparallel plagioclase crystals are flow oriented. Rare olivines form the only other identifiable mineral apart from palagonite and calcite, which occur in amygdales in some samples. Individual fragments of rock are typically pumiceous with numerous vescicles and amygdales. They may reach 2 mm in length.

D 209 49° 19′ S, 171° 45·7′ E. 81 m

100% freshly broken, weathered fragments of basic volcanic agglomerate, 1–3 cm across.

Agglomerate. A matrix of palagonitic volcanic ash encloses glassy, pumiceous, vescicular, and amygdaloidal fragments of basic volcanic rocks. Plagioclase (An_{60}) is the only readily identifiable mineral and commonly has subparallel orientation, indicating flow texture. Individual rock fragments may reach 4 mm in length.

D 210 49° 21′ S, 171° 53′ E. 353 m

Small sample of subangular fragments of glassy vesicular basalt and olivine basalt with maximum diameter of 2 cm.

Basalt D 210B. Plagioclase (An_{60-70}), magnetite, and the palagonite replacing some of the devitrified volcanic glass in this sample are the only identifiable minerals. The sample contains an inclusion of a very fine-grained and apparently holocrystalline basic volcanic.

Olivine Basalt D 210A. Texturally and mineralogically similar to D 210B, this sample contains rare olivine crystals and is veined by palagonite.

F 106 49° 30′ S, 172° 00′ E. 371 m

Small angular fragments of basic volcanics and poorly indurated foraminiferal limestone.

The Slope Surrounding the Auckland Islands Shelf

F 81 49° 32′ S, 167° 01′ E. 401 m

Freshly broken, angular fragments of calcareous quartzose conglomerate, up to 6 cm across.

Conglomerate F 81A. This sample is a section through four rounded pebbles, each up to 3 cm across, set in a calcareous, fine-grained matrix.

(1) Three pebbles consist of angular, subangular, and rounded quartz and feldspar grains, 0·5–1·0 mm long, densely set in a fine-grained calcareous matrix. Two-thirds of the minerals are quartz grains, of which some show wavy extinction indicative of strain, some contain inclusion trails indicating a probable metamorphic provenance, and others are probably derived from a granitic provenance. The remaining third are almost entirely potash feldspars, some containing inclusions, others perthitic in character. Fragments of microgranular quartz-feldspar aggregates are also observed. A granitic and metamorphic provenance of quartzo-feldspathic rocks is indicated for these grains. The mineral assemblage lacks ferromagnesian minerals, indicating a high degree of sediment maturity.

(2) One pebble of "muddy sand" consists of angular and subangular quartz grains, about 0·2–0·4 mm, together with small weathered rock fragments, potash feldspar grains, and glauconite grains, all set in a red-brown matrix. This consists of very fine particles, including some euhedral grains which indicate that the matrix may be a volcanic ash. Matrix minerals are evidently iron rich and the whole sample is heavily iron stained. Glauconite grains are typically well rounded and may have formed from rounded faecal pellets. Some glauconite is observed partly replacing ferromagnesian minerals. Quartz forms the main mineral constituent, equal in proportion to altered rock fragments and ferromagnesian minerals. Potash feldspars are subsidiary minerals. Since the mineral assemblage of this pebble bears some resemblance to other pebbles in the sample they may have a common provenance. The nature of the matrix of the iron-stained pebble however is taken to indicate volcanicity contemporaneous with sediment deposition.

(3) The matrix of the conglomerate is mainly calcareous. It consists of abundant Foraminifera and fine-grained, inorganic detrital minerals set in a finely comminuted calcareous groundmass. Quartz grains and subsidiary potash feldspar grains, about 0·1–0·2 mm long, form the detrital-mineral assemblage. Occasional glauconite grains are also observed.

Conglomerate F 81B. The matrix is a sandy limestone containing more than 50% of rounded to subangular quartz grains with subsidiary potash feldspar. Cementing these grains is a very fine-grained calcareous matrix with numerous foraminifera. Rounded glauconite grains form about 1% of the sample and appear to chiefly replace ferromagnesian minerals although some is observed within the tests of Foraminifera.

The margins of quartzose calcareous sandstone pebbles are diffuse and contain rare grains of glauconite and discrete Foraminifera. It is inferred that these marginal characteristics were obtained by the rounded, unconsolidated pebbles of sandstone rolling through a depositional environment characterised by foraminiferal ooze and glauconite formation. The sandstones themselves were probably formed in fairly shallow water near the source of the sand grains. The presence of strained quartz grains and fragments of granular metamorphic rocks indicates a granitic and metamorphic provenance. The absence of complete organic remains in sandstone pebbles indicates an environment of deposition affected by active processes such as a wave or current activity typical of near-shore environments. The foraminiferal limestone matrix was probably deposited in deeper water further from the source of detrital grains. However, that the ooze was deposited fairly near to the source area of its detrital components is evident from the large content of detrital fragments.

The glauconitic pebble of "muddy sand" is of different lithology and has coherent outlines, indicating a reasonable degree of induration prior to deposition. Although a similar quartzofeldspathic provenance is required for this sample, its distinct volcanic component places it in a different category to its fellow pebbles. Results of paleontological investigations are in Appendix III.

THE CAMPBELL PLATEAU

D 101 49° 20′ S, 166° 30′ E. 686 m

75% soft, friable, angular fragments of yellow-brown, limonitised, glauconitic foraminiferal and polyzoan limestone reaching 1·4 cm (four pebbles altogether).

D 138 48° 32′ S, 168° 19·5′ E. 668 m

One weathered, rounded pebble of fine-grained porphyritic basic volcanic 2 cm across.

B 172 48° 18·5′ S, 168° 32′ E to 48° 22·9′ S, 168° 30′ E. 732 m

50% fresh subrounded pebbles of soft, friable foraminiferal and polyzoan limestone reaching 2·7 × 2·2 × 1·5 cm.

F 122 48° 06′ S, 179° 57′ W. 252 m

100% fresh, subangular fragments of well indurated foraminiferal limestone up to 4 cm across.

F 132 49° 59′ S, 177° 32′ E. 1,335 m

The sample consisted entirely of Mn nodules. A section through the nucleus of one of these nodules showed it to contain rock fragments—

(1) Small fragments of basic volcanic rocks, including extremely fine-grained glassy basalt (?), one sample of which has well developed flow texture. One fragment has a well developed rim of probable zeolite crystals. These fragments, assumed to be basic, are up to 1·3 mm long and of probable local volcanic origin, although they may have been derived from the Antipodes Islands.

(2) Small fragments of biogenic limestone.

(3) Clear, unaltered, angular quartz crystals.

(4) Small, yellow-brown grains could be palagonitised volcanic glass.

D 176 51° 06′ S, 167° 48·5′ E. 582 m

55% amygdaloidal and vescicular olivine basalt, 1·5–10·5 cm.

43% basic volcanic agglomerate pebbles, 3·5–12·5 cm long, containing fragments of dark, fine-grained volcanic rocks, up to 2 cm long.

This is a large sample, elements of which have strong lithological similarity. Individual fragments are angular and subangular.

Olivine Basalt D 176C. Olivine, altered completely to brown iddingsite, is the sole phenocryst mineral. The groundmass is composed of plagioclase, small altered olivines, fresh augite microlites, euhedral magnetites, and interstitial radial fibrous aggregates of zeolites. These occur in obvious vesicles with calcite, but also in intercrystal spaces within the groundmass. Since these zeolites form 2–5% of the sample this rock is fairly alkaline.

Agglomerate D 176A. The agglomerate is composed of small fragments of basic volcanic rocks, up to 1 cm long, in a crystalline calcite matrix, probably of biogenic origin. Individual fragments are: palagonitic "pumiceous" volcanic fragments containing numerous vesicles and zeolite-filled amygdales, occasional discrete fragmented olivine crystals, and rare fragments of proxene crystals.

Agglomerate D 176B. Closely similar to the above sample, this contains vescicular and perlitic palagonitic rock fragments and fresh olivine crystals.

NEW ZEALAND CONTINENTAL SHELF

B 583 48° 00′ S, 167° 20′ E. 143·5 m

One freshly broken fragment of polyzoan shelly limestone containing whole mollusc shells in a matrix of comminuted polyzoan and shell debris, 3–7·5 cm across.

B 586 48° 00′ S, 166° 26·5′ E. 300 m

One fresh, angular pebble of muscovite granite, 2 cm across.

B 587 48° 00·2′ S, 166° 39′ E. 155 m

One slightly weathered, angular pebble of granite, 4 cm long.

B 589 48° 44′ S, 166° 30′ E. 188 m

One bryozoan nodule, 4·5 × 2 cm. The surface of this sample is extensively bored and encrusted with bryozoa. Internally however, it consists of a hard, well indurated crystalline limestone containing sparse molluscan fragments. In view of the similarity between this and other limestones from the shelf this sample is regarded as typical of shelf deposits and may be more or less *in situ.*

D 100 48° 02′ S, 166° 36′ E. 157 m

95% fresh, angular fragments of biotite granite, 3–8 cm in diameter.

5% one weathered, subangular fragment of mica schist, 3·5 cm across.

Granite. Slight granulation of quartz is evident, indicating mild cataclasis. This sample is mineralogically similar to D 155.

D 132 48° 06′ S, 167° 36·5′ E. 130 m

Large, angular, freshly broken fragments of polyzoan limestone containing entire mollusc shells. Results of paleontological investigation are in Appendix III.

D 153 48° 15·5′ S, 166° 16′ E. 347–256 m

100% fresh, angular and freshly broken fragments of shelly limestone, about 9–11·5 cm across, with a maximum diameter of 18 cm. Complete mollusc shells are preserved in a shell fragmental matrix.

D 155 48° 02′ S, 166° 38′ E. 141 m

60% fresh, subangular fragments of mica schist, reaching 14 cm.

30% fresh, subangular fragments of gneissic biotite granite, reaching 22 × 12 × 8 cm.

Mica Schist. Fine-grained schist with subparallel biotite and muscovite, up to 0·2 mm, separating granular quartz crystals about 0·1 mm long.

Gneissic Granite. Coarse-grained with orthoclase, up to 6 mm. Granulation of minerals occurs randomly between large crystals. Biotite is subparallel to shear planes, giving a gneissic texture. Orthoclase is the major mineral; quartz and plagioclase are subsidary.

TRANSPORTED ROCKS

D 5 56° 40·6′ S, 158° 45·5′ E. 1,280 m

20% of total sample consists of the following lithologies:

(1) Biotite-hornblende granite; coarse grained; slight chloritisation of biotite. Fresh elongate fragment, 4·7 × 2·5 × 1·4 cm.

(2) Biotite-hornblende granite; coarse grained; extensively chloritised; highly fractured and fragmented; Mn coated. Angular elongate fragment, 3·5 × 2·7 × 2·2 cm.

(3) Hornblende granite; medium grained; no visible biotite; Mn coated. Angular elongate fragment, 3·3 × 1·9 × 1·2 cm.

(4) Granite; fine grained; Mn coated. Angular elongate fragment 1·6 × 1·0 × 0·5 cm.

(5) "Acid" gneiss; medium grained; quartzofeldspathic; Mn coated. Subangular elongate fragment, 3·8 × 2·3 × 1·7 cm. (+ one smaller pebble).

(6) Mica schist; fine grained; weathered; narrow quartzofeldspathic banding; Mn-coated angular and subangular fragments (two), largest 2·9 × 1·8 × 1·5 cm.

(7) Pink arkose; fine grained; Mn coated. Subrounded pebble.

(8) Conglomerate; medium grained; quartzofeldspathic. Mn-coated, flattened, discoidal, and subrounded pebble, 2·0 × 0·7 cm.

(9) Greywacke; fine grained; Mn coated. Subangular, smooth, faceted, and elongated pebble, 4·5 × 2·8 × 2·3 cm.

(10) Mudstone; three pebbles; slightly slaty appearance. Mn-coated, angular fragments, maximum size 2·7 × 2·4 × 1·3 cm.

(11) Mudstone; hard, grey, and well-indurated. Mn-coated, subangular, faceted pebble, 2·0 × 2·0 × 0·9 cm.

(12) Basic volcanic; holocrystalline, fine grained, mid-grey, vesicular. Mn-coated, angular fragment, 3·8 × 3·6 × 3·5 cm.

D 169 50° 47′ S, 163° 57·5′ E. 896 m

30% of this sample consists of angular to subangular, fairly weathered fragments of acid gneiss, 1·0–4·5 cm maximum diameter. Two small quartz pebbles and a number of subrounded pebbles of phyllite also occur in this sample. All have a dark reddish brown Mn coating.

D 169A Gneissic Granite. Most crystals show a certain amount of marginal granulation. Granulation seems to follow parallel zones which may be related to shear planes. Biotite is aligned along these planes. Feldspar is present as orthoclase and subsidiary albitic plagioclase. Orthoclase is perthitic and myrmekite is developed between quartz and plagioclase. Individual feldspar crystals reach 1·8 mm. Strained quartz has been subjected to the most granulation: it is about 10% of the total mineral constituents. Brown biotite accounts for 1·0%. Accessory apatite and magnetite are present. This rock has been subject to cataclastic deformation which has caused granulation and straining of individual minerals. A certain amount of recrystallisation is indicated by the lineation of biotites parallel to poorly developed shear planes.

D 169C Granodiorite Gneiss. A completely recrystallised rock with a fairly even granular texture in which most crystals are anhedral. There is some tendency toward parallelism of ferromagnesian minerals. Oligoclase constitutes more than two thirds of the total feldspars, the potash feldspar being orthoclase. Quartz is present in small amounts. The dominant ferromagnesian mineral is a pale green pleochroic pyroxene, which could be augite or aegerine-augite; a dark green to pale olive-green hornblende is well developed as the subsidiary ferromagnesian. Sphene constitutes 1–2% of total minerals; apatite about 1·0%; and ferromagnesian minerals about 40%.

D 169D Fine-Grained Phyllite. An extremely fine-grained rock (average grain size less than 0·05 mm) consisting almost entirely of parallel oriented unidentifiable feldspars and hornblendes. Small amounts of strained and granulated quartz occur as occasional microporphyroblasts. Since this rock is coarser than a slate and does not have well developed slaty cleavage or schistosity, it has been termed a phyllite.

D 39A 50° 58′ S, 165° 45′ E. 549 m

About 5% of the total sample, is a single, fresh, subangular fragment of biotite granite, 7 cm long.

Biotite Granite. Orthoclase, the dominant mineral, is perthitic, and small amounts of microcline are also developed. Oligoclase forms the plagioclase feldspar. Biotite and dark green hornblende occur in approximately equal amounts and form about 5% of total mineral constituents, green hornblende commonly forming a corona around the biotite. Accessories are magnetite, apatite, and sphene.

D 80 50° 31·55′ S, 165° 59′ E. 104 m

10% is one pebble of fine-grained pinkish "acid" rock. The pebble is rounded, slightly encrusted, and measures about 2·3 × 1·7 cm.

D 101 49° 20′ S, 166° 30′ E. 686 m

25% consists of a single, fresh, angular pebble of quartzofeldspathic gneiss, 1·3 cm in diameter.

D 160 49° 31·5′ S, 166° 15·5′ E. 722 m

75% is weathered, subrounded pebbles of green and black basic volcanics with a maximum diameter of 2·9 cm. Angular fragments of polyzoan limestone, up to 1·4 cm in diameter, form the rest of the sample.

D 176 51° 06′ S, 167° 48·5′ E. 582 m

2% is one fresh, angular pebble of quartzofeldspathic gneiss, 5·5 × 3·5 cm. This is composed chiefly of granulated and strained quartz crystals with subsidiary orthoclase and plagioclase and accessory biotite and hornblende.

B 172 48° 18·5′ S, 168° 32′ E to 48° 22·9′ S, 168° 30′ E. 732 m

30% is a subangular pebble and granules of basic volcanics. The single pebble is 4·0 × 2·8 × 1·7 cm.

20% is fairly weathered, subrounded granules and a single pebble of granite, 3·2 × 1·9 × 1·7 cm.

F 80 49° 00′ S, 167° 01′ E. 621 m

One flattened, fresh, subangular pebble of calcareous sandstone, 4·0 × 3·5 × 1·0 cm.

70% rounded and subangular quartz, orthoclase, and rare plagioclase grains. Quartz and orthoclase are in approximately equal amounts. An iron-stained calcareous matrix is made up from pelagic foraminifera and bryozoan fragments.

F 144 53° 29′ S, 170° 56′ E. 596 m

One large unweathered pebble of "acid" pumice.

F 127 49° 22′ S, 176° 16′ E. 1,280 m

90% consists of very large fresh and angular boulders.

(1) Acid gneiss: one large boulder, 35 × 18 × 12 cm (+ one smaller boulder).

(2) Granite: one small boulder, 10 × 10 × 10 cm.

(3) Metamorphosed sedimentary conglomerate: one large boulder, 36 × 28 × 15 cm.

(4) Limestone: one large, bored boulder of fine-grained semicrystalline limestone, 19 × 10 × 10 cm.

F 129 49° 24′ S, 177° 59′ E. 978 m

25% consists of angular to subrounded fragments of assorted lithologies.

(1) 10% coarse-grained hornblende granite containing subsidiary biotite; fresh fragments, 1·0–6·5 cm.

(2) 5% medium- to coarse-grained, weathered hornblende granite fragments, 1·0–3·9 cm.

(3) 5% banded biotite gneiss fragments, 0·7–4·8 cm.

(4) 5% basic, fine-grained volcanics, both vesicular and/or porphyritic, 0·7–3·0 cm.

APPENDIX III

PALEONTOLOGY

Foraminifera —N. de B. Hornibrook*
Mollusca —P. A. Maxwell*
Marine Plankton—A. R. Edwards*

D 5 56° 40·6′ S, 158° 45·5′ E. 1,280 m

Calcareous ash or siltstone.
Foraminifera: Preservation indicates fossil character of

- *Globorotalia* aff. *scitula*
- *Globigerina quinqueloba*
- "Virgulina" sp.
- *Stilostomella* aff. *fijiensis* (Cushman)

Other species showing variable preservation are

- *Globigerina pachyderma* (nearly all sinistral and Mn-coated)
- *Globorotalia truncatulinoides* (rare)
- *Globorotalia inflata*
- *Globigerina bulloides*

The definite fossil assemblage consists almost entirely of a *Globorotalia* closely resembling *G. scitula,* although it is not identical with the Recent form, which is most unusual. This suggests that the assemblage is Pliocene and represents a southerly extension of *G. scitula* below the present Subtropical Convergence. The almost entirely sinistrally coiled *G. pachyderma* is in accord with the latitude of the station, suggesting a sub-Recent age.

This sample was probably deposited far from land, since the elements commonly found in a normal shelf fauna are absent.

Nannoplankton:

- *Coccolithus pelagicus* Opoitian-Recent
- *Coccolithus* sp.
- *Cyclococcolithus leptorus* Opoitian-Recent
- *Discoaster* sp.
- *D. variabilis* Altonian-Castlecliffian

Absence of warm-water species suggests deposition south of the Subtropical Convergence. The approximate 1:1 ratio of *C. leptoporus* to *C. pelagicus* indicates proximity of the convergence.

D 39 50° 58′ S, 165° 45′ E. 549 m

Polyzoan limestone.

Fauna: broken and worn polyzoa, rare barnacle plates, and echinoid spines. *Globorotalia* cf. *hirsuta* indicates post-Miocene age.

D 132 48° 06′ S, 167° 36·5′ E. 130 m

Polyzoan limestone.

Foraminifera: Numerous fresh Lagenidae, Miliolidae, Polymorphinidae, and Discorbiidae indicate a Recent or Quaternary age. Preservation of *Notorotalia zelandica* indicates a probable upper Wanganui Series age. This sample probably accumulated in water shoaler than 130 m.
Mollusca:

- *Plurigens phenax* Finlay
- *Micrelenchus* cf. *caelatus* (Hutton)
- *Estea rufoapicata* (Suter)
- *Murexsul* aff. *espinosus* Finlay
- *Zenepos totolirata* (Suter)

Plurigens phenax is most conspicuous, both as juvenile and adult shells. The type is from Otago Heads in 60 fathoms, and the shells also occur on Auckland Islands beaches.

Lower Pleistocene (Nukumaruan); possibly Upper Pliocene.

D 134 48° 16′ S. 168° 43·5′ E. 668 m

Phosphatised foraminiferal ooze.

Foraminifera:

- *Globigerina* spp.
- *G.* sp. (thick walls—could be *G. woodi*)

No definite *Globorotalia* visible.
Probably Oligocene - Lower Miocene (Landon-Pareora Series).
Open-water planktonic fauna.

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*New Zealand Geological Survey.

D 147 49° 31′ S, 167° 25′ E. 574 m

Phosphorite nodule.

Foraminifera:

Chiloguembelina cf. *cubensis*
Globigerina; minute, thin walled, 4–5 chambered.

A puzzling assemblage from which larger planktonic species might be expected (identified from thin section).

Upper Eocene-Oligocene (Arnold Series, Kaitaian stage—Landon Series, Whaingaroan stage).

D 153 48° 15·5′ S, 166° 16′ E. 347–256 m

Polyzoan limestone.

Mollusca:

Tawera sp.
Notirus sp.
Dentalium sp.

Wanganui Series, possibly Nukumaruan.

D 159 49° 01′ S, 164° 30′ E. 741 m

Polyzoan and foraminiferal limestone.

Foraminifera:

Globorotalia inflata
Globorotalia aff. *miozea sphericomiozea* (Walters)
Globigerina aff. *woodi* Jenkins
Globorotalia hirsuta
Globorotalia aff. *punctulata*
Globigerina pachyderma
Ehrenbergina hystrix
Planulina wuellerstorfi

The inflata-spherico-miozea complex is unlike any Recent populations and is typical of the Opoitian. This involves an apparent discrepancy in that Opoitian benthic species are not present. This may reflect a completely open-sea fauna. Probably lower Opoitian.

D 173 50° 53′ S, 166° 32′ E. 141 m

Polyzoan limestone.

Foraminifera:

Textularia (rare)
Quinqueloculina

Indeterminate age.

F 81 49° 32′ S, 167° 01′ E. 401 m

Calcareous quartzose conglomerate.

Foraminifera:

Globorotalia cf. *miozea*
Globigerina woodi or *Globigerinoides trilobus*

Southland Series (mid-Miocene).

F 122 48° 06′ S, 179° 57′ W. 252 m

Foraminiferal limestone.

Foraminifera:

Globorotalia sp. cf. *miozea* (thin walls)
Globigerina ? sp. (thick walls)
Ehrenbergina hystrix

Uncertain age, Miocene-Pliocene.

Foraminifera from six core samples were examined by Dr N. de B. Hornibrook. The six core samples, although all containing good microfaunas, are impossible to date accurately from available information. It can be said that all are post-Miocene but only general ages can be given. Until a sequence from top to bottom of each core is established there is too little information to allow accurate dating. Each core sample analysed was taken from the base of a 6–10-ft core.

D 101 *Globorotalia* cf. *sphericomiozea*
Globorotalia inflata
Globorotalia hirsuta
Globigerina bulloides
Lower Wanganui Series (?)

D 103 *Globorotalia inflata*
Globorotalia hirsuta
Globigerina bulloides
Upper Wanganui Series or younger

F 106 (X) *Haeuslerella* ? sp.
Euuvigerina cf. *rodleyi*
Bulimina echinata
Bulimina aff. *senta*
Ehrenbergina hystrix
Cibicides deliquatus
Globorotalia aff. *sphericomiozea*
Globorotalia hirsuta
Globorotalia pachyderma
Pliocene or Pleistocene

B 172 *Notorotalia* aff. *zelandica*
Globorotalia inflata
Globorotalia crassaformis
Globorotalia cf. *sphericomiozea*
Euuvigerina peregrina
Pliocene or Pleistocene

B 108 (X) *Globorotalia hirsuta*
Globorotalia inflata
Globigerina bulloides
Probably Pleistocene

F 121 (XI) *Globorotalia inflata*
Globorotalia truncatulinoides
Globigerina bulloides
Globigerina quinqueloba
Euuvigerina peregrina
Probably Pleistocene

F 127 49° 22′ S, 176° 16′ E. 1,280 m

Phosphorite nodule.

Foraminifera:

Globorotalia sp.
Globorotalia cf. *inflata*

Many planktonic species (not identified) and abundant Radiolaria. Upper Miocene to Pleistocene; deep water. Section through nodule given in Fig. 11.

APPENDIX IV

SUBMARINE MORPHOLOGY

INTRODUCTION

Brief descriptions of major morphological features, given below, are based on interpretation of the 1:1,000,000 charts and individual echo-sounding traverses. Reference to individual traverses is indicated by the appropriate traverse number in the text.

The continental shelf and shelves surrounding individual islands are approximately bounded by the 250-m isobath. True shelf dimensions cannot be given at this stage owing to the paucity of echo-sounding data. Shelf zones are separated from the regional levels of the Campbell Plateau by restricted zones of high gradient forming, in effect, poorly developed continental slopes. Local elevations or depressions occasionally interrupt their smooth surfaces.

SELECTED ECHO-SOUNDING TRAVERSES
(in pocket)

1a–f. Campbell Plateau
2a–h. Antipodes Islands, Bounty Platform, and Pukaki Saddle
3a–e. Shelf Regions, Cathedral Banks, and Snares Depression
4a–d. Aucklands Slope
5a–b. Emerald Basin and Aucklands Slope
6a–d. Macquarie Ridge, Solander Trough, and Aucklands Slope
7. Rough topography and sub-bottom reflections; Pukaki Rise
8a–f. Subantarctic Slope
9a–d. Continental Rise, Abyssal Plain, Abyssal Hill Province, and seamounts; Southwestern Pacific Basin

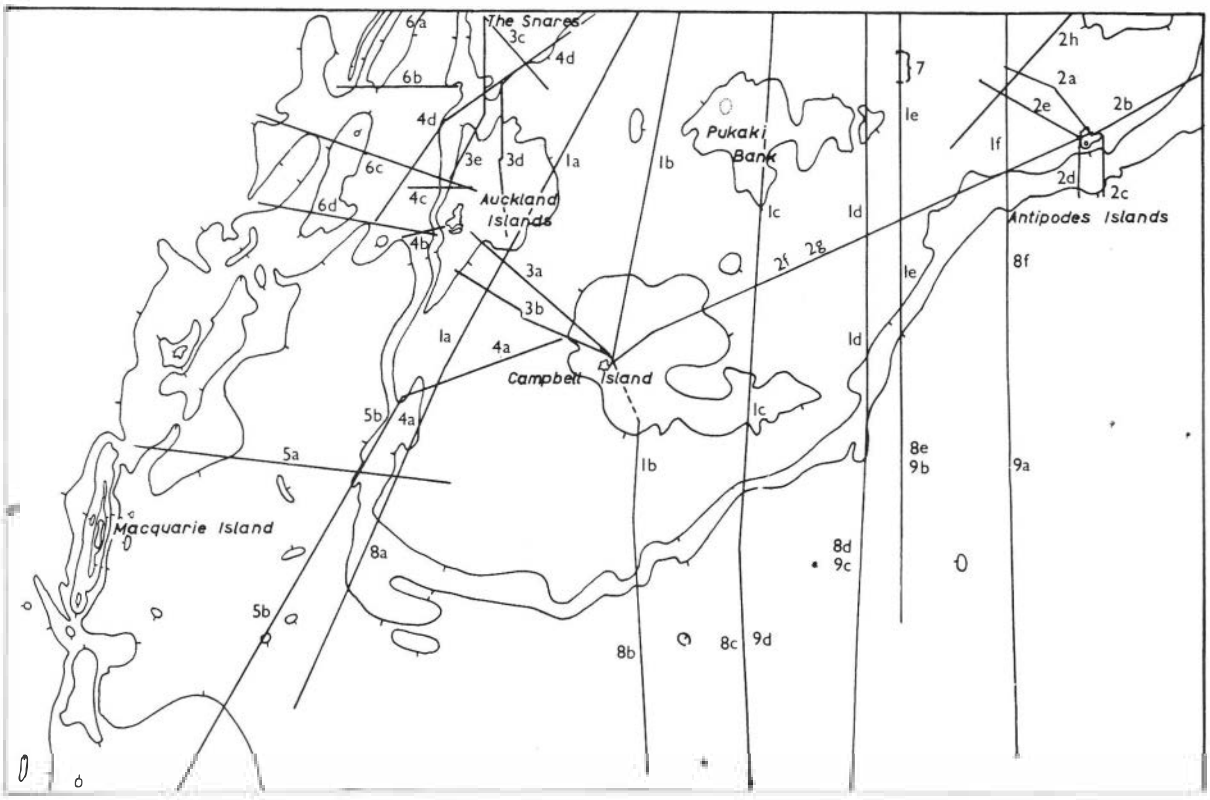

Fig. 31: Distribution of selected echo-sounding traverses.

CONTINENTAL SHELF OF NEW ZEALAND

The New Zealand continental slope forms the north-western margin of the Campbell Plateau. Above it the continental shelf (Traverses 3c and 3e) lies in about 150 m. Shelf-edge depths are commonly 150 m, locally reaching 190 m (Traverse 6a) in the west.

Upper parts of the continental slope reach 14° (Traverse 3c). Submarine canyons do not occur along the continental slope (Traverse 4c), but indentations in the shelf edge to the south-west are probably submarine canyons which lead south across the Aucklands Slope. Within the broad embayment associated with these shelf-edge indentations is a narrow 40-m-deep gully (Traverse 3e) which may be a canyon.

CAMPBELL PLATEAU*

Island Shelves and the Pukaki Bank

Four restricted zones of elevation on the Campbell Plateau rise above sea level to form the Subantarctic Islands of New Zealand. A fifth zone forms the Pukaki Bank, a flat-topped feature in depths typical of Pleistocene shelves and regarded as similar to the island shelves in this region.

Auckland Islands Shelf

The Auckland Islands Shelf covers an area of about 3,100 sq. miles. Depths greater than 90 m are typical of this smooth shelf; the western shelf lying in 180–210 m, the eastern in 90–105 m (Traverses 1a, 3d, 3e, 4c, and 6c). Depths of less than 90 m are found within 20 miles north-east and north of the islands. Shelf-edge depths range from 142 to 212 m.

A sinuous, shallow depression, the Enderby Channel (Fig. 2), north-east of the islands, connects with small indentations in the shelf edge. Similar shelf-edge indentations east of Adams Island are associated with the Carnley Sea Valley (Fig. 2) which extends from the Auckland Islands region south across the Campbell Plateau (compare Summerhayes, 1967). These channels and sea valleys may, on further investigation, prove to be parts of an extensive submarine canyon system.

The steep upper part of the poorly developed continental slope around the islands (Traverses 1a, 3d and 3e) is typically 4°–12°.

Campbell Island Shelf

The Campbell Island shelf covers an area of some 1,700 sq. miles. Much of the shelf lies in depths of 155–180 m, mainly greater than 90 m. Rough topography, commonly with vertical relief of about 145 m, is found up to 4 miles off shore. A localised but exceptional area of shoaling occurs 27 miles south-east of the island, where depths of 140 m are recorded. Shelf-edge depths (Traverse 1b, 2g, 3a, and 3b) lie between 236 and 183 m.

The continental slope around Campbell Island is commonly 0·2°–1·5° (Traverses 1b and 2g), steepening north-west, south, and west of the islands (Traverses 3a and b). Submarine canyons cutting this western slope (Traverse 3b) may extend west to join the Carnley Sea Valley (compare Fig. 6).

Bounty Islands Shelf

The Bounty Islands shelf, which forms an integral part of the Campbell Plateau, lies just to the north of the charted area (Fig. 14). The shelf is flat in 140–180 m. Changes taken to mark the shelf edge are remarkably slight (compare Traverse 2h) except to the north, where there is a very steep drop into the Bounty Trough (Krause, 1966). The shelf edge lies in 163–205 m.

Antipodes Islands Region

The Antipodes Islands are the subaerial expression of a localised region of volcanic activity. There is no apparent shelf development, and isobaths around the island descend steeply to the Campbell Plateau in 900–1,100 m (Traverses 2a–f).

A 12-mile-wide zone of submarine pinnacles surrounds the islands, covering about 200 sq. miles above 900 m. All individual peaks cannot be shown owing to the small-scale (1:1,000,000) chart used. Pinnacles are typically well formed, sharply defined features with little surface relief, and flanking slopes of 10°–30°, averaging 17° and occasionally reaching 60°. Relief of individual peaks is 200–260 m.

Pukaki Bank

The Pukaki Bank, situated at 49° 40′ S, 171° 59′ E, on the northern edge of the Campbell Plateau, was first recorded by RRS *Discovery II* in 1936. Since then it has been surveyed in detail on N.Z.O.I. cruises. It is an extensive feature, similar in overall characters to the island shelves, though lacking an emergent portion (Brodie, 1964). Covering some 250 sq. miles in 180 m, it forms a flat-topped eminence in 130–140 m. "Numerous steep-sided pinnacles rise from this surface to a recorded minimum of 60 m" (Brodie, 1964). The edges of the bank are defined, as for the continental shelf, by a pronounced steepening of slope which occurs in 150–200 m.

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*See footnote p. 83.

Narrow depressions, which may be small submarine canyons, indent the upper slopes around the bank but have little morphological expression below 250 m, except to the north where two depressions continue to 750 m.

PLATEAU SURFACE

The plateau surface, covering most of the north-eastern region, is a normally flat, smooth area, lying mainly in depths of 450–1,000 m, from which rise the Subantarctic Islands, the Pukaki Bank, and other minor elevations. It is bounded by a restricted zone of markedly high gradients forming the plateau marginal slope. At the base of the plateau marginal slope is a continental rise. The marginal zone starts in depths between 1,400 and 2,000 m. Below 1,500 m, the Campbell Plateau is a unit, but in shoaler water it is divisible into two major units, an eastern and a western. The western plateau surface is separated from the eastern surface—the Bounty Platform—by the Pukaki Saddle, a broad shallow depression in about 1,350 m.

BOUNTY PLATFORM

This eastern part of the plateau, from which rise the Bounty and Antipodes Islands, is a plateau-like feature in 750–950 m. A broad and gentle rise towards the Bounty Islands dominates its northern and eastern parts (Fig. 2h). The Antipodes Islands, rising abruptly from an almost flat sea floor at the southern edge of the platform (Traverses 2a–d and 2f) have no marked effect on the regional bathymetry. Westward the platform deepens in two broad horizontal "steps", each some 10 miles wide in 1,150 and 1,210 m, before the commencement of a 2° downward slope defines the edge of the Pukaki Saddle (Traverses 2f and h).

Fifty miles north-west of the Antipodes a zone of rough topography contains groups of small, sharply bounded, flat-topped elevations with maximum vertical relief of 150 m, and a minimum depth of 850 m (Traverses 2a and e). These features are flanked by 5°–40° slopes. Around them, the general level of the platform remains constant in about 940 m.

The northern boundary of the platform occurs about 30 miles north of the Bounty Islands where steep slopes plunge into the Bounty Trough from the edge of the Bounty Islands shelf (Krause, 1966). The southern-eastern platform edge lies in approximately 1,400–1,500 m (Traverses 2b–d) and is locally interrupted by the Antipodes massif.

A transitional zone with 0·5°–1·0° slopes occurs between the platform level in 950 m and the marginal slope in 1,500 m.

PUKAKI SADDLE

The Pukaki Saddle is a broad, northerly trending, shallow, V-shaped depression in about 1,350 m, with local maxima between 1,450 and 1,620 m, separating the eastern and western parts of the plateau surface (Traverses 2f and h). In 1,250 m it is about 45 miles wide but is constricted at its southern end to a width of 15 miles. Two sinuous depressions to 1,500–1,600 m are formed within the constricted part of the saddle. These continue south, with reduced expression, on to the marginal slope. Elsewhere in the saddle, sea-floor channels are not found.

Boundaries of the saddle are west, the edge of the flat plateau surface in 750 m, and east, the edge of the Bounty Platform in 1,250 m.

WESTERN PLATEAU SURFACE

The western plateau surface is by far the larger of the two plateau segments. North and east of Campbell Island and east of the Auckland Islands the greater part of the plateau surface lies in 500–750 m, and a considerable part in less than 600 m. Hence the plateau is not uniformly flat, its usually featureless surface being relieved by broad rises and depressions flanked by slopes of less than 0·5°.

PUKAKI RISE

Pukaki Rise is a broad feature, elongated east-west, of which the Pukaki Bank is the shoalest part. The crest of the rise lies in about 450 m, locally shoaling to 300 m. In 500 m it is 130 miles long, and locally reaches a width of 100 miles. Eastward it constricts the southern part of the Pukaki Saddle and westward extends at least 80 miles from the Pukaki Bank (Traverse 1b). Like most of the plateau the rise has a very smooth surface. It has a broad, flat top and is flanked by 0·5° slopes (Traverses 1b–e), interrupted along the north-eastern flank by a zone of small blocky elevations and depressions (Traverse 7a).

North of the rise 0·5° slopes lead down toward the Bounty Trough (Traverses 1b–e), and an arbitrary northern limit to the Campbell Plateau is set here in 1,250 m. Several small, poorly developed depressions, probably sea valleys connecting with the Campbell Channel (compare Fig. 14, based on Krause, 1966), cut this northern edge.

Campbell Island Rise

The plateau surface in 550 m between the Pukaki Rise and the Campbell Island Rise is interrupted sporadically by elevations to about 450 m, the shoalest recorded reaching 390 m.

The Campbell Island shelf is sited on a shallow dome-like feature forming the western end of the Campbell Island Rise. This rise, flanked by 1°–2° slopes, is an elongate feature extending some 155 miles east and 60 miles north of the island in 500 m. Eastward, the rise is some 30 miles wide in 500 m, reaching a maximum width of 60 miles. A perceptible crest extends for 50 miles along the rise in 330–400 m, between 171° 20′ E and 172° 30′ E at 53° 00′ S (Traverses 1c and d).

North-east of the rise, above the Subantarctic Slope, the plateau edge occurs in 1,350–1,380 m (Traverses 1d and 2). Separating the plateau edge from the flat 550 m plateau surface is a transitional zone, characterised by 1·5° slopes decreasing north to 0·5°. Minor blocky elevations occur along the western edge of this transitional zone (Traverse 1e).

Snares Depression: (Traverses 3c–e)

Between the continental shelf of New Zealand and the upstanding Pukaki Rise and Auckland Islands Shelf is a broad, elongate, flat-floored depression, the Snares Depression. It is about 80 miles wide in 640 m, narrowing to 12 miles in 500 m north of the Auckland Islands, where it is crossed by a minor ridge. West of this ridge the Snares Depression continues in about 650–750 m to the Auckland Slope. A few miles east of the ridge is a completely enclosed basin in about 730 m with a maximum depth of 766 m. Between the Pukaki Rise and the continental slope a broad, sinuous, well defined sea valley crosses the depression in about 730 m and deepens gently north-east toward the Bounty Trough. A median channel within the northern part of the sea valley reaches depths of 1,000 m or more locally but is not everywhere definable as a continuous feature owing to the poor coverage of the area by echo-sounding.

Auckland Islands Ridge

The western margin of the plateau is dominated by the Auckland Islands Ridge and associated shelf. As at Campbell Island, the Auckland Islands Shelf is situated on a dome-like elevation continued for some 120 miles south of the islands as a narrow submarine ridge. This, the Auckland Islands Ridge, is an extremely sharp, well defined feature running parallel to the Aucklands Slope as far as 53° S.

Southern Part of the Campbell Plateau

South of the Campbell and Auckland Islands the Campbell Plateau, although retaining the regional characteristics of smoothness, moderate depth, and a lack of distinctive topographic relief, differs in having a regional south-westward dip, and in development of marginal relief (Traverses 1a, 4a and 5a). Interpretation of the complex topography of this area is difficult since there is extremely poor echo-sounding cover.

Centred 45 miles south-east of the Auckland Islands is a localised zone of isolated elevations often with vertical relief of more than 250 m. This zone of banks—the Cathedral Banks—covers an area 50 miles long (Traverse 1a) by 10 miles wide (Traverses 3a and b), and has a notable north-north-east trend. Individual elevations appear as peaks flanked by slopes of 10° or more, sometimes reaching 50° (Traverse 1a). They rise abruptly from the flat sea floor to reach minimal depths of 350–460 m. Extending through this zone is a narrow, deep channel, difficult to chart along its length (Traverses 1a, 3a and b), which is probably an extension of the Carnley Sea Valley (Fig. 6).

South-west of Campbell Island much of the plateau lies in 850–950 m, but towards the plateau edge the depth increases gradually to 1,250–2,500 m. Along its western margin the plateau edge is diversified by several broad elevations and depressions (Traverse 5a). Between the Auckland Islands Ridge and the plateau surface is a deep, steep-walled, smoothly floored and relatively trough-like depression, the Cathedral Depression, extending north from 53° 10′ S to within 30 miles of the Cathedral Banks. The depression is 8 miles wide in 2,500 m (Traverse 4a), and is bounded west by a 9° slope flanking the ridge and east by a 26° slope up to the plateau surface. Southward the eastern wall of the depression merges with the Inner Aucklands Slope and the floor of the depression merges with a flat mid-slope bench in about 2,300 m.

Two major elevations rise from the western plateau edge. Northernmost is a broad eminence, flat-topped in about 850 m (Traverses 4a and 8a) and flanked by 1° slopes, except westwards where it abuts against the steep wall of the Cathedral Depression (Traverse 4a). Near the southern tip of the plateau a second, smaller eminence reaching a recorded minimum depth of 1,202 m, is separated from its northern counterpart by a broad depression in 1,500–1,600 m. The southern margin of the plateau is highly irregular and a plateau edge is not easily defined.

Due south of Campbell Island other broad gentle elevations diversify the plateau (Traverse 1b) before the Subantarctic Slope commences in about 2,035 m. Further east, the plateau edge occurs in about 1,400 m (Traverse 1c).

MARGINS OF THE CAMPBELL PLATEAU*

The plateau marginal slope is a narrow zone of high gradient separating the Campbell Plateau from the deep-ocean floor. The western part is known as the Aucklands Slope and the eastern as the Subantarctic Slope. Northward, a broad zone of very low gradient, which is not a true plateau marginal slope, separates the Campbell Plateau from the Bounty Trough. Since it is impossible to define precisely a northern edge of the plateau, a boundary is arbitrarily selected in 1,250 m.

Aucklands Slope

The steep slope forming the plateau's western margin is continuous with the continental slope of New Zealand. Together they form a distinct linear feature stretching from Stewart Island (47° S) to 55° S. This feature, the Aucklands Slope, is slightly sinuous but maintains an approximate north–south trend over its 480 miles. It has a regular surface, broken in intermediate depths by a variety of elevations and depressions. Locally, the slope is interrupted by fairly broad indentations but, owing to the lack of echo-sounding traverses parallel to the slope, it is difficult to determine their exact nature.

As pointed out by Krause (1966) the typical continental slope, as a single feature, is rarely found. Such slopes are commonly divisible into inner and outer slopes separated by broad zones, platforms, or benches in intermediate depths. Krause defines the inner slope as that commencing at the edge of the continental shelf and the outer slope as that terminating on the deep sea floor. Slopes bordering the Campbell Plateau are in general closely similar to continental slopes elsewhere and are similarly benched in intermediate depths. Krause's terminology is here amended to define the inner slope as that commencing at the edge of the Campbell Plateau.

Echo-sounding traverses across the Aucklands Slope adequately illustrate certain aspects of the slope morphology (Traverses 4a–d, 5a, b, and 6a–d). Inner slopes are typically 4°–12° compared with outer slopes which are much steeper (10°–25°). Benches are not ubiquitous and such features are lacking due west of the Auckland Islands (Traverse 6d), (compare Summerhayes, 1967).

Bench surfaces are commonly undulating and dip slightly toward the plateau. The junction between bench and inner slope is usually sharp. An extensive bench in 2,250–3,400 m is developed along the south-western margin of the plateau (Traverse 5a). At the base of the slope, the junction between the slope and the floors of the Solander Trough and Emerald Basin may be smooth or abrupt (Traverses 6a–d).

Off the south-western part of the Campbell Plateau a linear zone of steep-sided elevations (Traverses 5a and b) parallels the slope between 53° S and 53° 30′ S.

Subantarctic Slope

This feature, with a pronounced north-east trend, stretches from the eastern end of the Chatham Rise to the southern end of the Aucklands Slope forming the eastern and southern margins of the Campbell Plateau.

At its southernmost point the slope is highly complex and a zone of ridges 100 miles wide separates the Campbell Plateau from the South-western Pacific Basin. An abrupt break in slope in 1,300 m marks the plateau edge.

Typically, the slope may be divided into an inner slope and an outer slope separated by benches in intermediate depths. Slopes are commonly 10°–15°. Typical echo-sounding traverses across the slope illustrate certain aspects of the morphology (Traverses 8b–f).

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*Note Added in Proof:

Seismic reflection profiles obtained by Lamont Geological Observatory during cruises in New Zealand waters by *R. V. Vema*, *R. V. Conrad*, and *R. V. Eltanin* have been published since completion of this bulletin (Houtz, Ewing, Ewing, and Lonardi, 1967). Profiles across Campbell Plateau show that:

(1) basement topography is fairly regular and follows surface topography quite closely although with slight exaggeration;

(2) horizontally bedded sediments reach maximum thicknesses of about 1 km between rise crests;

(3) thinning of sediments on rise flanks is in part due to lateral transgression but is also due to the relatively greater accumulation of sediment in the "basins" between rises;

(4) along the western margin of the plateau is a zone of localised tectonic distortion not found elsewhere on the plateau;

(5) the pinnacles comprising Cathedral Banks are composed of rocks which extend at least to basement.

These data support the conclusions put forward in this bulletin for:

(1) the originally erosional character of the regular plateau surface;

(2) the later development of rises and depressions;

(3) the probably volcanic origin of Cathedral Banks;

(4) the tectonic character of the western plateau margin.

At approximately 174° E the slope is broken by isolated elevations forming the Endeavour Banks (Traverse 8d). The largest of the three banks has a gently sloping top in about 2,470 m, is distinctly elongated north–south, and has steep flanks. A narrow depression which may form part of a submarine canyon system occurs in the depression between the largest bank and the slope (Traverse 8d).

A pronounced embayment indents the slope south of the Pukaki Saddle. South of the Antipodes Islands the slope is interrupted by a north-west-trending ridge-like feature with steep southern flanks (Traverse 2c).

CONTINENTAL RISE

A broad zone of markedly gentle gradients forms a continental rise obscuring the junction between the steep Subantarctic Slope and the level Southwestern Pacific Basin floor. The rise generally commences in 4,500 m ± 400 m and slopes at 0·5° — 0·1° toward the ocean basin (Traverses 9a–d). Boundaries of the rise extrapolated from available profiles, are shown in Fig. 6. It is 10–100 miles wide—widest east of the Antipodes Islands, and narrowest a few miles north-east of the Endeavour Banks. Occasional local elevations and depressions up to 200 m in vertical relief occur on the rise. A continental rise does not occur at the foot of the southernmost part of the Subantarctic Slope west of 169° E. Locally, at the foot of the Subantarctic Slope, narrow depressions are found (Traverses 8a, f, and 9a).

The junction between the Aucklands Slope and the floor of the Solander Trough is partly obscured by a feature resembling a continental rise (Traverse 6a) which is present at intervals along the slope. This is described under the heading Solander Trough. South of the Auckland Islands a small continental rise is developed between the Aucklands Slope and the Emerald Basin flanking the Auckland Islands Ridge. Further south there is no continental rise (Traverses 5a and b).

Where it flanks the Solander Trough the Macquarie Ridge is mantled by a continental rise which is an integral part of the trough floor (Traverse 6a). A rise is not developed on other parts of the Macquarie Ridge.

MACQUARIE RIDGE

The ridge is rugged, with peaks reaching locally 200 m or less. It has, in places, a double crest of two narrow parallel ridges. Individual peaks on the ridge crest are commonly split (Traverses 6a–d) by deep, steep-walled, V-shaped depressions, which may or may not be continuous along the ridge. The best development of a double crest occurs in the "block" on which Macquarie Island is sited. The island forms the emergent part of the eastern crest of the ridge. The western crest forms several flat-topped elongate banks in about 160 m off the west and south coasts of the island, from which they are separated by deeps of about 800 m. Peaks on the eastern side of the ridge are commonly the highest, and broad benches are developed on the western flanks, making the ridge distinctly asymmetric. Owing to the paucity of echo-soundings in this region it is not possible to say whether the benches (particularly on the eastern flanks of the ridge) are continuous.

Westward the flanks of the ridge plunge steeply into the Tasman Basin; eastward they form the walls of the Solander Trough and Emerald Basin. The flanks commonly have 10°–15° slopes, reaching 40°–50° locally (Traverses 6a–d).

Over the distance covered by the charts used, the ridge is continuous in depths below 3,750 m. In shoaler water it is a series of isolated, elongate blocks separated by narrow east–west-trending deeps.

Although not continuous features, narrow trenches with depths of 5,000–6,000 m are found on both sides of the ridge. Eastward these trenches separate the main Macquarie Ridge from a sub-subsidiary ridge which extends from 51° S to the latitudes of Macquarie Island (Traverse 5a).

The main Macquarie Ridge is offset some 50 miles to the west at about 51° S. However, the line of the northern part of the main ridge is continued southward by a subsidary and much reduced ridge in about 3,500 m. This feature forms an integral part of the Macquarie Ridge complex. It locally reaches depths of 2,780 m, varies in width from 6 miles (north) to 35 miles (south) and is flanked by 1–7° slopes. The trench separating the subsidiary from the main Macquarie Ridge reaches depths of 5,700 m. Nowhere are the crests of both parts of the Macquarie Ridge more than 35–40 miles apart (compare Fig. 12). Like the main ridge, the subsidiary ridge is rugged (Traverse 5a).

The flanks of both main and subsidiary ridges are diversified by benches or well developed elevations and depressions in intermediate depths. Individual elevations are very steep sided with slopes commonly exceeding 10°. Because of the magnitude and steepness of slope of these elevations on the flanks of the main ridge (Traverse 7) it is impossible to make precise representation of the bathymetry on the 1:1,000,000 scale used for the charts.

Narrow, steep indentations in the eastern flank of the Macquarie Ridge may be submarine canyons. Two canyons near the latitudes of the Auckland Islands appear to be directly continuous with a channel on the floor of the Solander Trough. Other probable canyons occur in the eastern flank of the ridge off Macquarie Island.

The Macquarie Ridge is slightly arcuate, changing in trend from approximately north-south in the southern part of the area, to approximately north-east in the north.

SOLANDER TROUGH

The Solander Trough separates the Macquarie Ridge from the Campbell Plateau. The northern part of the trough from the New Zealand mainland to the Snares (Fig. 9) has been described by Brodie (*in* Harrington and Wood, 1958). More bathymetric detail has been revealed by subsequent surveys, and Brodie and Dawson (1965) described the trough as descending to the Emerald Basin on the eastern side of the Macquarie Ridge.

The trough is a large, linear depression with steep flanking slopes and a smooth floor, sloping very gently southward. It terminates at the flat Emerald Basin floor at about 51° 30′ S, near the Auckland Islands. From the Solander Islands in the north to the Auckland Islands in the south the trough is 310 miles long. The northern part is 20 miles wide in 1,800 m and 50 miles in 450 m (Brodie, 1958). Off the Auckland Islands it is 55 miles wide in 3,000 m and about 60 miles wide in 1,800 m. From the Solander Islands to 49° S the trough width remains fairly constant at about 15–20 miles in its deeper parts. Further south it abruptly increases in width to 50 miles and maintains this width to its southern limits. The trough walls are more or less parallel over the whole of its length.

The walls of the Solander Trough have been described in detail in earlier sections. The trough floor has a southward slope of 0·25° (Traverse 4d). It is 2,780 m deep off the Snares (Traverse 6a) and deepens to 4,000–4,250 m at 51° 30′ S. In profile the trough is broadly U-shaped, its smooth floor sloping almost imperceptibly toward the centre. Slightly west of its centre is a narrow median channel some 5 miles wide and 160 m deep (compare Traverse 6c). In profile, the channel is V-shaped with walls inclined at 2·5°–3·0°. Levees are not developed along margins. It appears to be continuous north of the charted area (Hatherton, 1967).

West of the Auckland Islands three seamounts rise sharply from the centre of the Solander Trough (compare Traverse 6d). A major peak reaches a minimum recorded depth of 915 m and a minor peak reaches 2,377 m. These features are aligned approximately north–south along the centre of the trough, are 20–30 miles apart, and are flanked by 10°–30° slopes. Between these seamounts and the Aucklands Slope are narrow, depressions, 100 m deep, which trend parallel to the slope. Similar depressions also occur further north toward the Snares.

EMERALD BASIN

Emerald Basin is a broad elongate basin lying between the Macquarie Ridge and the Campbell Plateau. The subsidiary Macquarie Ridge forms the western margin of the basin over most of its length. Brodie and Dawson (1965) do not describe the Emerald Basin although they mention the sill in just less than 4,000 m separating the Emerald Basin from the Southwestern Pacific Basin. Southward the basin terminates against a broad rise toward the Macquarie Ridge. The Emerald Basin forms a broad, deep, flat-floored continuation of the Solander Trough as far as 56° S.

Echo-sounding coverage of the Emerald Basin is extremely poor but is sufficient to illustrate some aspects of its morphology.

Emerald Basin is divided into two large basins separated by a sill in a little less than 4,500 m. The northern part lies in 4,250–4,500 m, and the southern in 4,500–4,750 m. Westward the sill joins the subsidiary Macquarie Ridge at its widest point. The Emerald Basin is floored by an abyssal plain sloping gently south and west. Isolated, steep-sided, abyssal hills rise from the plain to a recorded minimum depth of 3,850 m (Traverses 5a–b). A 5,140 m deep occurs in the centre of the basin, separating an eastern abyssal plain from a western region with a more rugged floor, at the same general level (Traverse 5a).

A poorly developed continental rise occurs locally at the foot of the Aucklands Slope west of the Auckland Islands Ridge. Southward the junction between the basin floor and the Aucklands Slope is abrupt and a continental rise is not developed (Traverses 5a and b). Locally, shallow depressions occur in the abyssal plain near the foot of the Aucklands Slope (Traverse 5a).

SOUTHWESTERN PACIFIC BASIN

ABYSSAL PLAINS

Flat or very gently sloping featureless plains form the floor of the Southwestern Pacific Basin at the foot of the continental rise (Traverses 9a–d). These plains are restricted basinward by the regional development of an abyssal hill province (Fig. 6). They generally lie between 4,100 and 5,400 m, are 60 miles wide north-west of Endeavour Banks, and only 10 miles wide south of the banks. Locally, the plains are diversified by abyssal hills, either single or in small groups. Abyssal plains have not been found at the foot of the Subantarctic Slope further west than 169° E.

ABYSSAL HILLS

South of the abyssal plains the Southwestern Pacific Basin floor is characterised by rough topography. Numerous small elevations with vertical relief of less than 1,000 m rise from the general level of the basin floor. These abyssal hills cover the whole of this region (Traverses 9a–d), forming an abyssal hill province. Individual hills are flanked by 0·5°–25·0° slopes, have flat or rugged tops, and vary in width across the base from 10 to 100 miles.

The bathymetry of the Southwestern Pacific Basin was based on soundings recorded at approximately 15-minute intervals along traverses oriented north–south across the area. From these soundings it is possible to determine the distribution of depth within the basin. Owing to the non-random distribution of sounding traverses the depth distribution and the average depth determined will have an inherent bias toward north–south-sounding traverses. Frequency of depth distribution within the abyssal hill province is derived from the given soundings plotted in 100 m-depth intervals (Fig. 18). The average depth of the Southwestern Pacific Basin abyssal hill province is 5,130 m, about which the depth distribution approaches a normal or Gaussian form (Fig. 18).

SEAMOUNTS

Six large seamounts rise high above the regional level of the basin, one reaching a recorded minimum depth of 3,060 m at 57° 23′ S, 172° 57′ E. The distribution of echo-soundings indicates that these recorded minima are isolated; so they are probably seamounts (Traverse 9d).

None of these features has been adequately surveyed and it is not possible to determine whether or not they are guyots. Incidence of seamounts in the abyssal hill province is responsible for the skewed nature of the depth distribution (Fig. 18).

RISES

A broad, elongate, gentle rise trends north–south across the basin between 55° 30′ S and 52° 30′ S. It is some 30–60 miles wide and reaches a general elevation of 4,850 m. An isolated elevation along the rise reaches 3,600 m forming one of the seamounts of the above group.

TASMAN BASIN

A small part of the Tasman Basin lies within the area covered by the Auckland and Macquarie 1:1,000,000 charts. Marked paucity of data from this region makes detailed discussion of the morpology impossible.

The eastern margin of the basin is formed by the Macquarie Ridge. Between the Tasman Basin floor and the Macquarie Ridge a narrow, discontinuous trench locally reaches depths greater than 6,000 m. This trench is separated from the regional depths of the Tasman Basin by a broad elevation to 3,000–4,000 m. The regional depth of the basin over the northern part of the area is about 4,250 m, increasing southward to 4,750 m, slightly shoaler than the regional average depth of 4,850 m (Standard, 1961). From available echo-sounding data, it appears that the basin floor is not smooth, and vertical relief of up to 500 m may occur locally (Traverses 6a, c).

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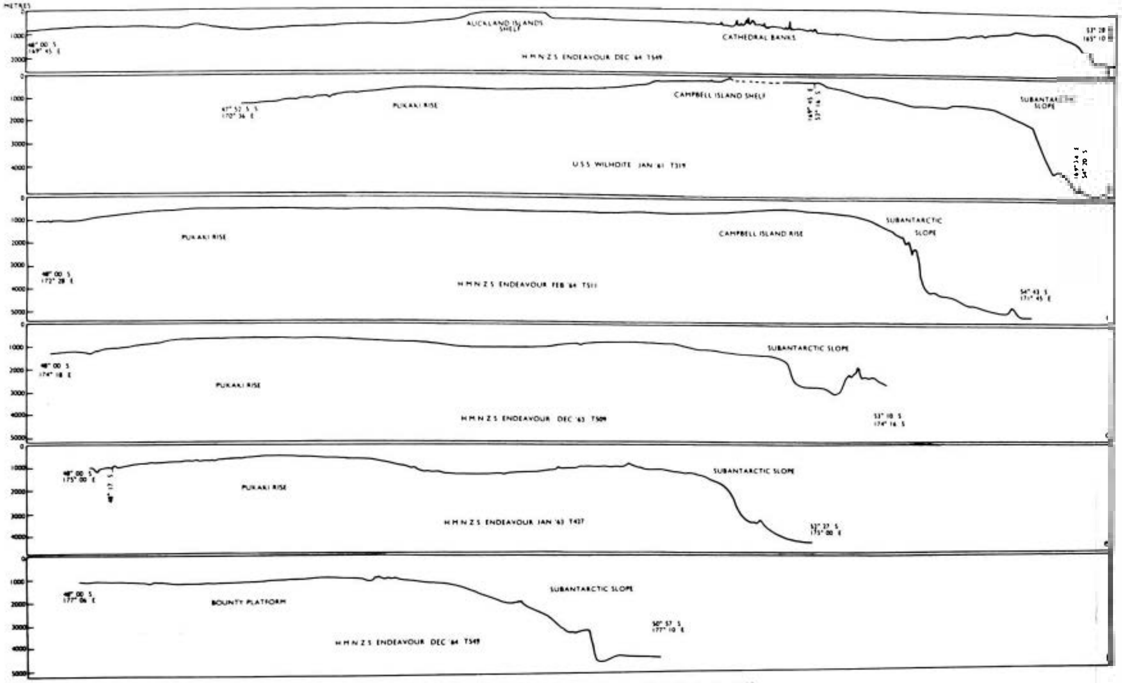
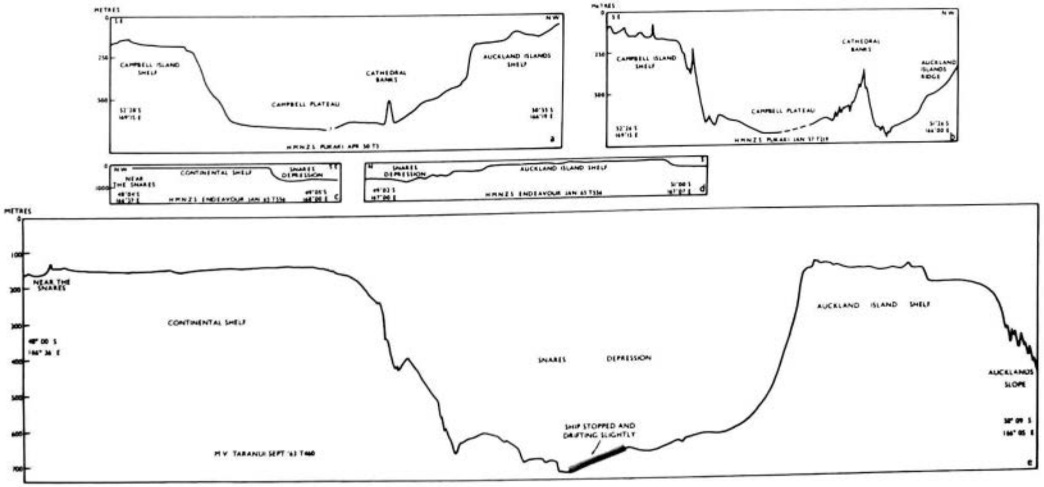
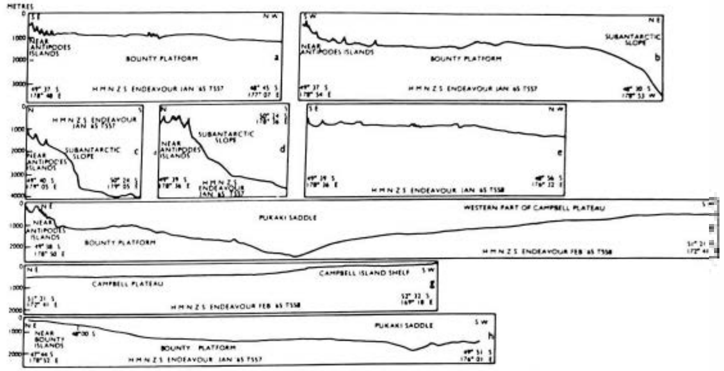


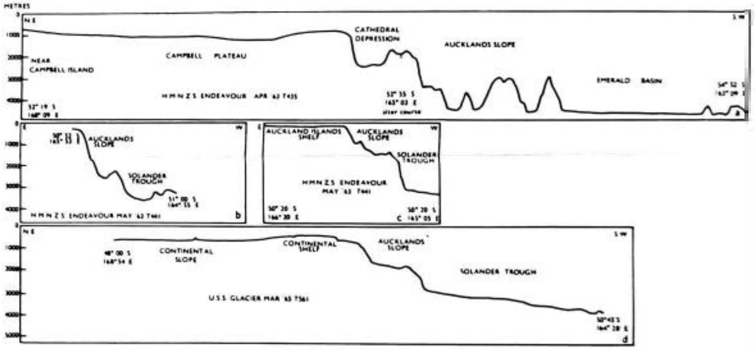
TRAVERSES 1a - f: Campbell Plateau. (e also published by Brodie, 1965.)



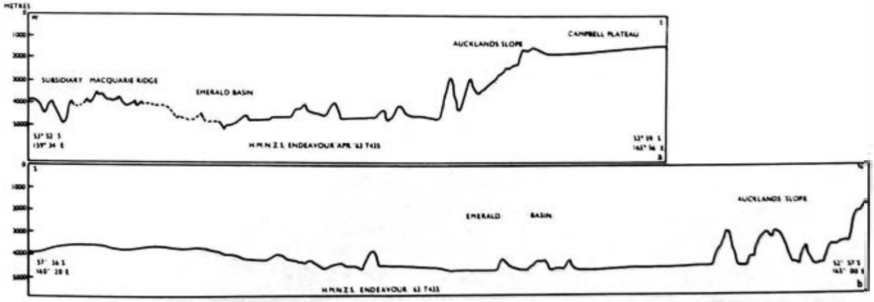
TRAVERSES 3a - e: Shelf Regions, Cathedral Banks, and Snares Depression.



TRAVERSES 2a - h: Antipodes Islands, Bounty Platform, and Pukaki Saddle



TRAVERSES 4a - d: Aucklands Slope.



TRAVERSES 5a - b: Emerald Basin and Aucklands Slope.

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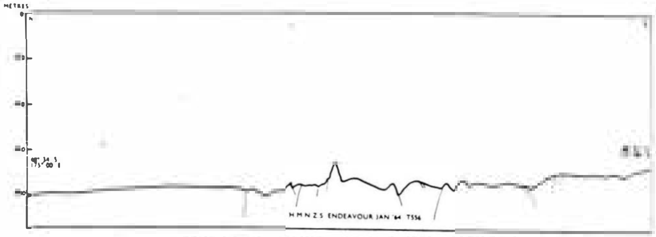

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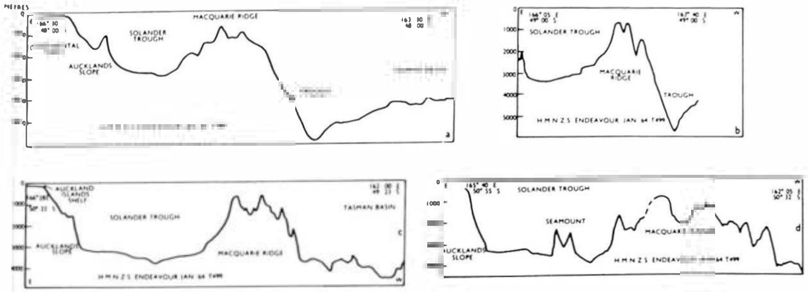

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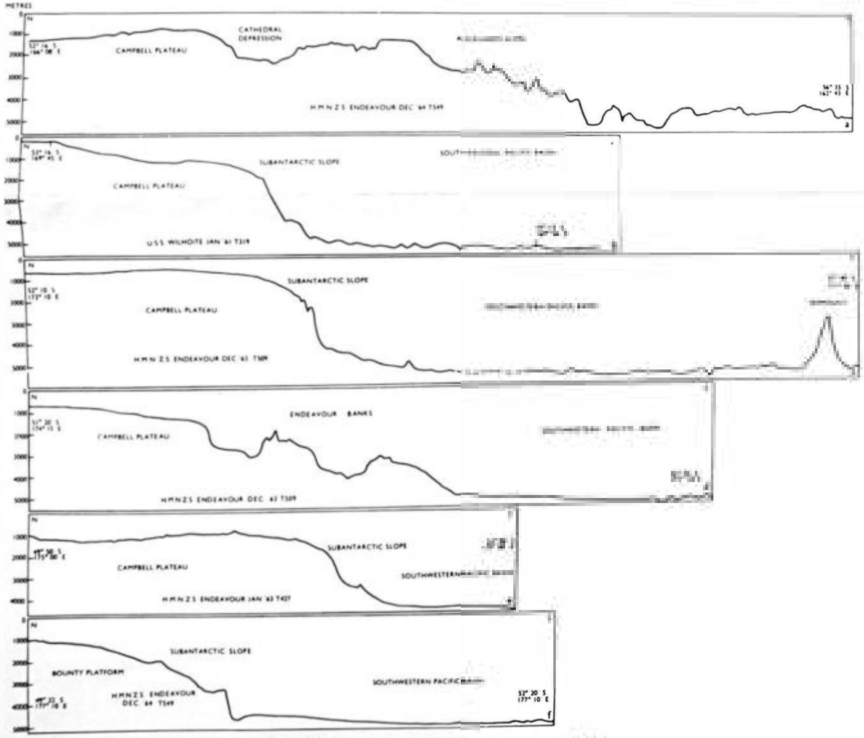
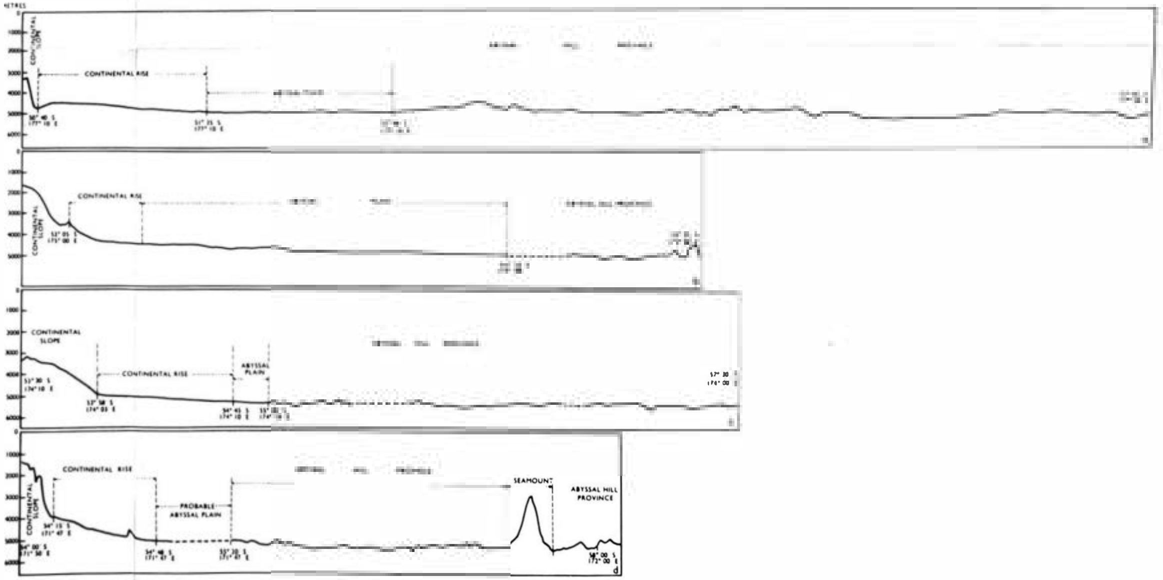


TRAVERSES 8a - f: Subantarctic Slope. (e[illegible] published by Brodie, [illegible]



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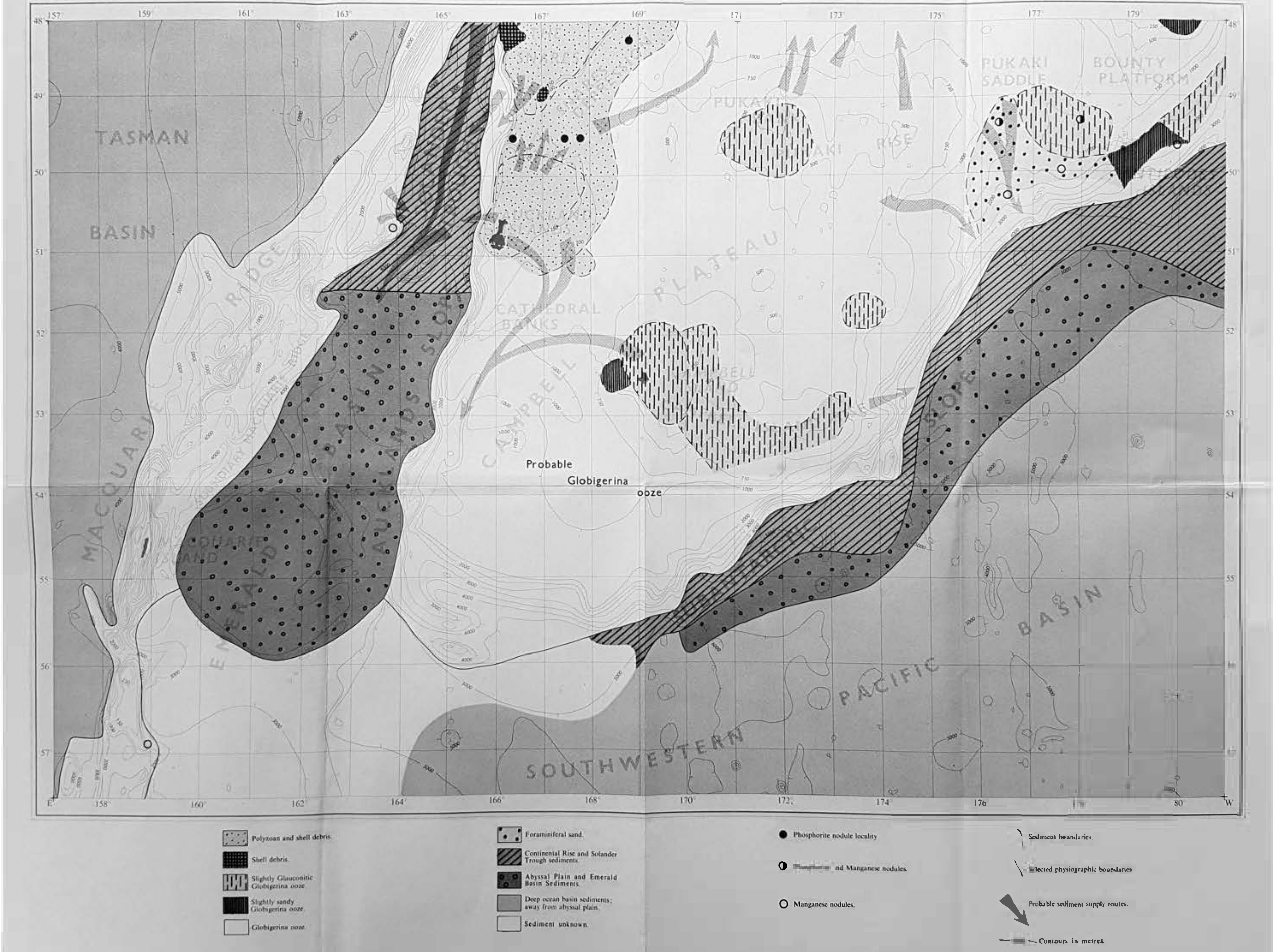

Fig. [illegible] Sediment distribution, sedimentary environment, and sediment movement patterns

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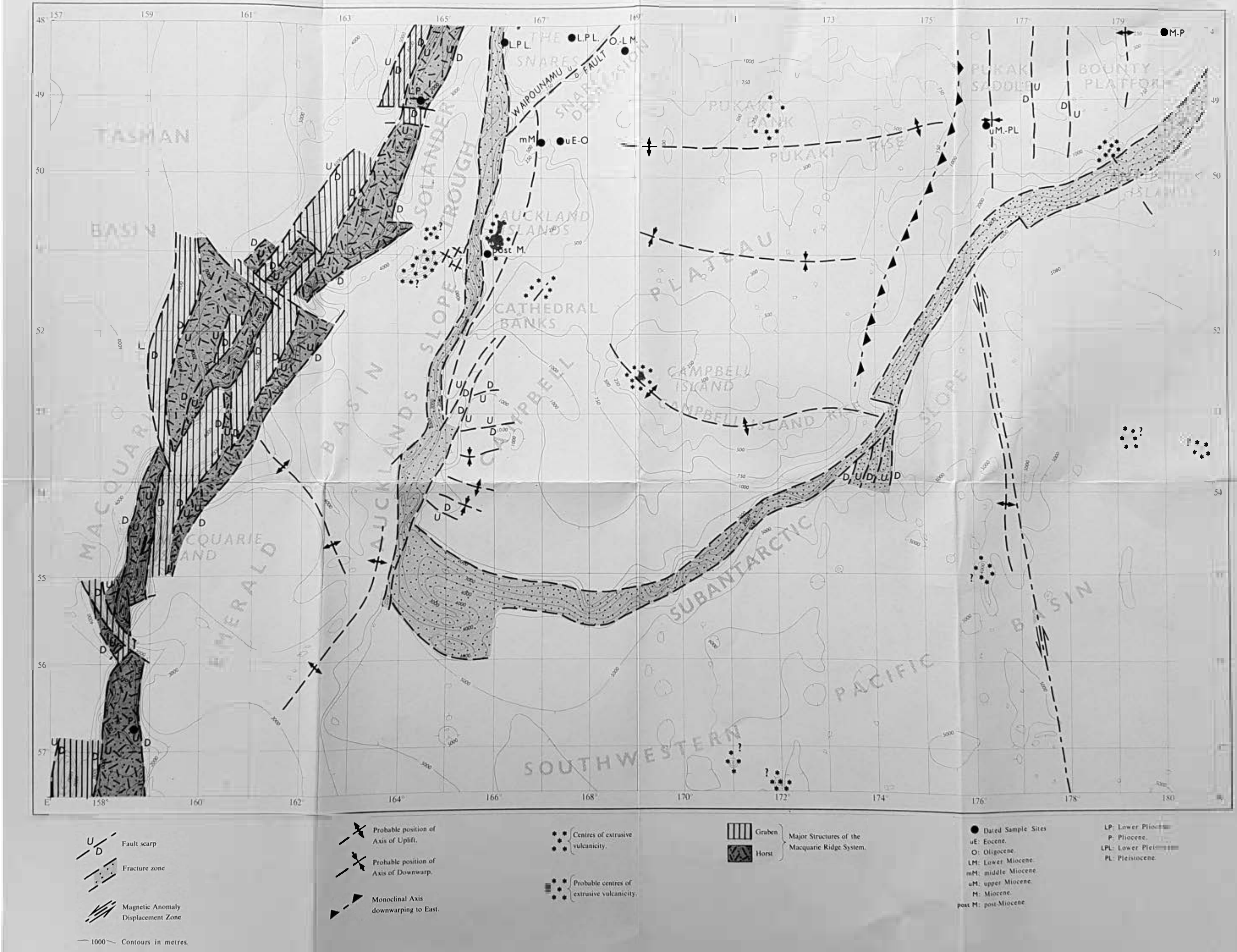

Fig. 11. Major geological [illegible]