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RECORD 1975/118

"PAPERS FOR G.A.M. TAYLOR MEMORIAL VOLUME TO
BE PUBLISHED AS A BOOK BY ELSEVIER"

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CONTENTS

# VOLCANISM IN AUSTRALASIA

A collection of papers in honour of
the late G.A.M. Taylor, G.C.
Edited by
R.W. Johnson

*Bureau of Mineral Resources, Geology, and Geophysics (Canberra)*

**G.A.M. TAYLOR, G.C.**

## PREFACE

The range of subjects covered in this volume, and the relative proportions of papers covering different topics, are a close indication of G.A.M. Taylor's particular interests in volcanism. Because his overriding concern was for an understanding of the eruptive mechanisms of island-arc volcanoes in the Australasian region — and especially in Papua New Guinea — it is appropriate that most of the papers presented here deal with Australasian eruptions and with the Quaternary geology of Melanesian volcanoes. However, Mr Taylor's outlook was much broader, and during the last years of his life he was involved with other subjects, such as seismic refraction studies of island-arcs, the petrology of volcanic rocks, the practicalities of establishing volcanic surveillance networks, and geological field studies of ancient volcanic formations. The papers presented here reflect this broad overview of volcanism in Australasia.

All the papers (except one) deal with specific areas in the region, and the chosen order of presentation is therefore a geographic one, progressing clockwise from Australia to New Zealand by way of Indonesia, Papua New Guinea, the Solomon Islands, and Tonga. This route takes the reader from an intra-continental environment into the complex island-arc systems that essentially define the northeastern and eastern margins of the Indo-Australian plate.

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R.W. Johnson
October, 1975

Canberra

# OBITUARY OF G.A.M. TAYLOR

by N.H. Fisher

This volume is dedicated to the memory of G.A.M. Taylor, whose premature death whilst engaged on field work on Manam volcano in August, 1972, was a great shock to his many friends and colleagues. Ever since graduation, Mr Taylor's whole professional career had been devoted to the study of volcanoes and volcanic centres, mainly of Papua New Guinea and Australia, but also of other countries, and he was well known internationally for his contributions to volcanology and particularly for his study of Mount Lamington after the disastrous eruption of 21 January, 1951.

George Anthony Morgan Taylor, always known as Tony Taylor, was born in Moree, northern New South Wales, on 30 October 1917. His father was an Englishman who had come to Australia in the early 1900's, and settled at Weemelah, about 75 km northwest of Moree and 25 km east of the Queensland border town of Mungindi. Tony Taylor received his early education at the local primary school, and then attended Maitland High School near Newcastle, but in 1936 he moved to Sydney and at the end of that year passed his Matriculation Examination from Sydney Boys' High School.

At school his favourite subject was chemistry and he maintained at home a 'laboratory' which his two younger sisters were occasionally allowed to visit as a rare treat to observe his 'experiments'. He was a member of the Young Naturalists and keenly interested in all aspects of nature study, including geology, though it must be admitted that the alluvial plains of the Moree-Mungindi district are not a very fertile field for the study of rocks. This interest was to develop later.

8/ After matriculation he joined the Broken Hill Proprietary Company (BHP) as a staff trainee analytical chemist, and remained in the employment of this company until he enlisted in the Second A.I.F. in April, 1942. His position at BHP/ involved shift work, with on-the-job training, and attendance four nights a week at a chemistry diploma course at the Newcastle Branch of the Sydney Technical College, a course which he had not completed at the time of his enlistment. This schedule did not leave much time for other activities but he played Rugby Union Football for a couple of years as a winger with one of the teams of the Newcastle Wanderers Club. At school he had played football and had also represented the school in athletics — running and high jump.

8/ During his period with BHP/ it was discovered that Tony Taylor had a degree of colour-blindness and as some of the analytical tests depended on accurate perception of fine colour differences this imposed some limitations on his future as 8#/ an analytical/chemist.

On 29 April 1942 he enlisted in the Australian Imperial Force and was posted to the 2/4 Ordnance Stores Company. He spent most of the next three years in North Queensland, where his first interest in volcanoes was aroused by the numerous volcanic forms and evidences of fairly recent activity that are so abundant in the Cairns Hinterland.

In May 1945 he embarked in Brisbane on the 'Duntroon' for Jacquinot Bay, New Britain, and moved on 9 September, after the Japanese surrender, into Rabaul where he remained until the end of October 1946. Here he had the opportunity to observe at length the many manifestations of volcanism around Rabaul Harbour.

On 8 January 1947 Tony Taylor was discharged from the Army with the rank of Warrant Officer 1st Class and in March of the same year he began a Science Degree Course at the University of Sydney under the Commonwealth Government's Post-War Reconstruction Training Scheme. He completed the requirements for a Bachelor of Science degree in three years and joined the Bureau of Mineral Resources of the Australian Government's Department of National Development as a Geologist Grade 1 on 20 March, 1950.

an/ The Bureau of Mineral Resources had been established in May/June 1946 and one of its responsibilities was to provide geological and volcanological services for the Administration of the Territory of Papua ~~and~~ New Guinea. Lack of available geologists in Australia and absence of geological training in universities during the war years had made it very difficult for the BMR to build up its staff in its early years, until the post-war graduates started to become available. Consequently the Volcanological Observatory in Rabaul, which had been started in 1940 and destroyed during the war, had not been re-established and this was Taylor's first task when he was posted to New Guinea in April 1950. The Observatory was rebuilt on the same site, over the same instrument cellars, which were intact and usable, the *Benioff* seismographs destined for the Observatory which had been on loan for several years to Queensland Uniersity, were installed, and systematic observations of the Rabaul volcanic centres resumed. Re-establishment of the Rabaul surveillance system occupied most of Taylor's attention, but the volcanologist's responsibilities included the provision of advice on the activity of all of Papua and New Guinea's volcanoes. One of his early field visits was to Bougainville Island to examine Balbi and Bagana volcanoes and to report on the possibility of Pelean-type eruptions from the latter, one of the Territory's most active volcanoes. But it was the catastrophic and disastrous eruption of Mount Lamington in Easter Papua on 21 January, 1952 that catapulted Taylor, normally one of the most reserved and retiring of men, ito public prominence and provided the opportunity to put into effect the studies of the effects of volcanic eruptions of various types that he hda been assiduously pursuing.

Mount Lamington was not previously known even to be a volcanic site, and although preliminary eruptive activity began several days before the great Pelean outburst, inadequate communications and transport difficulties prevented any expert advice arriving on the scene before the main eruption. Taylor therefore did not reach Popondetta, the nearest point of access to Mount Lamington, until immediately after the disaster, when the bodies of 3000 recent inhabitants lay dead around the fertile northern slopes of Mount Lamington, surrounding the village of Higaturu. The volcano was relatively quiet but nobody knew what type of activity it was likely to produce in the immediate future, and the tasks of care and rehabilitation of the injured and displaced people, assessment of the damage and of the future of the area, as well as disposal of the dead, had to proceed immediately.

An observation post, equipped with seismographs, was set up at Sangara Plantation, 16 km from the volcano, and systematic observations commence. Excursions were made into the crater area to examine the growing lava dome and the distribution of products of the main eruption, and to collect samples. Of particular interest was the correlation of the growth of the lava dome with the almost continuous seismic activity recorded on the seismographs at the Observation Post.

Much of Tyalor's activity uring the next two years was devoted to the surveillance of Mount Lamington, studying the later eruptions, advising the Administration on the location of evacuation camps, the movement of people in the devastated areas, the return of village groups to their areas and the resumption of agricultural activity. Some of the later erutions of the nuee ardente type were particularly fearsome and awe-inspiring, hurtling down the valley of the Ambogo River in huge clouds and threatening to completely engulf and destroy the Observation Post. It was Taylor's stadfastness and devotion to his work during this troubled period that led to his being recommended for, and awarded in April 1952, the George Cross, the citation for which read as follows.

> ~~CITATION~~ *Mount Lamington, in Papua, began to erupt on the night of 18th January 1951. Three days later there was a violent eruption when a large part of the northern side of the moutain was blown away and steam and smoke poured from the gap for a considerable time afterwards. The area of extreme damage extended over a radius of about eight miles, while people near Higaturu, nine miles from the volcano, were killed by the blast or bued to death. This and subsequent eruptions caused the death of some 4,000* persons, and considerable damage. Dust and ash filled every stam and tank for some miles around, and there was urgent need of food, water and medical supplies. Rescue parties were hampered by suffocating pumic dust and suphurous fumes, and hot ashes on the ground, and the advance post of relief workers at Popondetta was threatened with destruction by other eruptions during several days following. Further tremors and explosions occurred during February. As late as March 5th a major erution occurred which threw as far as two miles pieces of volcanic dome, 15 ft. by 12 ft. by 10 ft. and caused a flow of pumice and rocks for a distance of nine miles, the whole being so hot as to set fire to every tree in its path*

*For a prolonged period Mr Taylor showed conspicuous courage in the face of great danger. He arrived at Mount Lamington on the day following the main eruption and from that day onwards, over a period of several months he visited the crater by aircraft almost daily, and on many other occasions on foot. On some occasions he stayed at the foot of the volcano throughout the night. During the whole of this period the volcano was never entirely quiet. Several eruptions took place without any warning or any indications from the seismographical data which he had collected. Without regard for his personal safety he entered the danger area again and again, each time at great risk, both in order to ensure the safety of the rescue and working parties and in order to obtain scientific information relating to this type of volcano, about which little was known. His work saved many lives, for as a result of his investigations in the danger zone he was able, when necessary, to warn rehabilitation parties and ensure that they were prevented from entering an area which he so fearlessly entered himself. (London Gazette: 22nd April 1952.).*

During 1951, in addition to his Mount Lamington work and general oversight of the Rabaul surveillance system, Taylor was called upon again to investigate reported activity of volcanoes on Bougainville, and to examine other volcanoes in Papua to check against possible sympathetic reactivation in response to the Mount Lamington eruption. He also was made available by the Australian authorities, at the request of the Condominium Administration of the New Hebrides, to examine and report on Ambrym volcano which had been erupting for a year causing great devastation, and to advise on the safety of the inhabitants still remaining on the island.

During the next few years Taylor continued his general studies of the New Guinea volcanoes, making specific field investigations wherever and whenever signs of eruptive activity became apparent. The first to show such signs was the long dormant Langila volcano at the western end of New Britain, in 1952. Eruptions of Long Island and Bam took place the following year, Tuluman, a submarine volcano near the southwestern tip of Manus, and Langila in 1954, and Manam began an eruptive period of major proportions in 1956 after a dormancy of 11 years.

A particularly exciting incident took place in June 1955 when Taylor and one of the local people attempted in an improvised boat to examine the then active Motmot island crater in the southern part of the lake that occupies the caldera of Long Island, Lake Wisdom. An explosion occurring as the frail craft was approaching the crater placed the expedition at considerable risk and presented the intrepid investigators with unsolicited samples of the ejecta.

During the period 1954 to 1956 Taylor spent a considerable part of his time in Canberra, studying the Mount Lamington seismograms and other records, carrying out laboratory work and experiments of various kinds, and writing a comprehensive report on the Mount Lamington 1951 eruption. This report was published in 1958 as Bulletin No. 38 of the Bureau of Mineral Resources and ranks as probably the most complete and authoritative analysis of a major Pelean-type cataclysmic eruption and subsequent phenomena that has been written. In 1957 he was awarded the degree of Master of Science by Sydney University in recognitiion of his Mount Lamington work.

Returning early in 1957 to Rabaul, Taylor found the Manam eruption increasing in severity, culminating in a series of climactic eruptions during the first three months of 1958. Two instrument stations were established on the island, and the 3000 inhabitants were evacuated to the mainland in December 1957. They were returned to the island in August the following year.

Meanwhile the development of the Rabaul system of Observation Posts was proceeding, with improvements in the seismic instrumentation and the installation of tiltmeters. Observations posts were established at Sulphur Creek and later in the caldera wall near Tavurvur, in a tunnel at Taviliu overlooking Vulcan, and at

Rabalankaia. Activity at Manam continued with another series of major eruptions in March-May 1960, but evacuation of the island was avoided by defining danger areas and maintaining a 24-hour vigilance. Bam Island and Langila erupted in 1958. Advice was given to the administrations of the British Solomon Islands Protectorate and of the New Hebrides on volcano surveillance systems.

For some years Taylor had been studying the possible effects of luni-solar phenomena in triggering off eruptions and also the possible correlation between regional seismic acitivity and the incidence of eruptions at the volcanic centres of Papua and New Guinea. In 1959 he attended a meeting in Paris of the International Association of Volcanology (IAV) and presented a paper on the prediction concepts, including those mentioned above, used on Manam volcano. During this overseas trip he examined volcano research and surveillance centres as Vesuvius, in Japan, and in the Philippines.

In February 1961 Taylor was promoted to the position of Senior Resident Geologist, Port Moresby, and became responsible for supervision of all official geological and volcanological work in Papua and New Guinea. Here he was able to advance plans to upgrade the volcano surveillance instrumentation at Rabaul by establishing a telemetered seismic network connecting the Central Observatory with the Observation Posts situated, as mentioned above, at strategic locations around the caldera, and also to establish permanent volcanological observatories at Manam, Mount Lamington, and Esa-ala in easter Papua. In 1961 he attended a meeting of the IAV in Japan.

In 1963 Taylor became the first scientist to make the difficult descent to the caldera floor of Karkar volcano to examine the craters and volcanic products within the calder. This volcano had last erupted in 1895 and was to become active again in 1974.

At the end of 1963 Taylor transferred his headquarters to Canberra, but continued to devote most of his attention to Papua New Guinea volcanological problems and spent a considerable amount of time in Papua New Guinea on field work and on development of volcanological stations. During the succeeding years he played a major part in planning and carrying out crustal study projects in the New Britain region. In these projects shots were fired at sea at locations along two extended lines at right angles to each other and the resultant seismic waves measured at a number of field stations set up especially for the purpose. The 1967 exercise involved the use of two ships and twelve field parties, the 1969 one three ships and eighteen field parties. Co-operating institutions as well as the Papua New Guinea resident staff and the Bureau of Mineral Resources included the University of Hawaii, University of Queensland, Australian National University, and the Queensland Geological Survey. These projects made an important contribution to the knowledge of tectonics and crustal structure in the Melanesian region and to the determination of travel times of seismic waves and thereby the precise location of points of origin of local earthquakes.

In 1966 Taylor examined recent lava flows and volcanic centres in North Queensland and Victoria, and the following year prepared for the IAV a report on work in Palaeovolcanology and Plutonism in Australia and for the Upper Mantle Committee a progress report on volcanological work and its results. Much of his time was occupied with preparation of data sheets and maps concerning over 800 post-Miocene volcanoes in Australia and Melanesia in time for a meeting of the IAV at Zurich as part of the IAV's compilation of post-Miocene Volcanoes of the World.

An interesting project during 1967 was the investigation of a very high ~~[illegible]~~ of a landslip area at 'the Crater', at Koranga, near Wau. The occurrence lasted only about three months, and the gases appeared more likely to have been generated by oxidising sulphides at depth rather than by genuine rejuvenating vulcanicity.

temperature (at t... thermal vent which... in the middle

During 1968 Taylor was the convenor of a symposium on palaeovolcanology in Australia and Papua New Guinea organised by the Geological Society of Australia; he contributed a paper on post-Miocene volcanoes in Papua New Guinea, and edited the volume that was subsequently published.

The crustal study projects and various field investigations in Papua New Guinea occupied most of Taylor's time during the next few years. He prepared a booklet for a volcanological excursion in Papua New Guinea, attended the ANZAAS Congress in Port Moresby in August 1970, and led the post-congress excursion. The same year he paid, by invitation, a visit to New Zealand to examine and discuss volcano surveillance techniques and to present a paper at the Annual Meeting of the New Zealand Geological Survey. In Papua New Guinea he took part, together with resident staff, in the investigation of a full-scale eruption of Ulawun on New Britain, advised the Administration on the return of village population to affected areas, and prepared a report on the eruption. He also continued his long-term study of Manam volcano, about which he had accumulated a tremendous amount of data during its many years of eruptive activity. His death on August 19, 1972, from a heart attack, whilst actually carrying out field work on this volcano, prevented the completion what would no doubt have been another major contribution to volcanological literature.

Taylor was quiet and reserved by nature. He was not normally loquacious although he was always willing to discuss at length the subjects in which he was especially interested, particularly, of course, volcanoes, volcanism and volcanic prediction. He considered carefully what he wished to say and was tenacious in argument where his convictions were concerned. He was a careful writer with the special gift of presenting his material in an interesting, even absorbing way, and many of his papers make fascinating reading. He had the ability to convey to the reader a special sense of excitment and involvement in the subject being discussed. He had the patience that is often characteristic of quiet men and pursued his aims with persistence and diligence. He was a good craftsman, with a wide understanding of construction methods and instrument operation, a most useful accomplishment in a volcanologist engaged in the development of surveillance systems in remote locations, where he has to plan, supervise installation, adjust and improvise to meet the requirements of the situation. He never allowed consierations of comfort or personal safety to affect his approach to his volcanological studies, and in fact tended to pay insufficient attention to his material requirements whilst on field excursions.

Taylor had a deep and continuing interest in, almost an obsession with, volcanology, and the main thrust of his work was devoted towards establishing and improving methods of prediction of volcanic eruption and putting them into practical effect. His work was his main hobby but he was a keen fisherman, played occasional golf and tennis, and was a successful gardener.

Taylor had a strong, somewhat dry, sense of humour. He was always well respected by his colleagues and regarded with deep affection by those fortunate enough to penetrate his quiet reserve and gain his friendship. He was married on April 4, 1956, to Lindsay Hudson; he was a devoted family man, and is mourned by his wife, two sons and a daugher.

Plate 1: Oblique aerial photograph (camera tilted) of lava flow source, Ulawun volcano, 5 October 1973. Width of the source is about 3 m. The identity of the fin-like monolith (upper left corner) produced during the 1970 eruption, is not known. Photograph by R.J.S. Cooke.

Plate 2: Explosion at crater 3, Langila volcano, 22 July 1973. Incandescent lava surface is about 10m across. Photograph by R.J.S. Cooke.

Plate 3: Lava jet at crater C, Bagiai cone, Karkar volcano, mid-May 1974. Height of jet is about 180 m. Photograph by D.. Wallace.

Plate 4: Explosion at crater C, Bagiai cone, Karkar volcano, mid-May 1974. Photograph by D.A. Wallace.

Plate 5: Lava source 20, base of Bagiai cone, Karkar volcano, 16 January 1975. Width of source is about 3 m. Photograph by C.O. McKee.

* Although the exact number of dead was never determined, the usually quoted figure, and the one used by Taylor, is almost 3000. Ed.

FLOOD BASALTS OF PROBABLE EARLY CAMBRIAN AGE IN NORTHERN AUSTRALIA

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ABSTRACT ——> centre

The Antrim Plateau Volcanics, a flood-basalt sequence of probable early Cambrian age, crop out over an area of about 35 000 sq km in the East Kimberley and Victoria River regions of northern Australia, and are known from drill-hole intersections to underlie most of the Daly River and northern Wiso Basins containing predominantly Lower Palaeozoic sediments. Small, isolated outcrops of basic volcanic rocks exposed around the northern and western margins of the Georgina Basin and the eastern margin of the Daly River Basin probably originally formed part of the same extensive lava field but have become separated from the main mass of volcanics by erosion and a cover of younger sedimentary rocks.

Little-altered, massive rocks from the central parts of flows consist essentially of labradorite, clinopyroxene, opaque minerals, and a mesostasis of devitrified glass or a quartz-alkali feldspar residuum. Pseudomorphs of olivine are present in a few flows, but the great majority are olivine-free. The overall tholeiitic character of the suite is indicated by the common presence of modal pigeonite and quartz. Chemically the lavas form a homogeneous group; all have hypersthene in the norm and many also bear normative quartz. They range in composition from ~~under-~~saturated olivine tholeiite to quartz tholeiite and tholeiitic andesite. Distinctive chemical characteristics of most of these lavas are their relatively high $SiO_2$ (between about 52 and 54 per cent), ~~and~~ low $TiO_2$ (between about 0.8 and 1.4 per cent), and total Fe (between about 8 and 11 per cent, as FeO) contents compared with other flood-basalt sequences. They are not high-alumina types, a characteristic shared with most flood basalts.

Fig 1.

## INTRODUCTION

The Antrim Plateau Volcanics are an extensive, but hitherto little-studied, sequence of flood basalts* ~~[illegible]~~ in northern Australia (Figs. 1, 2). They form one of the more extensive accumulations of flood basalts ~~[illegible]~~ in the world (Table 1) and this paper represents a first attempt to provide a comprehensive account of their geology and petrology.

Extensive inland erosional plains covered by black and grey cracking clay soils, benches, mesas, buttes, and low rounded residual hills are the characteristic landforms developed on the essentially flat-lying (regional dips less than $<1^{\circ}$) lavas and interbedded sediments in the eastern Victoria River region (Fig. 2). In contrast, the volcanics in the East Kimberley and western Victoria River regions have been more deeply dissected and the country is more rugged. Plateaux ~~and benches~~ with boulder-strewn mesas and hills up to about 100 m high have formed where the lava flows are flat-lying or gently dipping (regional dips less than $<5^{\circ}$; Fig. 3). Valley walls are steep and are generally broken by structural benches. Cuestas and hogbacks are well developed on the moderately dipping ($20^{\circ}$ – $30^{\circ}$) lava flows around the western margin of the Hardman Basin (Fig. 4).

UPPER CASE

### Previous Geological Investigations

Antrim Plateau Volcanics in the East Kimberley region were first described by Hardman (1885) who was government geologist attached to the Kimberley Survey Expeditions. The first recorded observations on the geology of the Victoria River hinterland were made by the explorer A.C. Gregory (~~[illegible]~~ 1858). H.I. Jensen (1915) described the volcanic rocks in the Victoria River region and referred to reported discoveries of native copper from several places in the volcanics. A.B. Edwards (Edwards & ~~and de C.~~ Clarke, 1940) analysed and described basic volcanic rocks collected by ~~de C.~~ Clarke in 1927 from the East Kimberley region and concluded that the lavas formed a single homogeneous petrographic province. In 1945, Matheson & ~~and~~ Teichert (1948) made a geological reconnaissance of the Cambrian rocks in the East Kimberley

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*Following common practice the term 'basalt' is used in a general sense throughout the text to describe the lavas of the Antrim Plateau Volcanics and their correlatives.

region. Their report contains an excellent bibliography and summary of the literature up to 1945. Pyroclastic intercalations in the volcanics were reported by Traves (1955). More recently, Dunn ~~and~~ & Brown (1969) correlated probable Early Cambrian basic volcanics in other parts of the Northern Territory and northwestern Queensland with the Antrim Plateau Volcanics.

STRATIGRAPHY

The Antrim Plateau Volcanics consist predominantly of basaltic ~~[illegible]~~ lava flows, and minor flow breccia and agglomerate. Explosive volcanism appears to have played only a very minor role in the igneous activity. Thin beds (generally less than 10 m thick) of quartz sandstone, siltstone, chert (commonly stromatolitic), sedimentary breccia, and limestone are intercalated with the lava flows in many places. Most of the interbeds are of small extent.

The formation is separated from the underlying Precambrian rocks by a marked erosional and regional angular unconformity. Locally, the lavas rest conformably on thin layers of sandstone or conglomerate mapped as Lower Cambrian, because of their unconformable relationship with underlying Proterozoic strata.

The volcanics are overlain by early Middle Cambrian sediments consisting predominantly of shallow-water marine limestone, dolomite, shale, siltstone, and sandstone with an aggregate thickness of less than 600 m. South of Wyndham the Antrim Plateau Volcanics and the overlying Blatchford Formation are reported to be separated by a low-angle unconformity (Kaulback ~~and~~ & Veevers, 1969). Traves (1955) found basalt pebbles in basal shales overlying the volcanics in this area. Farther south the volcanics are overlain by the Negri Group, now mainly confined to the Hardman, Rosewood, and Argyle Basins. Regional angular unconformity between the Headleys Limestone (the basal formation of the Negri Group) and the volcanics has not been observed by the writer (however, see Hardman, 1885), but the lava field was eroded prior to the deposition of the Headleys Limestone. In several places around the margins of the Hardman and Rosewood Basins, Headleys Limestone is underlain by thin, poorly exposed, irregular lenses of predominantly sedimentary breccia, siltstone, and sandstone ranging from about 1 m to 7 m thick. The breccias consist of large, angular to sub-rounded clasts (2-60 cm across) of intensely altered vesicular basalt ~~(sensu lato)~~ in a matrix of mainly soft, dark red-brown, calcareous siltstone and fine-grained, red-brown, friable sandstone.

Weathering profiles up to 6 m thick have been reported (unpublished BMR Record 1964/104) in volcanics underlying Headleys Limestone around the northern and western margins of the Hardman Basin implying that, although the two formations appear to be conformable, there had been a period of subaerial weathering and possible erosion of the lava pile before the Headleys Limestone was deposited.

l.c.

Extent, Thickness, and Correlation

The formation has been extensively eroded since the ~~Lower~~ Early Cambrian, and much of the lava field covered by younger, predominantly Palaeozoic sediments (Fig. 2). Volcanics forming the eastern belt of Traves (1955) are exposed along the western margins of the Wiso and Daly River Basins. Volcanics of the western belt (Traves, 1955) crop out extensively in the East Kimberley and western Victoria River regions. Continuity of outcrop between the eastern and western belts cannot be established because of an extensive veneer of laterite and superficial deposits in the southern part of the area. However, there is sufficient borehole data to indicate that the volcanics are continuous below surface. The laterite is believed to have formed during periods of intense weathering in the Tertiary (Traves, 1955; Hays, 1967).

Fig 2

The formation is thickest in the East Kimberley region, where a maximum of 1 500 m is attained west of Halls Creek (Roberts, et al., 1968, 1972). Around the western margin of the Hardman Basin, total thicknesses range from about 470 m in the south (Gemuts & Smith, 1968) to about 960 m in the north. South of Wyndham the formation is up to about 1 250 m thick.

These thicknesses are much greater than those indicated by stratigraphic drilling for the eastern belt of volcanics (less than 250 m), and those measured south and north of Tanami (less than 30 m; Blake, in press; Hodgson, in press a, b).

Small outcrops of basic lavas and interbedded sediments mapped as Peaker Piker Volcanics (Smith ~~and~~ & Roberts, 1963), Colless Volcanics (~~Carter and Opik, 1963~~ (Carter, Brooks ~~and~~ & Walker, 1961), Helen Springs Volcanics (Randal ~~and~~ & Brown, 1969) and Nutwood Downs Volcanics (Dunn, 1963a) are exposed around the northern and western margins of the Georgina Basin and the southeastern margin of the Daly River Basin (Fig. 2). These formations may have originally formed part of the same extensive lava field as the Antrim Plateau Volcanics and have been separated from the main mass of volcanics by erosion and a cover of younger sedimentary rocks. The lavas are similar petrographically and chemically to Antrim Plateau lavas. The formations are unconformably or disconformably overlain by early Middle Cambrian sediments and are separated from underlying late Proterozoic rocks by marked angular unconformities, or, in the case of the Nutwood Downs Volcanics, are conformable on sandstone mapped as Lower Cambrian.

Table l

The Nutwood Downs Volcanics have a maximum recorded thickness of 122 m (Dunn, 1963a); the Helen Springs Volcanics, 37 m (Randal & Brown, 1969); the Peaker Piker Volcanics, 37 m (Smith & Roberts, 1963); and the Colless Volcanics, 61 m (Carter, Brooks ~~and~~ & Walker, 1961).

Borehole data indicate that lavas of the Antrim Plateau Volcanics underlie most of the Daly River and northern Wiso Basins (Fig. 2). Basic volcanics correlated with the Peaker Piker Volcanics have been intersected below early Middle Cambrian sediments in a stratigraphic hole, in mineral exploration holes, and in water-bores drilled in the northern part of the Georgina Basin (unpublished BMR Records 1970/114, 1972/74); but their subsurface extent is not known. Several wells and bores drilled in this part of the basin have bottomed in Precambrian basement rocks without intersecting basic volcanics. The eastern sequence of volcanics appears to have been relatively thin and it is possible that, in places, all traces of the volcanics were eroded before the deposition of the Middle Cambrian sediments and that the lavas are now preserved mainly as discontinuous sheets. Alternatively the lava flows may have originally been confined mainly to topographic depressions.

The Antrim Plateau Volcanics and their correlatives are exposed over an area of about 35 000 sq km (Fig.2) and extend over about a further 115 000 sq km beneath the sediments of the Hardman, Rosewood, Argyle, northern Wiso, and Daly River Basins. Small erosional remnants west of Halls Creek (Roberts et al., 1968, 1972), north and south of Tanami (Blake, in press; Hodgson, in press a, b), around the eastern margin of the Daly River Basin (Dunn, 1963a, b; Randal, 1963, 1969), south and north of Wyndham and in the central Victoria River region indicate that the area covered by lavas was formerly much more extensive. Lava flows also underlie parts of the northern Georgina Basin (Fig. 2). It is therefore likely that the lava field originally had an area of at least 300 000 - 400 000 sq km.

Small outcrops of tholeiitic basalt have also been described from around the margins of the Officer Basin (Peers, 1969; Lowry et al., 1972; Krieg et al., in press) several hundreds of kilometres south and south-southwest of the area investigated by the writer (Fig. 1). Basic volcanics have also been intersected in holes drilled in the Officer Basin, and seismic data suggest that they underlie most (at least 130 000 sq km, M.J. Jackson, pers. comm., 1975) of the basin (Lowry et al., 1972; Krieg et al., in press). The lavas are probably early Cambrian (Compston, 1974) and may also correlate with the Antrim Plateau Volcanics, which they resemble petrographically and chemically (unpublished data).

## Age

The age of the Antrim Plateau Volcanics must be inferred from the stratigraphic evidence, because no diagnostic fossils have been found in the interbedded sediments. The most definitive age limits are set in the East

Kimberley region where lavas unconformably overlie the Albert Edward Group (Dow ~~and~~ & Gemuts, ~~1967~~ 1969) and are overlain by the fossiliferous Blatchford Formation and Negri Group. Measurements of rubidium and strontium isotopic abundances in shales from the Albert Edward Group have yielded minimum ages of deposition of 666 ± 43 m.y. and 654 ± 48 m.y. for the sediments (Bofinger, 1967). Furthermore, the Albert Edward Group overlies the Duerdin Group, which contains ~~containing~~ glacial rocks deposited during a glacial epoch that affected widespread areas in Australia during the late Adelaidean (Dow, 1965). The Blatchford Formation contains the oldest faunal assemblage of the Ordian Stage (the oldest stage of the Middle Cambrian) and is slightly older than the Negri Group, the lower part of which is also Ordian (Öpik, 1967).

The available data, therefore, restrict the age of the Antrim Plateau Volcanics to between late Adelaidean and early Middle Cambrian. Because of the nature of the contacts with the overlying and underlying rocks, the volcanics are regarded as being little older than the overlying sediments and, therefore, most probably ~~early~~ Early Cambrian. However, the possibility that volcanism commenced in some areas in the late Adelaidean cannot be rejected on the available evidence. Walter (1972) suggested a late Precambrian, probably Vendian, age for silicified stromatolites from chert interbeds in Antrim Plateau Volcanics from the southern part of the area.

The Nutwood Downs Volcanics are conformably underlain by the Bukulara Sandstone, which rests on eroded Proterozoic sediments as a shallow-dipping capping (Dunn, 1963a). Sandstone 'dykes' in the basal lava flow (Dunn, 1963a) suggest it was extruded before the underlying Bukulara Sandstone was lithified. The Bukulara Sandstone contains <u>Scolithus</u>-like structures which are generally considered to be confined to the Phanerozoic (Plumb ~~and~~ & Derrick, in press).

Ages ranging from 395 ± 10 m.y. to 511 ± 12 m.y. have been obtained on 12 lava samples (10 from the Antrim Plateau Volcanics and one from each of the Helen Springs and Nutwood Downs Volcanics) submitted to the Australian Mineral Development Laboratories for isotopic age determination by the K-Ar whole-rock method. The most likely explanation for these anomalous Upper Cambrian to Siluro-Devonian ages is that the samples have lost varying amounts of radiogenic argon since their formation (A. Webb, in unpublished Australian Mineral Development Laboratories report).

ital

**Structure**

Fig 3

The lavas and interbedded sediments of the Antrim Plateau Volcanics are mainly flat-lying or gently dipping (~~<~~ less than $5^{\circ}$) and show little deformation, except in the western part of the area where they have been downfolded and cut by numerous faults and lineaments that trend predominantly northwest ~~-~~ to north-northwest~~-~~. ~~trending faults and lineaments.~~ The increase in the degree of deformation correlates with the proximity of the Halls Creek Mobile Zone (Dow ~~and~~ & Gemuts, 1969; Plumb ~~and~~ & Derrick, in press; Fig. 1). The volcanics and overlying Negri Group have been downwarped and, in places, downfaulted to form the markedly asymmetrical Hardman, Rosewood, and Argyle Basins (Matheson ~~and~~ & Teichert, 1948; Fig. 2). These basins probably formed as a result of isostatic movements and readjustment of basement blocks in the western part of the area after the outpouring of vast quantities of lava. Dips range from about $30^{\circ}$ in volcanics bordering the western margin of the Hardman Basin to subhorizonal in volcanics around the southern and eastern margins of the basin, farther away from the mobile zone. Parts of the southeastern margin of the Hardman Basin and of the northern margin of the Rosewood Basin are bounded by steep monoclinal flexures in the Headleys Limestone. The underlying volcanics responded to the stresses responsible for the downfolding of the Headleys Limestone by fracturing or movement along pre-existing fault planes. Displacements on most faults in the volcanics have been small (less than 100 m) and mainly vertical.

Fig 4

The Helen Springs Volcanics, in the east, have not undergone any major tectonic deformation. However, small domes ranging from about 0.5 km to 2.5 km across have been produced in the volcanics by the upwarping of underlying siltstones (Randal and Brown, 1969). The basal lava flow dips outward from the domes at angles ranging from $5^\circ$ to $30^\circ$. The Nutwood Downs Volcanics are flat-lying or gently dipping ($<5^\circ$) in a westerly direction. The Peaker Piker and Colless Volcanics appear flat-lying.

## FLOW CHARACTERISTICS

### Internal Fabric, Extent, and Thickness of Individual Flows

The general scarcity of phenocrysts, flow banding, and pronounced elongation of vesicles, indicate that most of the lavas of the Antrim Plateau Volcanics and correlative formations were highly fluid at the time of their extrusion and that they crystallized under relatively static conditions. The lavas were extruded subaerially: flow tops are generally highly oxidized and pillow lavas and palagonite (or altered palagonite) have not been definitely identified. It has not been possible to determine the extent of individual lava flows, or to correlate sequences from different parts of the lava field, because of the lack of continuous outcrop, the similarity in hand-specimen appearance of most lava flows, and the general absence of persistent marker units.

Secondary alteration, mainly as a result of low-grade burial metamorphism, but also partly due to weathering and probably deuteric or hydrothermal processes, has affected all the primary phases in most flows from the Antrim Plateau Volcanics and their correlatives to varying degrees. At the base of most flows is a thin, slightly to moderately vesicular or amygdaloidal, extensively altered chilled zone, generally less than 1 m thick. This zone grades upwards into relatively fresh, fine- to coarse-grained massive basalt making up the bulk of most flows and in which columnar jointing is rarely well developed. Higher up is a transitional zone of altered, slightly to moderately vesicular or amygdaloidal basalt, commonly characterized by spheroidal weathering. At the top of each flow is a highly vesicular or amygdaloidal zone of intensely altered basalt comprising about 15-30 per cent of the total flow thickness. The predominantly spheroidal or almond-shaped vesicles range in size from less than 1 mm to 20 mm across and are commonly filled with a variety of secondary minerals, the most common being quartz, prehnite, calcite, chalcedony, hematite, agate, pumpellyite, natrolite, analcime, and chlorite. The poorly exposed upper and lower surfaces of most flows appear to be planar; rubbly layers characteristically found in the upper and basal parts of aa and blocky lava flows (Macdonald, 1972) are rare. Individual thicknesses of 43 lava flows, penetrated in 9 stratigraphic holes drilled in essentially flat-lying sequences in the eastern and western belts of Antrim Plateau Volcanics, range from about 10 m to about 114 m, the average flow thickness being about 38 m.

Flow breccias or agglomerates (Traves, 1955; Sweet et al., 1974), or both, form extensive and prominent layers up to 40 m thick, at or near the top of the sequence in parts of the East Kimberley and Victoria River regions. The breccias consist of unsorted, closely packed, angular to subrounded fragments (up to about 60 cm across) of massive to highly vesicular basalt ~~(sensu lato)~~ set in a matrix of intensely altered, fine-grained volcanic detritus and, locally, minor red-brown siltstone and sandstone, confined to the upper parts of the units. Some of the breccias may be autobrecciated parts of lava flows (see Macdonald, 1972); in at least three places around the margins of the Hardman Basin, breccia grades downwards into massive non-brecciated basalt ~~(sensu lato)~~. Elsewhere, the breccias appear to form discrete units; they may have resulted from violent explosive activity.

this section to be printed in lower point setting (up to p.13)

Mineralogy

Massive basalts ~~(sensu lato)~~ from the Antrim Plateau Volcanics and correlative formations are typically pale to dark grey and fine- to medium-grained. The rocks commonly contain dark red-brown patches and veinlets rich in secondary iron oxides and small dark green clots of chlorite. A few flows are distinguished by small, prominent glomeroporphyritic aggregates of plagioclase phenocrysts.

The lavas are tholeiitic in character and consist~~ing~~ essentially of plagioclase, clinopyroxene, opaque minerals, and a mesostasis of red-brown to almost colourless devitrified glass or a quartzo-feldspathic residuum (e.g., see Wilkinson, 1967). Quartz, hornblende, biotite, and apatite are common accessories. A few flows contain scattered chlorite or 'iddingsite' pseudomorphs after olivine. Intersertal textures are the most common. Porphyritic textures are rarely well developed, the phenocryst content generally being less than 10 per cent. Subophitic and ophitic textures are poorly developed in coarse-grained rocks from the central and lower parts of a few thick flows.

Rapidly quenched upper and basal parts of flows intersected in stratigraphic holes consist predominantly of dark red-brown to almost black semi-opaque devitrified glass containing microlites of plagioclase, clinopyroxene, and opaque oxide, and, rarely, scarce euhedral laths (0.5 mm x 15 mm) of plagioclase and rectangular prisms (0.2 mm x 0.1 mm) of clinopyroxene. The devitrified glass is commonly partly replaced by chlorite, more rarely by prehnite, pumpellyite, calcite, and quartz.

In most massive basalts the groundmass laths and granules range from 0.1 mm to 0.3 mm in length. Most phenocrysts are less than 2 mm long. Phenocrysts rarely show flow alignment, and the groundmass grains are generally randomly orientated. Small pegmatoid veinlets are present in some samples. The well-formed constituent minerals are the same as those in the enclosing host rocks, but the pegmatoids are characterized by coarse grainsize and high contents (over ~~>~~50 per cent) of red-brown devitrified glass charged with opaque oxide granules.

Fresh and little-altered groundmass plagioclase grains in the massive basalts ~~(sensu lato)~~ are about sodic labradorite in composition ($An_{50-63}$; determined optically by the maximum extinction angles method).

In the highly crystalline rocks feldspar mainly forms small euhedral tabular grains showing normal zoning. In addition most massive basalts ~~(sensu lato)~~ contain minor amounts of euhedral to subhedral plagioclase phenocrysts, mainly as scattered ~~,~~ isolated laths, rarely as glomeroporphyritic aggregates. Some phenocrysts appear to be partly resorbed.

Plagioclase grains, particularly phenocrysts, are commonly partly replaced by sericite. In intensely altered rocks calcic plagioclase is extensively replaced by prehnite, pumpellyite, albite, and rare calcite.

Clinopyroxene forms mainly small ~~,~~ subhedral ~~,~~ blocky grains in the massive, highly crystalline basalts ~~(sensu lato)~~. Some medium- to coarse-grained rocks contain relatively large subhedral to anhedral grains (0.3 – 0.5 mm) and phenocrysts (0.5 – 2.0 mm) of clinopyroxene. The phenocrysts are commonly twinned and weakly zoned; many enclose fine submicroscopic exsolution lamellae parallel to the (001) plane of the host clinopyroxene. The proportion of clinopyroxene granules to feldspar laths decreases markedly in the fine-grained ~~,~~ rapidly chilled rocks, and in some rapidly quenched basalts ~~(sensu lato)~~ ~~,~~ clinopyroxene is absent.

Augite appears to be the most common pyroxene, but ~~However,~~ pigeonite ($2V = 0^{\circ}–15^{\circ}$) and ~~or~~ subcalcic augite ($2V = 15^{\circ}–30^{\circ}$), generally associated with a more calcic clinopyroxene (2V exceeds ~~$\not>$~~ $30^{\circ}$), have ~~has~~ been identified in many flows. The pyroxenes appear as colourless grains in thin sections and are indistinguishable in general appearance. Pigeonite, as small groundmass grains, is most abundant in the quartz-rich rocks. Massive basalts ~~(sensu lato)~~ (specimens 70770872A, 70770874, Table 2) from the interiors of two particularly thick (more than ~~$\not>$~~ 90 m) lava flows intersected in stratigraphic holes contain phenocrysts of pigeonite.

In most massive basalts ~~(sensu lato)~~ clinopyroxene shows only minor replacement by green or yellowish-green chlorite or, rarely, by pale yellowish-green to blue-green, slightly pleochroic, fibrous amphibole. In the more intensely altered amygdaloidal basalts ~~(sensu lato)~~ from the upper and basal parts of flows, however, clinopyroxene is commonly completely replaced by amorphous brown to black secondary iron oxides.

Magnetite and subordinate ilmenite form mainly small, subhedral to anhedral, interstitial grains in massive, highly crystalline rocks. Small euhedral and partly resorbed magnetite phenocrysts, commonly containing small inclusions of plagioclase and, rarely, clinopyroxene, are present in a few flows. Magnetite grains are commonly extensively replaced around boundaries and along fractures by secondary hematite and hydrated iron oxides (goethite or lepidocrosite, or both).

The bulk of the interstitial late-stage residuum crystallized as fine-grained, micrographic intergrowths of quartz and alkali feldspar in many of the coarser-grained, relatively slowly cooled basalts ~~[illegible]~~ from the central and lower parts of thick flows. Rarely, groundmass plagioclase grains are mantled by narrow overgrowths of alkali feldspar. Many of the highly crystalline basalts ~~[illegible]~~ from the interiors of thick flows also contain minor interstitial primary quartz and traces of primary poikilitic hornblende and biotite (Table 2). The hornblende is distinctly pleochroic from greenish yellow to pale pink and the biotite from dark brownish red to nearly colourless. Quartz, alkali feldspar, and primary hornblende and biotite have rarely been observed in flood-basalt flows, many of which are thinner and cooled more rapidly than the relatively thick lava flows of the Antrim Plateau Volcanics. Konda ~~and~~ & Green (1974; p. 1192) reported alkali feldspar, quartz, and amphibole in the groundmasses of flows of "basaltic andesite" and a dyke of "andesite or trachyandesite" from the Keeweenawan lava sequence of Minnesota. Tholeiitic dolerites commonly contain a quartzo-feldspathic mesostasis and minor primary amphibole and biotite (Wilkinson, 1967).

Minor olivine crystallized as euhedral to subhedral grains (0.1 mm x 0.05 mm) and small phenocrysts (0.4 mm x 0.3 mm) in only a few flows, and in the thin sections examined by the writer has been completely replaced by chlorite or "iddingsite" and opaque minerals. Glover (in Traves, 1955) described specimens of basalt from the Antrim Plateau Volcanics that contain abundant small phenocrysts of unaltered olivine.

Chemical composition

Table 2

Chemical and modal analyses and CIPW norms of 18 samples (selected from a total of about 55 analysed specimens) are listed in Table 2. Only the least-altered samples from the massive central parts of flows were analysed. On the whole the lavas are remarkably uniform in composition, and most oxides display only very limited ranges in composition (Table 2; Figs. 5a, b). For example, silica contents in the specimens analysed, ~~for example,~~ range from about 50 ~~per cent~~ to about 56 percent, nearly all ~~the vast majority of~~ values falling between 52 and 54 percent (Fig. 5a).

Fig 5

The majority of the lavas from the Antrim Plateau Volcanics and their correlatives are characterized by relatively high $SiO_2$ and MgO contents, and by low $TiO_2$ and total Fe (as FeO) contents compared with most flood-basalt sequences. The lavas contain normative hypersthene and many bear normative quartz; some have bulk chemical compositions more akin to andesites than basalts (Table 2). The lavas show marked enrichment in alkalies, particularly $K_2O$, and most ~~the majority~~ have alumina contents ranging between 13.5 and 15 percent. $Fe_2O_3$ contents range from 1.15 to 7.95 percent. The high $Fe_2O_3$ values and correspondingly low FeO contents are mainly the result of secondary oxidation and alteration. The rocks display a tholeiitic trend of moderate iron enrichment on an AFM diagram (Fig. 6), but because of their relatively high alkali contents the northern Australian lavas plot closer to the alkalis corner than many other tholeiitic suites. The wide scatter of points is probably mainly a reflection of partial alteration of some of the samples. The only specimen from the Colless Volcanics, for example, is fairly extensively altered and has an anomously high $K_2O$ content (Table 2).

Fig 6

$TiO_2$ contents in the Antrim Plateau Volcanics and their correlatives show a bimodal distribution with maxima between 0.8 and 0.9 percent and 1.2

and 1.3 per cent (Fig. 5b). These values are significantly lower than $TiO_2$ contents of most lavas analysed from most of the better known flood-basalt sequences such as the Deccan Traps and Columbia River Group (Sukheswala ~~and~~ & Poldervaart, 1958; Waters, 1961; Bose, 1972; Wright et al., 1973; Fig. 7). Fig 7 Karroo basalts from Lesotho are also characterized by relatively low $TiO_2$ contents (Cox ~~and~~ & Hornung, 1966), but $SiO_2$ and $K_2O$ contents in these basalts are generally lower than in most lavas from the Antrim Plateau Volcanics and correlatives.

Modal and normative mineralogy indicate that the northern Australian lavas range from olivine tholeiites to tholeiitic andesites, quartz tholeiite being by far the most abundant rock type. Using a modal classification individual lava flows commonly show a range in rock types suggesting that some differentiation occurred during consolidation of the flows. However, most of the modal variations can be attributed to differing degrees of crystallinity. Chemical analyses of massive basalts collected from different levels in several relatively thick flows show no significant variations in composition within the individual flows. Analyses of rocks from the chilled vesicular upper and basal parts of flows, however, reflect the changes imposed upon them as a result of redistribution of elements during secondary alteration and depart significantly from the original compositions.

## Alteration

The pattern of alteration in the lavas shows a persistent tendency for the degree of alteration to increase from the massive part of each flow to the highly amygdaloidal or vesicular top and base of each flow. Joints and fractures in the lavas and the vesicular parts of flows acted as channelways Table 3 for migrating connate and meteoric fluids which produced the sequence of secondary phases within the flows. The oxides which show the greatest susceptibility to metasomatic redistribution are $CaO$, $K_2O$, $Fe_2O_3$, $FeO$, $H_2O^+$, $H_2O^-$, and $Na_2O$ (Table 3).

Extensively altered amygdaloidal basalts from the basal and upper parts of most flows are significantly enriched in $K_2O$, $Na_2O$, $H_2O^+$, and $Fe_2O_3$ and depleted in $SiO_2$, $FeO$, and $CaO$ compared with relatively fresh, massive basalts from the central parts of the flows (Table 3). $MgO$ and $TiO_2$ are slightly higher in the altered rocks ~~and~~ but $Al_2O_3$ and total iron (as $FeO$) display little variation. Volatile contents are generally much higher in the altered amygdaloidal basalts. Similar trends have been described in tholeiitic basalts from the Portage Lake Series (Jolly & Smith, 1972) and in altered andesitic flows from a Cretaceous volcanic sequence, central Chile (Levi, 1969). Table 4

MODE AND LOCATION OF ERUPTION

The lavas are characterized by a general lack of recognizable eruptive centres. Small swarms of northwest-oriented dykes cut the volcanics around the south-southeastern margin of the Hardman Basin. Most of the dykes are completely brecciated and consist of angular fragments (up to 15 cm across) of intensely altered, mainly massive to slightly vesicular basalt in a finely comminuted, extensively chloritized basaltic matrix. Slickensides are common on the walls of the dykes. The dykes closely resemble the flow breccias exposed around the margin of the Hardman Basin. Only two, thin (less than 50 cm wide), poorly exposed dykes of extensively altered massive basalt have been found.

Flood basalts are generally regarded as having been erupted relatively quietly from extensive fissures developed in zones of crustal tension, (e.g., Washington, 1922; Gibson, 1966; Clifford, 1968; Choubey, 1971; Macdonald, 1972). In many other flood-basalt sequences, dykes or other structures indicating sites of former vents have not been reported from large areas of the lava fields (e.g., Butler ~~and~~ & Burbank, 1929; West, 1959; Swanson, 1967; Schmincke, 1967), and it is widely accepted that the featureless character of the final lava plains resulted from the burying of any vents by the superposition of thickening lava flows during the later stages of the eruption. The Antrim Plateau lavas mostly cover an area that has been relatively stable since at least the early Carpentarian (Plumb ~~and~~ & Derrick, in press), and there is little evidence (except for the few dykes) of any tensional features which may have acted as loci for feeders to the flows. The northern Australian lavas are thickest in and ~~adjacent to~~ near the Halls Creek Mobile Zone, but there is no evidence of major extrusion of lava along the length of the zone in the ~~Lower~~ Early Cambrian.

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Captions of figures 1 to 7

Fig. 1. Distribution of basic volcanics of probable early Cambrian age in northern and southwestern Australia.

Fig. 2. Distribution of Antrim Plateau Volcanics and correlative formations in northern Australia.

Fig. 3. Terraced topography developed on a succession of flat-lying lava flows of the Antrim Plateau Volcanics, southeastern margin of Hardman Basin.

Fig. 4. Moderately dipping lava flows of the Antrim Plateau Volcanics, southwestern margin of Hardman Basin (view looking south). The base of the formation is near the foot of the scarp on the right hand side of the photograph.

Fig. 5. Frequency distributions of $SiO_2$ (a) and $TiO_2$ (b) in lavas from the Antrim Plateau, Helen Springs, Nutwood Downs, Peaker Piker, and Colless Volcanics.

Fig. 6. AFM plot of chemical data from the Antrim Plateau, Helen Springs, Nutwood Downs, Peaker Piker, and Colless Volcanics. $A = Na_2O+K_2O$ $F = FeO+0.8998Fe_2O_3$, $M = MgO$.

Fig. 7. $MgO$ versus $TiO_2$ variation diagram for the Antrim Plateau lavas and correlative formations (symbols same as ~~used~~ in Fig. 6), showing the fields of variation delineated by Wright et al. (1973) for the Picture Gorge Basalt and 'lower' basalt (P) and Lower Yakima Basalt (Y) of the Columbia River Group. All the other chemical types defined by Wright et al. from the group are characterized by higher $TiO_2$ contents, and $MgO$ contents ranging from about 6.5 percent to about 2.5 percent.

41

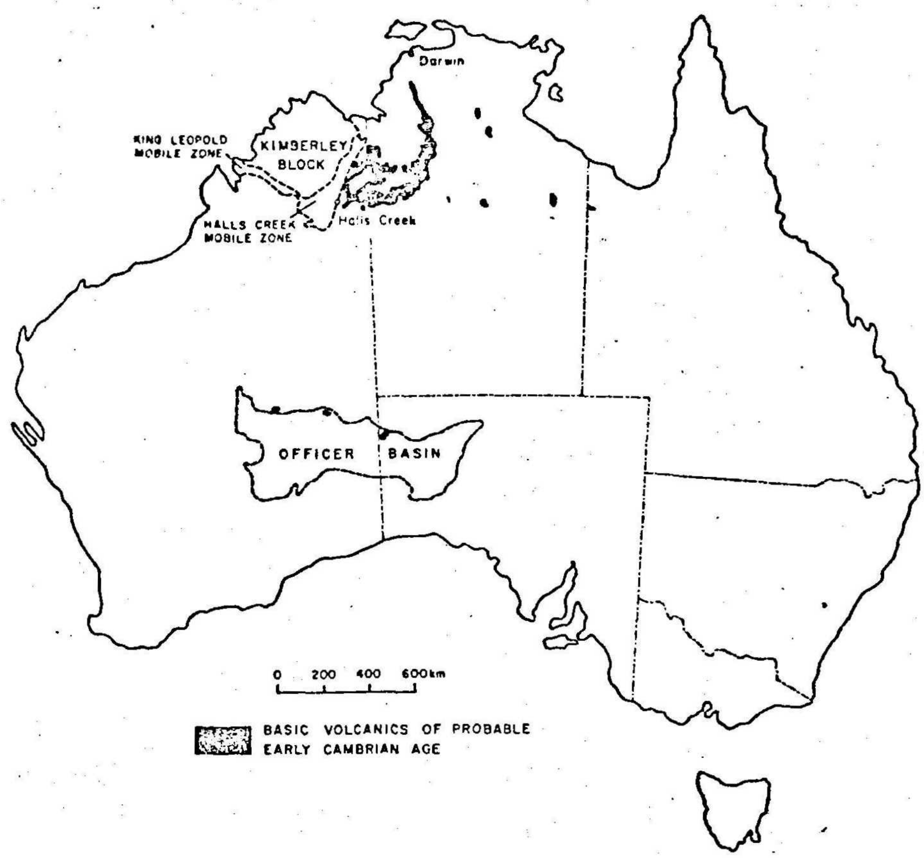

Darwin
KING LEOPOLD MOBILE ZONE
KIMBERLEY BLOCK
HALLS CREEK MOBILE ZONE
Halls Creek
OFFICER BASIN
0 200 400 600km
BASIC VOLCANICS OF PROBABLE EARLY CAMBRIAN AGE

Fig. 1

(#1)

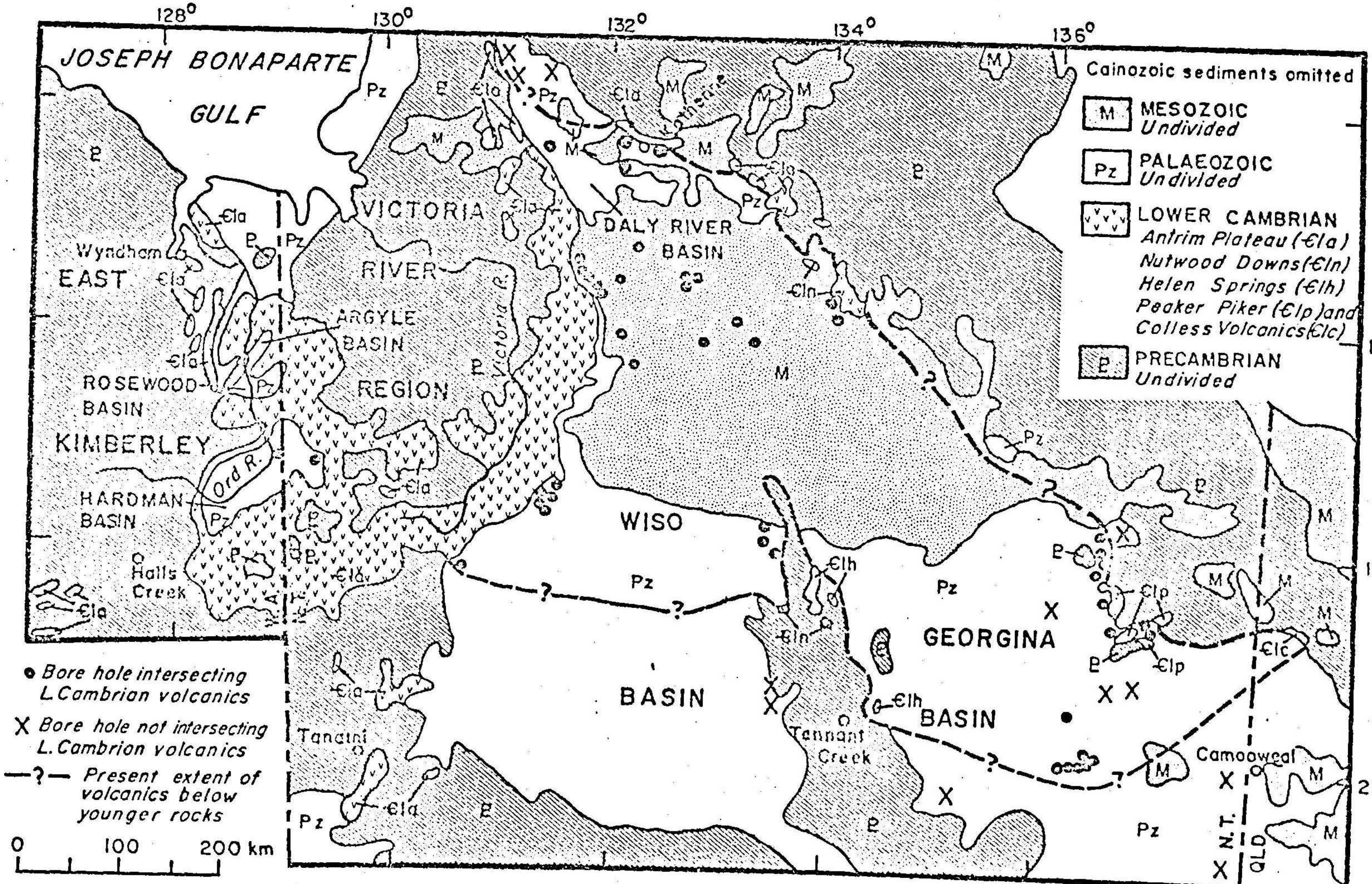



Fig.1 Distribution of Lower Cambrian volcanics in northern Australia

Fig 2

Fig. 3
Neg. No.
901/25

Fig. 4
Neg. No
1395/28.

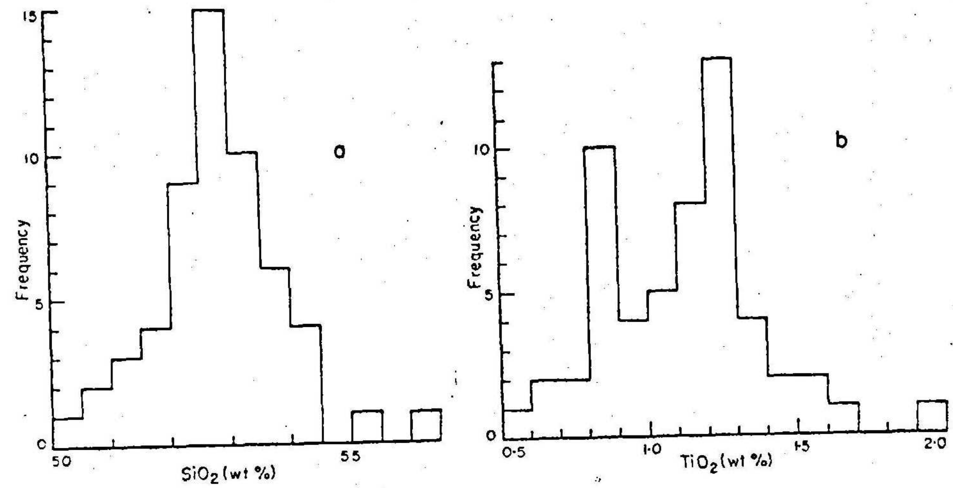



Fig 5

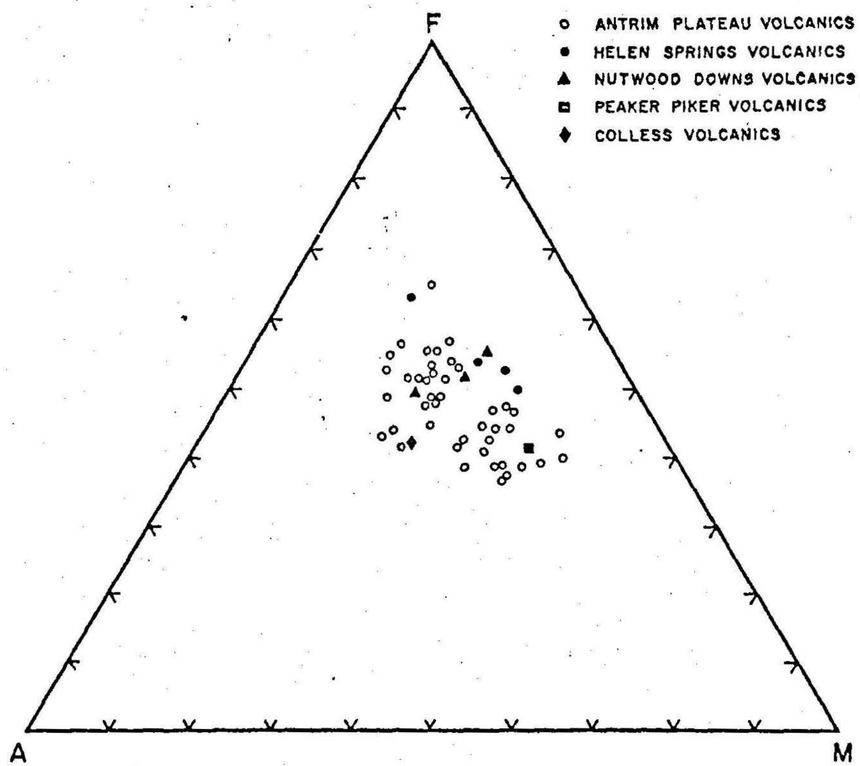
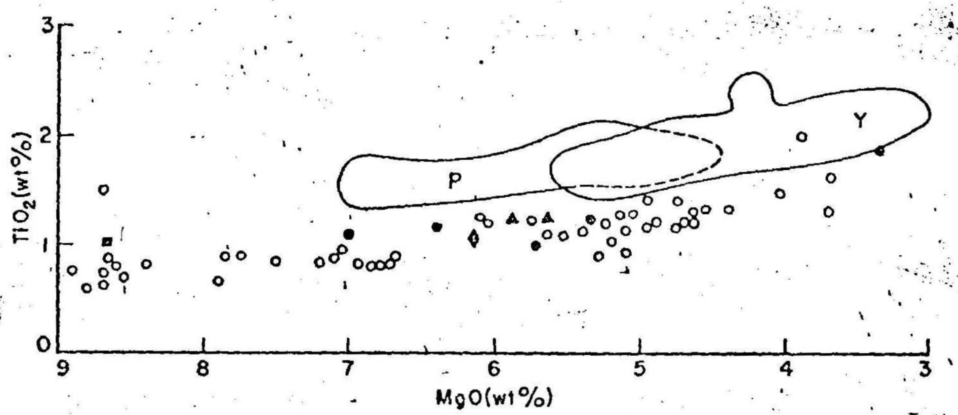

F
A
M
ANTRIM PLATEAU VOLCANICS
HELEN SPRINGS VOLCANICS
NUTWOOD DOWNS VOLCANICS
PEAKER PIKER VOLCANICS
COLLESS VOLCANICS

Fig 6




○ ANTRIM PLATEAU VOLCANICS
● HELEN SPRINGS VOLCANICS
▲ NUTWOOD DOWNS VOLCANICS
■ PEAKER PIKER VOLCANICS
◊ COLLESS VOLCANICS

Table 1. Comparison between Antrim Plateau Volcanics and other selected flood-basalt sequences

| nit | Age l.c. | Maximum Reported Thickness (metres) | Present-day Extent (sq km) | Reported Lithologies of the Igneous Rocks | Selected References |
|---|---|---|---|---|---|
| ntrim Plateau Volcanics northern Australia) | early Cambrian | 1 500+ | 150 000+ | tholeiitic basalt and **tholeiitic andesite** | **this paper** |
| olumbia River roup (USA) | Miocene-Pliocene | 3 000+ | 130 000 | tholeiitic basalt | Waters, 1961; Schmincke, 1967; Swanson, 1967; Snavely et al., 1973; Wright et al., 1973 |
| eccan Traps India) | late Cretaceous - early Tertiary | 3 000 | 510 000 | tholeiitic basalt; minor acid and alkaline lavas and intrusives | Sukheswala ~~and~~ & Poldervaart, 1958; Bose, 1972; Macdonald, 1972 |
| erra Geral Formation (South America) | early Cretaceous (mainly 120-126 m.y.) | 3 000+ | 1 200 000 | tholeiitic basalt and dolerite; minor acid and alkaline variants | Amaral et al., 1966; McDougall ~~and~~ & Rüegg, 1966; Macdonald, 1972 |
| oko basalts South-West Africa) | early Cretaceous | 900 | ? | tholeiitic basalt and dolerite | Siedner ~~and~~ & Miller, 1968 |
| rroo basalts southeastern Africa) | mainly late Triassic-early Jurassic | 9 000 | 140 000 | tholeiitic basalt; minor rhyolite, nephelinite, limburgite, teschenite, shoshonite | Cox, 1971, 1972 |
| rtage Lake Lava ries (southern Lake perior region, USA) | late Proterozoic | 5 000+ | ? | tholeiitic basalt; minor andesite and rhyolite | Butler ~~and~~ & Burbank, 1929; White, 1960; Jolly ~~and~~ & Smith, 1972 |
| rth Shore Volcanic oup (northern Lake perior region, USA) | late Proterozoic | 7 500 | ? | olivine tholeiite to quartz tholeiite; subordinate basaltic andesite, andesite, trachybasalt, latite and rhyolite | Konda ~~and~~ & Green, 1974 |
| ap rocks of the berian Platform (USSR) | late Carboniferous-early Triassic | 2 000-3 000 | 1 500 000 | basaltic lavas, tuffs, intrusives, minor alkaline variants | Nalivkin, 1973 |
| opermine River Group orthern Canada) | late Proterozoic | 4 300 | | tholeiitic basalt | Baragar, 1969 |
| lcanics of eastern eenland | Palaeocene-Eocene | 7 000 | 60 000 | mainly quartz-normative tholeiites; minor alkalic and nephelinitic types; rare tuff and rhyolite | Brooks, 1973; Fawcett et al., 1973 |
| lcanics of western eenland | Palaeocene-Eocene | 10 000 | ? | picrite, quartz and olivine normative tholeiites; minor trachyte, rhyolite and alkaline variants | Brooks, 1973; Fawcett et al., 1973 |
| teau basalts of land | Tertiary | 10 000 | ? | predominantly basaltic lava flows, minor rhyolite, andesite and intrusives | Einarsson, 1973 |

Table 2. Chemical and modal analyses of selected volcanic rocks from the Antrim Plateau Volcanics and correlative formations (sample localities are listed in Table 4)

| | 7077-2102 | 6977-0107 | 6977-0118 | 7077-0719 | *7077-6004 | 7077-2142 | 7077-2141 | 7077-0539 | 7077-0715 | 7077-2104 | 7077-2166 | *7077-3000 | 7077-2106 | 7077-0863 | 7077-0872A | 7077-0721 | 7077-0615 | 7077-0248 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 50.6 | 50.9 | 51.6 | 52.0 | 52.0 | 52.1 | 52.5 | 52.6 | 52.7 | 52.9 | 52.9 | 53.0 | 53.1 | 53.5 | 53.7 | 54.2 | 55.1 | 56.4 |
| $TiO_2$ | 0.60 | 0.76 | 2.05 | 0.84 | 1.10 | 1.09 | 1.00 | 0.81 | 0.95 | 0.84 | 1.25 | 1.01 | 1.09 | 0.90 | 0.84 | 1.27 | 0.94 | 1.64 |
| $Al_2O_3$ | 16.1 | 15.7 | 13.0 | 15.6 | 14.2 | 14.7 | 14.8 | 15.0 | 15.4 | 14.8 | 14.4 | 14.2 | 14.3 | 14.7 | 15.1 | 14.4 | 14.8 | 12.8 |
| $Fe_2O_3$ | 1.69 | 1.15 | 4.10 | 1.55 | 6.60 | 2.25 | 6.25 | 1.55 | 2.00 | 2.10 | 3.50 | 5.25 | 3.45 | 3.50 | 1.33 | 3.50 | 7.95 | 3.70 |
| FeO | 6.05 | 6.75 | 10.70 | 7.70 | 3.90 | 7.85 | 4.75 | 7.50 | 7.40 | 7.20 | 7.55 | 3.90 | 7.30 | 5.05 | 6.65 | 7.35 | 1.10 | 8.20 |
| MnO | 0.13 | 0.15 | 0.20 | 0.17 | 0.16 | 0.17 | 0.15 | 0.15 | 0.16 | 0.15 | 0.18 | 0.05 | 0.17 | 0.14 | 0.15 | 0.18 | 0.13 | 0.29 |
| MgO | 8.80 | 8.70 | 3.90 | 7.20 | 6.15 | 7.00 | 5.70 | 6.85 | 7.05 | 6.70 | 5.70 | 8.65 | 5.55 | 6.70 | 6.75 | 5.15 | 5.10 | 3.70 |
| CaO | 11.10 | 8.85 | 7.90 | 10.60 | 5.25 | 10.30 | 9.50 | 10.40 | 9.30 | 10.10 | 9.15 | 5.65 | 9.25 | 9.35 | 10.00 | 8.40 | 7.05 | 5.80 |
| $Na_2O$ | 2.25 | 3.30 | 2.50 | 2.20 | 3.20 | 2.10 | 2.55 | 2.35 | 3.05 | 2.50 | 2.20 | 2.10 | 2.70 | 2.35 | 2.20 | 2.40 | 3.00 | 2.45 |
| $K_2O$ | 0.36 | 1.24 | 1.43 | 0.98 | 4.30 | 0.81 | 0.88 | 0.88 | 1.35 | 1.10 | 1.12 | 1.41 | 1.28 | 1.24 | 1.01 | 1.50 | 2.10 | 2.80 |
| $P_2O_5$ | 0.08 | 0.07 | 0.16 | 0.12 | 0.11 | 0.11 | 0.11 | 0.10 | 0.11 | 0.11 | 0.13 | 0.11 | 0.10 | 0.11 | 0.11 | 0.16 | 0.13 | 0.15 |
| $H_2O^+$ | 0.84 | 1.87 | 1.63 | 0.61 | 1.97 | 0.68 | 0.37 | 0.54 | 0.27 | 0.44 | 0.92 | 2.60 | 0.55 | 0.81 | 0.51 | 0.75 | 1.60 | 1.53 |
| $H_2O^-$ | 1.26 | 0.31 | 0.59 | 0.58 | 0.77 | 0.90 | 1.13 | 0.86 | 0.47 | 1.01 | 0.81 | 4.10 | 0.78 | 1.40 | 1.17 | 0.33 | 0.58 | 0.45 |
| $CO_2$ | 0.12 | 0.03 | 0.05 | 0.05 | 0.05 | 0.08 | 0.08 | 0.11 | 0.01 | 0.16 | 0.08 | 0.30 | 0.08 | 0.13 | 0.16 | 0.11 | 0.29 | 0.07 |
| Total | 99.98 | 99.78 | 99.81 | 100.20 | 99.76 | 100.14 | 99.77 | 99.70 | 100.24 | 100.12 | 99.89 | 99.63 | 99.70 | 99.88 | 99.68 | 99.70 | 99.87 | 99.98 |
| $\frac{100\,Mg}{Mg + Fe^{++}}$ (mol.) | 72.1 | 69.7 | 38.2 | 56.1 | 59.5 | 61.5 | 55.5 | 61.9 | 63.0 | 62.4 | 55.1 | 71.1 | 55.0 | 66.4 | 64.4 | 53.3 | 60.0 | 42.9 |
| CIPW norms | | | | | | | | | | | | | | | | | | |
| Q | 0.3 | 0.0 | 6.8 | 1.8 | 0.0 | 4.5 | 5.1 | 3.4 | 0.0 | 3.5 | 7.3 | 5.6 | 4.6 | 6.4 | 6.1 | 8.3 | 7.2 | 11.8 |
| or | 2.2 | 7.5 | 8.7 | 5.9 | 26.3 | 4.9 | 5.3 | 5.3 | 8.0 | 6.6 | 6.8 | 9.0 | 7.7 | 7.5 | 6.1 | 9.0 | 12.8 | 16.9 |
| ab | 19.4 | 28.6 | 21.7 | 18.8 | 28.0 | 18.0 | 22.0 | 20.2 | 25.9 | 21.4 | 19.0 | 18.2 | 23.2 | 20.4 | 19.0 | 20.6 | 26.1 | 21.2 |
| an | 33.5 | 25.0 | 20.5 | 30.1 | 12.1 | 28.7 | 26.9 | 28.3 | 24.5 | 26.3 | 26.6 | 27.2 | 23.5 | 26.6 | 29.0 | 24.5 | 21.3 | 16.0 |
| ne | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 |
| di wo | 9.0 | 8.1 | 7.6 | 9.1 | 5.8 | 9.1 | 8.3 | 9.5 | 8.8 | 9.5 | 7.6 | 0.1 | 9.2 | 8.1 | 8.3 | 6.7 | 5.0 | 5.0 |
| di en | 5.9 | 5.1 | 3.1 | 5.3 | 3.3 | 5.3 | 4.5 | 5.5 | [illegible] | 5.6 | 4.1 | 0.1 | 4.9 | 5.2 | 5.0 | 3.6 | 2.9 | 2.2 |
| di fs | 2.4 | 2.5 | 4.6 | 3.5 | 2.2 | 3.4 | 3.6 | 3.7 | [illegible] | 3.5 | 3.2 | 0.0 | 4.0 | 2.4 | 3.0 | 3.0 | 1.8 | 2.7 |
| hy en | 16.5 | 3.7 | 6.9 | 12.8 | 1.6 | 12.4 | 10.0 | 11.9 | 11.1 | 11.3 | 10.3 | 23.1 | 9.2 | 12.0 | 12.2 | 9.5 | 10.1 | 7.2 |
| hy fs | 6.7 | 1.8 | 10.4 | 8.4 | 1.1 | 7.9 | 8.1 | 6.0 | 6.9 | 7.1 | 8.1 | 8.7 | 7.4 | 5.6 | 7.3 | 7.9 | 6.2 | 8.7 |
| ol fo | 0.0 | 9.4 | 0.0 | 0.0 | 7.7 | 0.0 | 0.0 | 0.0 | 0.9 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 |
| ol fa | 0.0 | 5.0 | 0.0 | 0.0 | 5.6 | 0.0 | 0.0 | 0.0 | 0.5 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 |
| mt | 2.5 | 1.7 | 5.3 | 2.3 | 3.9 | 3.3 | 3.7 | 2.3 | 2.9 | 3.1 | 4.1 | 3.9 | 3.8 | 3.6 | 2.0 | 4.1 | 3.6 | 4.7 |
| he | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 |
| il | 1.2 | 1.5 | 4.0 | 1.6 | 2.2 | 2.1 | 1.9 | 1.6 | 1.8 | 1.6 | 2.4 | 2.1 | 2.1 | 1.8 | 1.6 | 2.5 | 1.8 | 3.2 |
| ap | 0.2 | 0.2 | 0.4 | 0.3 | 0.3 | 0.3 | 0.3 | 0.2 | 0.3 | 0.3 | 0.3 | 0.3 | 0.2 | 0.3 | 0.3 | 0.4 | 0.3 | 0.4 |
| cc | 0.3 | 0.1 | 0.1 | 0.1 | 0.1 | 0.2 | 0.2 | 0.3 | 0.1 | 0.4 | 0.2 | 0.7 | 0.2 | 0.3 | 0.4 | 0.3 | 0.7 | 0.2 |
| $\frac{100\,an}{an + ab}$ | 63.3 | 46.6 | 48.7 | 61.6 | 30.2 | 61.0 | 55.0 | 58.3 | 48.6 | 55.1 | 58.4 | 58.6 | 50.3 | 56.6 | 60.4 | 54.3 | 45.0 | 43.1 |
| Modal Analyses (volume per cent) | | | | | | | | | | | | | | | | | | |
| Plagioclase | 40 | 38 | 31 | 33 | | 38 | 45 | 40 | 35 | 40 | 47 | | 34 | 36 | 41 | 33 | 36 | 31 |
| Clinopyroxene | 55 | 46 | 26 | 49 | | 32 | 37 | 45 | 49 | 42 | 32 | | 48 | 35 | 43 | 47 | 27 | 31 |
| Opaque Minerals | 1 | 1 | 7 | 3 | | 2 | 4 | 4 | 2 | 3 | 4 | | 5 | 4 | 3 | 8 | 4 | 6 |
| Olivine (altered) | - | <1 | <1 | - | | - | - | - | - | - | - | | - | - | - | - | - | - |
| Quartz | - | - | - | <1 | | - | - | 2 | - | 2 | - | | 3 | 2 | 3 | - | 1** | 6** |
| Mesostasis | 4 | 15 | 37 | 15 | | 28 | 14 | 10 | 14 | 13 | 17 | | 10 | 21 | 11 | 12 | 33 | 26 |
| Hornblende | - | - | - | <1 | | <1 | - | <1 | - | - | <1 | | <1 | - | <1 | - | - | - |
| Mica | - | - | - | <1 | | - | - | <1 | - | <1 | - | | <1 | - | - | - | - | - |

Specimens analysed at Australian Mineral Development Laboratories, Adelaide.

$Fe_2O_3$ contents adjusted in all normative calculations according to the formula: per cent $Fe_2O_3$ = per cent $TiO_2$ + 1.5; any "excess" $Fe_2O_3$ being recalculated as FeO (see Irvine and Baragar, 1971).

* These are the least-altered specimens of Peaker Piker ~~Volcanics~~ (70773000) and Colless (70776004) Volcanics collected; however, the thin sections and chemical analyses indicate that the rocks have been fairly extensively altered.

** Mainly secondary.

Table 3. Comparison of analyses of little-altered massive specimens from flow interiors and extensively altered amygdaloidal and vesicular specimens from the upper and basal parts of flows intersected in Stratigraphic Holes drilled in the Antrim Plateau Volcanics (sample localities are listed in Table 4).

| | Interior of flow | | | Upper part of flow | Basal part of flow | Central part of flow | Upper part of flow | Basal part of flow | Central part of flow | Upper part of flow |
|---|---|---|---|---|---|---|---|---|---|---|
| | 70770539 | 70772104 | 70770874 | 70773053 | 70770518 | 70770872A | 70773052 | 70772034 | 70776088 | 70776087 |
| $SiO_2$ | 52.6 | 52.9 | 53.0 | 51.5 | 51.3 | 53.7 | 50.9 | 48.8 | 51.5 | 48.4 |
| $TiO_2$ | 0.81 | 0.84 | 0.85 | 0.86 | 0.89 | 0.84 | 0.93 | 0.88 | 0.89 | 0.93 |
| $Al_2O_3$ | 15.0 | 14.8 | 14.9 | 14.0 | 15.2 | 15.1 | 14.7 | 14.9 | 15.4 | 14.6 |
| $Fe_2O_3$ | 1.55 | 2.10 | 3.35 | 7.55 | 3.70 | 1.33 | 8.75 | 5.15 | 3.55 | 5.35 |
| FeO | 7.50 | 7.20 | 5.95 | 1.97 | 5.70 | 6.65 | 1.88 | 2.4 | 4.75 | 5.05 |
| MnO | 0.15 | 0.16 | 0.14 | 0.09 | 0.15 | 0.15 | 0.12 | 0.09 | 0.12 | 0.07 |
| MgO | 6.85 | 6.70 | 6.75 | 8.10 | 7.05 | 6.75 | 8.05 | 7.75 | 6.75 | 10.2 |
| CaO | 10.40 | 10.10 | 9.80 | 3.00 | 7.65 | 10.00 | 2.70 | 5.9 | 9.8 | 1.50 |
| $Na_2O$ | 2.35 | 2.50 | 2.20 | 4.60 | 3.55 | 2.20 | 4.10 | 1.38 | 2.35 | 3.05 |
| $K_2O$ | 0.88 | 1.10 | 1.04 | 2.40 | 1.56 | 1.01 | 2.00 | 5.5 | 1.20 | 3.0 |
| $P_2O_5$ | 0.10 | 0.11 | 0.10 | 0.09 | 0.10 | 0.11 | 0.10 | 0.10 | 0.10 | 0.10 |
| $H_2O^+$ | 0.54 | 0.44 | 0.57 | 3.60 | 2.10 | 0.51 | 3.25 | 3.25 | 1.11 | 4.25 |
| $H_2O^-$ | 0.86 | 1.01 | 1.31 | 2.05 | 0.93 | 1.17 | 2.70 | 1.84 | 2.4 | 3.0 |
| $CO_2$ | 0.11 | 0.16 | 0.17 | 0.28 | 0.08 | 0.16 | 0.08 | 2.35 | 0.15 | 0.15 |
| Total | 99.70 | 100.12 | 100.13 | 100.09 | 99.96 | 99.68 | 100.26 | 100.29 | 100.07 | 99.65 |
| Total Fe (as FeO) | 8.90 | 9.09 | 8.96 | 8.76 | 9.03 | 7.85 | 9.75 | 7.04 | 7.95 | 9.86 |
| CIPW norms | | | | | | | | | | |
| Q | 3.4 | 3.5 | 7.0 | 0.0 | 0.0 | 6.1 | 2.3 | 2.4 | 5.12 | 0.0 |
| C | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 1.4 | 1.7 | 0.0 | 4.5 |
| or | 5.3 | 6.6 | 6.3 | 15.1 | 9.5 | 6.1 | 12.5 | 34.1 | 7.3 | 19.2 |
| ab | 20.2 | 21.4 | 18.9 | 41.2 | 31.0 | 19.0 | 35.8 | 12.3 | 20.6 | 27.9 |
| an | 28.3 | 26.3 | 28.2 | 11.1 | 21.6 | 28.0 | 13.0 | 14.5 | 28.9 | 6.3 |
| ne | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 |
| di wo | 9.5 | 9.5 | 8.2 | 0.9 | 6.8 | 8.3 | 0.0 | 0.0 | 8.3 | 0.0 |
| di en | 5.5 | 5.6 | 5.3 | 0.8 | 4.7 | 5.0 | 0.0 | 0.0 | 5.9 | 0.0 |
| di fs | 3.7 | 3.5 | 2.2 | 0.0 | 1.7 | 3.0 | 0.0 | 0.0 | 1.6 | 0.0 |
| hy en | 11.9 | 11.3 | 11.8 | 15.4 | 10.5 | 12.2 | 21.3 | 20.3 | 11.5 | 27.2 |
| hy fs | 8.0 | 7.1 | 4.9 | 0.0 | 3.7 | 7.3 | 0.0 | 0.0 | 3.1 | 3.7 |
| ol fo | 0.0 | 0.0 | 0.0 | 3.6 | 2.1 | 0.0 | 0.0 | 0.0 | 0.0 | [illegible] |

Table 4. Locality index of chemically analysed specimens from the Antrim Plateau, Helen Springs, Nutwood Downs, Peaker Piker, and Colless Volcanics

## Antrim Plateau Volcanics

| | |
|---|---|
| 69770107 | Hilltop at 129°30'24"E, 16°25'30"S |
| 69770118 | Hilltop at 131°30'18"E, 16°17'00"S |
| 70770248 | Lava flow overlain by Headleys Limestone at 129°17'42"E, 17°32'48"S |
| 70770518 | Cuttings from 177 - 178 m interval, flow no.3 (64-178 m), in stratigraphic hole, Limbunya No. 2, sited at 130°00'06"E, 17°52'12"S |
| 70770539 | Cuttings from 303 - 305 m interval, flow no.5 (195-305+m), in stratigraphic hole, Limbunya No. 1, sited at 129°22'36"E, 17°25'00"S |
| 70770615 | Basal flow at 128°06'24"E, 18°17'18"S |
| 70770715 | Second lowermost flow, western margin of Hardman Basin - at about 128°14'E, 17°39'S |
| 70770719 | Second lowermost flow, northwestern margin of Hardman Basin - at about 128°19'E, 17°22'S |
| 70770721 | Third lowermost flow, northwestern margin of Hardman Basin at about 128°19'E, 17°22'S |
| 70770863 | Core from depth of about 70 m, flow no.2 (46-93 m), in stratigraphic hole, Waterloo No. 1, at 129°05'06"E, 16°31'06"S |
| 70770872A | Core from depth of about 138 m, flow no.3 (64-178 m), in stratigraphic hole, Limbunya No. 2, at 130°00'06"E, 17°52'12"S |
| 70770874 | Core of massive basalt from depth of about 237 m, flow no.5 (195 - 305+m), in stratigraphic hole, Limbunya No. 1 |
| 70772034 | Cuttings from 128 - 130 m interval, flow no.4 (104 - 130m), in stratigraphic hole, Wave Hill No. 1, at 131°14'42"E, 17°21'48"S |
| 70772102 | Cuttings from 119 - 122 m interval, flow no.2 (62 - 148 m) in stratigraphic hole, Waterloo No.2, at 129°24'24"E, 16°25'00"S |
| 70772104 | Cuttings from 273 - 274 m interval, flow no.5 (195 - 305+m), in stratigraphic hole, Limbunya No. 1 |
| 70772106 | Cuttings from 50 - 52 m interval, flow no.1 (0-56 m), in stratigraphic hole, Waterloo No. 2 |
| 70773052 | Cuttings from 66-67 m interval, flow no.3 (64-178 m), in stratigraphic hole, Limbunya No. 2 |
| 70773053 | Cuttings from 198-200 m interval, flow no.5, in stratigraphic hole, Limbunya No. 1 |

Continued

(Table 4 continued)

| | |
|---|---|
| 70776087 | Cuttings from 105-107 m interval, flow no.4 (104 - 130 m), in stratigraphic hole, Wave Hill No. 1 |
| 70776088 | Cuttings from 63-64 m interval, flow no.2 (62 - 148 m), in stratigraphic hole, Waterloo No. 2 |
| | <u>Helen Springs Volcanics</u> |
| 70772141 | Massive basalt in hillside at about 133°54'06"E, 18°26'06"S |
| 70772142 | Small gully at about 133°52'42"E, 18°26'42"S |
| | <u>Nutwood Downs Volcanics</u> |
| 70772166 | Lava flow at about 134°04'42"E, 15°36'54"S (Flicks Waterhole) |
| | <u>Peaker Piker Volcanics</u> |
| 70773000 | Core sample from about 149 - 150 m interval in stratigraphic hole, Alroy No. 2, at 136°17'30"E, 19°30'50"S |
| | <u>Colless Volcanics</u> |
| 70776004 | Floater of massive basalt at about 138°19'E, 18°39'S |

Conti

ERUPTIVE HISTORY OF MANAM VOLCANO, PAPUA NEW GUINEA

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ABSTRACT

Manam, an imposing basaltic stratovolcano rising 1725 m~~etres~~ above the sea, lies 12 km off the north coast of mainland Papua New Guinea. It was probably first recognized as an active volcano~~e~~ during the mid-sixteenth century, but until the 1870s recorded sightings of the island were very infrequent; since then probably all the major eruptions have been recorded. Manam is clearly one of the most active volcanoes in the country and after it was brought under constant surveillance in 1957, it became apparent that the volcano is more-or-less constantly active. A compilation of all available accounts of past activity shows that the longest period of quiescence during the last 100 years is 9 years. A few large eruptions have occurred at the main crater but most strong activity takes place at the southern crater. The larger eruptions are typically of powerful strombolian character, with frequent but usually brief lava flows and occasional nuées ardentes of basaltic composition. Heavy falls of ash and scoria result from paroxysms, typically of only a few hours duration, but often recurring after a few days to a few weeks.

INTRODUCTION

Manam, 12 km from the north coast of mainland Papua New Guinea (Fig. 1), is one of the most active volcanoes in the country (Fisher, 1957). Frequent strombolian eruptions have been recorded, many involving lava flows, and the volcano's longest repose period during the last one hundred years appears to have been only about nine years. The occurrence of nuées ardentes during some of its eruptions is well established (Taylor, 1963). Lavas produced during recent eruptions have been olivine-bearing quartz tholeiite basalts and low-silica andesites (Table 1). Some 5000 people live in a dozen or so village groups around the coast, and at least some of these people are in considerable danger during the most violent eruptions.

Fig. 1

Full-time volcanological studies commenced in Papua New Guinea with the establishment of a volcanological observatory in Rabaul in 1938. The observatory was re-established in ~~1951~~ 1950 after a break of ten years during and following the Second World War. Volcanological observations commenced at Manam in June 1957 when, following a period of activity earlier in the year, an observation post was established by G.A.M. Taylor. Almost continuous observations were made by Taylor from this and from another, temporary, post during the violent eruptions of 1957 to 1960. A well-equipped permanent volcanological observatory, planned by Taylor, was established on the island in 1966 (Fig. 2)*.

This paper surveys the known eruptive history of Manam, also referred to in earlier years as Manumudar, Mammamur, Manamur, Vulcan Island or Hansa Island. The compilation of this account has involved the searching of explorers' journals dating from before the time of European settlement

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* See Myers (this volume). Ed.

(1870s), of explorers', naturalists' and administrators' accounts from the period of German administration (till 1914) and from various (mostly non-scientific) sources from the period of Australian administration, up until 1951. Many of the older German texts are not generally available in Australia and the descriptions of eruptions at Manam have been translated and included in the following account. To give a balanced picture of activity, descriptions (mostly originally in English) of activity from the period 1914 to 1951, have also been included although these texts are generally more accessible. Until the 1870s, observations had been possible at only very long intervals, and their main value lies in their indication of the character of activity, rather than as contributions to a connected eruptive history. Since then, observations have been intermittent, although it is probable that at least the major events have been reported. In recent years it has become clear that eruptive activity of some sort is more-or-less continuous. Much of the activity consists only of minor ejections of dust, occasional weak glows, and minor rumblings. However, periods of vigorous strombolian eruption occur at intervals, and, for the period since the 1950s, only activity of this type is covered in this paper. A detailed study of the activity since 1957 is in progress, and will be reported elsewhere.

TOPOGRAPHY

Manam is an impressive island stratovolcano, with a basal diameter of about 25 km, and a height of about 2800 m (1725 m above sea-level). It appears to be very symmetrical at a distance, but a closer view shows that its flanks are strongly dissected and that it has a complex inner "core" Fig. 3). The lower parts of the island, to about 1250 m above sea level, are at the planeze stage of erosion (Kear, 1957), four planezes of similar dimensions being separated by four major radial, amphitheatre-headed valleys in the northeast, southeast, southwest, and northwest quarters (Fig. 2). Rising above this level is the central eroded core buttressed by a dense network of flows and dykes as seen exposed in the rugged headwalls and upper sidewalls of the valleys.

Fig 2

Recent activity has occurred chiefly at two craters: the main crater, at the head of the northeast valley, and partly breached towards the northeast, and the southern crater, perched high near the summit at the head of the southeast valley. The position of the southern crater seems to shift about within a small area from eruption to eruption, and sometimes there may be two or three active vents in the vicinity. Although there are four parasitic cones (Fig. 2), historical activity is known only from the summit.

Recent activity is rapidly building up an apron of ejecta about the flanks of the central core. The apron is most strongly developed in the northwest and southeast valleys, less so in the northeast valley, and hardly at all in the southwest valley (Fig. 3). The apron consists of large talus cones built up against the valley headwalls; lava flows are incorporated in the talus in both the northeast and southeast valleys, at least. Nowhere has the apron yet built up sufficiently to overflow onto the outer planezes of the island, although eruptions in the late 1950's filled in the head of the southeast valley (Taylor, 1963).

The amount of apron material present can be accounted for by comparatively recent activity (perhaps 500-1000 years). Imaginary removal of this young material would reveal the deeply eroded skeletal core of the island, and studies of erosion rates on shield and stratovolcanoes (McDougall, 1964, 1971; Ruxton and McDougall, 1967) suggest that this degree of dissection may have needed at least tens of thousands of years to develop. Detailed geological studies of Manam are in progress, but as yet there is no stratigraphic history available to determine if the topographic features result from two or more separate periods of volcanic activity.

## EARLY EUROPEAN OBSERVATIONS

Manam may have been first observed by European navigators, and identified as an active volcano, during the mid-sixteenth century. A volcano was observed in this area during the voyage of de Retes in 1545 (Wichmann, 1909), although its identity cannot be positively established. Certainly, a volcano is marked on a European map of 1601 (Sharp, 1960, Plate 1) in this vicinity; possibly it really lies among the islands just south of Manus Island (300 km to the northeast.).

Schouten and le Maire observed Manam in eruption on 6-7 July 1616 (Spilbergen, 1906); their map is detailed enough to allow positive identification of the volcano, which they describe as emitting flames and smoke.

Tasman also saw Manam in eruption on 21-22 April, 1643; it "burned steadily directly out of the top with flames of fire" (Sharp, 1968). Again, the volcano can be positively identified from contempory maps of the region. Tasman made a sketch of an erupting volcano, but its identity is not clear, as he also saw Karkar (Fig. 1) erupting at about the same time.

Dampier observed Manam in 1700: "On Tuesday the 2d of April, about 8 in the Morning, we discovered a high peeked Island to the Westward, which seem'd to smoak at its Top. The next Day we past by the North-side of the Burning Island, and saw a Smoak again at its Top; but the Vent lying on the South-side of the Peek we could not observe it distinctly, nor see the Fire." Dampier's map and sketch of the volcano both unmistakably indicate Manam (~~Dampier,~~ Williamson, 1939).

In a fanciful account of his voyage along the coast near Manam in 1830, Morell (1832) claimed to have seen seven active volcanoes, five of which were actually erupting, and the other two smoking. Six of these were islands, and the other was said to be in the coastal ranges of the mainland, where no youthful volcanoes are now known; only four recently active volcanic islands are known in the area described by Morell. In spite of the doubt about the veracity of the account, circumstantial detail of the only eruption actually described seems to suggest Manam: "In the direction of north-north-east from this cape [stated to be at 4°59'S, 145°16'E; that is 50 km inland from the present coast] is a small volcanic island, lying about six leagues from the mainland, which was in full blaze... the flames ascended upwards from the lofty summit of the isolated mountain, at least one thousand feet; while the red burning coals of pumice-stone were carried to the north-west..."

OBSERVATIONS SINCE THE 1870s

1877. An eruption was observed between 11 and 13 November 1877 by von Miklucho-Maclay (1878) and was first seen from the vicinity of Karkar. "During the whole night one could watch the flickering of the red light". "The fire occurred at intervals of several minutes and lasted each time for only about ½ to 2 minutes". Early on 12 November, "the mountain-top was covered with a smoke-and-cloud cap above which a high, nearly vertical column of smoke towered up" and in the afternoon, from a distance of about 60 km, von Miklucho-Maclay saw "three different columns of smoke which rose from the crater and, as its southerm rim was a little lower, one could see the new eruption cone out of which rose two strong columns of smoke, while the third one, being a bit lower, rose at the southwest rim of the crater. From time to time great quantities of white smoke emerged. They were thrown out at similar intervals as the flickering of the light during the previous night had appeared. The three columns of smoke combined to form a huge whitish-grey cumulo-stratus cloud. As it grew darker I saw, several times, mighty fork-lightnings flash through the darkening clouds. The white smoke clouds, periodically thrown out, changed into columns of fire which flared up in the same way as they had done the day before." In a footnote von Miklucho-Maclay stated that the eruption had been noted in the logbook of the schooner "Flower of Jarrow" on 29th October, 1877.

Insert →

1885. Finsch (1888) observed the island from a distance: "A rosy glow surrounded the crater wall and in the collapsed western vent could be seen clearly a fiery glow like a burning chasm which appeared more intense after nightfall, until the mountain was, stage by stage again covered by cloud". Sapper (1917) noted that this observation was made during May 1885. Later in his account, ~~Furthermore,~~ though probably referring to the same eruption, he described the "mighty roar" and the "enormous white masses of cloud" emanating from the volcano.

1887-88. Grabowsky (1895) described an apparently more violent eruption in June 1887 which was clearly visible from Hatzfeldhaven, 35 km distant, on the mainland (Fig. 1). "Weak tremors were observed several times while the double craters (Auroka and Ujap) of the volcano Manamur ... were continuously active, and in the night from 27 to 28 June 1887 a violent eruption with a large lava flow took place ..."

Similar activity was seen by Zöller (1891) who was stationed at Hatzfeldhaven in 1888. "During the day one sees only a column of smoke whilst in the evening the gleam of the embers from the outbreaks of flames which are repeated at intervals of a few seconds gives much more variety to the scene. At times one saw molten lava flowing down to half way down the mountain." Sapper (1917) made the comment "apparently glowing ash", in relation to the lava flow. This is possibly correct as from the distance of Hatzfeldhaven it might be difficult to distinguish lava flows from hot ash avalanches or nuées ardentes.

1889-95. Sapper (1917) quoting Schleinitz (1889) stated that Manam was active during 1889. However, Schleinitz only remarked generally on activity.

Zdekauer (1899) is quoted by Sapper (1917), who gave his source as Hammer (1907), as having "observed sparks of fire and lava flows" sometime during 1889. However, Hammer stated that Zdekauer "noted lava producing eruptions which occurred every eight minutes". As Zdekauer's book (Zdekauer, 1899) was not seen by us, neither the reliability of the observation nor the date of the eruption could be checked.

According to Tappenbeck (1901) activity continued at Manam until about 1895: "Until the mid 1890's the crater of Vulcan Island was highly active and ejected lava almost continuously". Subsequently: "Vulcan crater quietened down and shown no sign of its inner activity for long periods of time". Tappenbeck equated the change in activity with the eruption of Karkar in 1895.

1898-1900. No further activity was reported until some time between 1898 and 1900 when Pflüger (1901) noted that "The summit is jagged and produces two mighty columns of smoke".* von Hesse-Wartegg (1902) also noted the twin smoke columns. He further stated that "on the northern slopes can be seen the flow of the fresh lava streams which stem from the latest eruption".*

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* Sapper (1917) gave the dates of these observations as the publication dates of each book. However, it is possible that both accounts are of the same event.

1904. Pöch (1908) made the following observations from Potsdamhafen (on the mainland opposite Manam; now abandoned) on volcanic activity at Manam during October 1904; "On 26 October, at 2 p.m., a very high column of smoke suddenly appeared above the summit of the volcano". "At about 4 p.m. the cloud became larger and darker and one could see a shower of ash falling into the sea to the west of the island". "After sunset a powerful column of fire could be seen at the summit". "One could hear rumblings and rifle-shot-like detonations. I could quite clearly see through my binoculars the fire column which issued from the westerly vent from time to time, always to be followed by a detonation. From the second and easterly crater vent the same weak but steady smoke column persisted just as before the eruption. This vent, though very close to the westerly one, was not at all affected by the eruption". "I could further see through the binoculars that large glowing boulders fell to the west of the fire column, and a lava stream rolled down the western valley. The lava had set fire to a forest and one could see burning trees or the glowing, burnt-out trunks. The lava reached the foot of the mountain and came to rest where the angle of slope diminished; it did not reach the sea. It was said that villages were destroyed. Next morning, 27 October, tremendous explosions occurred, but at 11 a.m. activity ceased. The westerly vent then smoked less than before the eruption. There were no more eruptions or tremors as long as I was at Potsdamhafen (till 25 November)". An illustration of the volcano by Pöch on 26 October shows a tall ash column rising to 2000 m or more above the summit. Two other, much shorter accounts, of this eruption (Pöch, 1907 a & b) state that the lava flow reached the sea. Pöch was told that eruptions had occurred at Manam regularly in the past and that the last eruption prior to his visit had been during April or May 1904.

1909. Under Sapper's entry for 1909 is, "strong smoke development"; the source is given as Friederici ms. This reference was not seen by us so it could not be checked. G. Häntke (pers. comm., 1974) noted that "explosions and lava" occurred during 1909; no source ~~is~~ was given.

1911. Manam was visited briefly in December 1911 by Scholz (1912) who made the following observation. "On the evening of my arrival (12 December) it was active and ejected smoke and flaming fire. Every now and then there occurred deep rumblings inside the mountain and once a strong and persistant earth tremor was felt. Wide furrows on the slopes mark the the paths of lava flows towards the sea. In one particular place, about a year ago, the volume of lava built up a kind of knoll".

1913. Behrmann (1917), on his return from a voyage up the Sepik River in September 1913, noted that "the easternmost crater ejected a brown smoke column. About every 10 minutes an eruption occurred concurrently with an underground thunder, the ejected material falling back into the crater. At night the reflection of fiery lava masses could be seen". He stated that there had been several eruptions during the course of his expedition.

1917. A brief explosive eruption following a strong earthquake occurred sometime during 1917. Stanley (1923) stated that the eruption lasted a few minutes only and that "boulders 3 or 4 cubic yards in mass were hurled to a height of 2000 feet".

A much more violent eruption during 1917 is implied by a report in

al. the mission journal, Steyler Missionsbote (1921-22). There, the quite violent eruption of December 1920 (see below) is described and compared in magnitude with one which took place during 1917.

<u>1919.</u> The eruption of 11 August 1919 was one of the most violent ever reported; Stanley (1923) gave the following account derived from a missionary who observed it. The eruption was preceded by several small earthquakes. "At about 2 p.m. large volumes of steam and black vapour rising in the form of a vortex formed about the apex and soon obscured the sun, but the glow from the crater illuminated the lower clouds. A spasmodic rumbling at the base of the mountain commenced, and soon after 2 o'clock lava commenced to flow down the Eastern side and in about 5 minutes reached the sea near (B) causing the water to boil furiously, and huge volumes of steam and black vapour to rise to a great height, which covered the whole mountain and surrounding country". "At 2.55 p.m. the red steam of lava could be clearly seen, but ceased to flow, and dense grey-coloured clouds of vapour were emitted from the crater. When the rumbling ceased a grey-brown halo of dust encircled the mountain, which gradually spread for miles over the mainland". "At 3 p.m. the South-east wind cleared the Western centre of the island, but the lower Western portion was black with dense vapour and dust clouds. At 3.3 p.m. the largest crater was alone in eruption, the lava continuing to flow from the Eastern edge for a short distance. At 3.45 p.m. the lava ceased to flow, and only a small quantity of black vapour rose from the crater. The villages of Zogari and Josa [on the west coast of the island] were covered with dust, scoria, and hot fragments of vesicular lava". Further on in his account Stanley noted: "At night, it was possible to read a newspaper 8 miles away by the reflection of the lava on the clouds". The point (B) mentioned is shown on Stanley's sketch map at the foot of the southeast valley.

1920-22. A slightly less violent eruption commenced in December 1920. A report in *Steyler Missionsbote* (1921-22) dated 19 December 1920 states; "Opposite our station (Monumbo, near Potsdamhafen) lies the fire-ejecting mountain Manam." "For a fortnight, it has been ejecting a fiery column many metres high above which hang black smoke clouds. Out of the clouds, twice, ash has rained for many hours. Other times it ejects glowing lava out of its fiery vent which rolls, destroying everything, down the mountainside and into the sea". Another description is given in a further passage. "Day and night one can hear for many hours its underground rumblings. It ejects glowing lava house-high which rolled into the sea as a fiery stream".

The eruption probably continued until at least March 1921 as Stanley (1923) recorded an eruption at that time "damaging a few gardens and extruding some lava, but it was not as violent as that in 1919". Stanley (1923) also reported that "the crater was in partial eruption, emitting huge volumes of steam and black dusty vapours which darkened the whole of the North-western portions of the island" during his visit in 1922 (the precise date is not reported).

1925. G.A.M. Taylor (unpublished ms) recorded that an eruption took place during 1925. He was informed by a local resident that it was possible to read a newspaper at Dugomur plantation on the mainland opposite the island by the night-time glow from incandescent lava being ejected from the volcano.

This eruption is not recorded by either the *Official Handbook of the Territory of New Guinea* (published 1936), or the *Pacific Islands Year Book 1935*, both of which state that the last big eruption was in March 1921. However, Höltker (1942), drawing on Mission archives, stated that between 1917 and 1926 six large eruptions were observed, one of which presumably was that mentioned by Taylor.

1926-28. G. Häntke (pers. comm., 1974) ~~says~~ ~~states~~ informed us that strombolian activity occurred between March 1926 and February or March 1928. However, the original source of the information is not known.

1933-34. The anthropologist, Camilla Wedgwood was on Manam between January 1933 and February 1934. She wrote (Wedgwood, 1934): "it is not uncommon to hear the sound of volcanic explosions and to see a red glow at night in the sky above the mountain top. Sometimes too, a thick brown-black cloud of dust rolls down over the upper slopes, and during the latter part of my stay there was a very brief shower of fine, glowing cinders". Höltker (1942) also mentioned that he has photographs showing smoke plumes in 1932 and 1934.

1936-39. Höltker (1942), an anthropologist, who was stationed on Manam Island between July 1936 and April 1939, stated that, between the beginning of October and the middle of December 1936, "the volcano was subjected to numerous large paroxysms. Six large eruptions followed one another at intervals of five to seven days". He stressed that "Mostly the eruptions began towards evening, and lasted with only short interruptions into the early hours of the morning. On only one occasion did an eruption take place in the morning, and this was on 11 October 1936". He described their general pattern as follows: "In the late afternoon the usual smoke plume became darker and denser. One could notice a peculiar unrest in the mountain; earthquakes of different strengths and underground rolling. One could already see in the sky a red glow above the crater (the east crater was active). The rolling became stronger and stronger until the first eruption began as projections of glowing ejecta rose up hundreds of metres, and one could distinguish plainly large glowing blocks of lava. These glowing sheaves now played up and down like a fountain, as if the driving force were sometimes stronger, sometimes weaker. In between, the whole sky was covered with heavy black clouds of ash. Such an outburst lasted for several minutes, and sometimes up to a quarter of an hour. Then the glowing fountain sank back beyond the crater rim, the glow of the lava above the open crater remained, and the smoke clouds rose higher until after a short time fresh rolling in the interior of the mountain heralded a new outburst. Each outburst followed the previous one fairly frequently in the first hours of the night, at intervals of about 10 to 15 minutes. Towards morning the pauses became longer until after a last big outburst the volcano again became quiet, and only the usual smoke plume remained above the summit". Then followed "a pause of 6 to 7 days in which, throughout the day, the usual smoke plume, and at night the glow of the lava, could be seen, but no eruption followed".

Höltker did not describe a nuée ardente, but in one of his published photographs, taken on 11 October 1936 he identified, "a glowing cloud descending" the southeastern valley. This identification appears to be accurate although details are not particularly clear in the photograph. On the same day: "Out of the clouds there fell on us, not very thickly but yet continuously, ash and little glowing pieces of scoria". "From time to time small and large lapilli poured down onto the corrugated iron roof of our house". "After every outburst there was a layer of ash on the leaves of the taro plants in our garden, up to a centimetre thick, and more".

A report by the captain of a vessel (data file, Rabaul Observatory) states that a "heavy lava flow" was observed at 7 p.m. on 18 October 1936, ceasing at 8.20 p.m., and that "terrific eruptions" occurred between 6.30 p.m. and 10 p.m. on 24 October, accompanied by lava and by the ejection of incandescent blocks 1000 to 1200 metres into the air.

Fisher (1939) gave dates for peaks in the 1936 activity at 17, 25, and 31 October and stated that "A similar outbreak occurred on 15 March 1937". Höltker (1942), however, noted that, after a lull from December 1936, eruptions began again in May to June 1937; the renewed activity taking the form of quiet outpourings of lava. He noted that "After a couple of months it [the lava ] could be observed from Bogia Harbour on the mainland on almost every evening; sometimes it was stronger, sometimes weaker". "During these fiery outpourings glowing fragments bounced and rumbled down the hill; these were probably the larger lava blocks". "This activity on Manam, with outpourings of lava at night, continued almost unbroken during the following years to the end of my stay in New Guinea" [April 1939]. Höltker further stated that two craters were active during his stay on the island. From the descriptions, these can be readily identified with the present main and southern craters. The main crater was active in 1936, and the southern crater was "constantly and intensely active".

Another report concerning activity in January and February 1938 (Rabaul Times, 1 April 1938) notes that "the crater on the south side, which is beneath the main one has been very active, shooting lava into the air hundreds of feet", and that the volcano had been in eruption "night and day" for some time.

A correspondent to the Pacific Islands Monthly of June 1938 described activity at an unknown date early in 1938. "The occasional wisps of steam and smoke normally seen issuing from the crater have given place to an almost constant eruption-cloud, ever-changing in shape and size with the wind and intensity of the eruption. Down the southern side of the cone may frequently be seen a sloping streak of steam, rising from flowing lava. Apparently the lava has not yet reached more than half-way from the crater to sea-level but it is being watched with interest for further developments. Rumblings and explosions have been heard on the mainland more than 12 miles away".

1946-47. There appears to be no report of volcanic activity during the period of the Second World War. Best (1956) reported: "Towards the end of 1946 an eruption even more severe than that of 1919 commenced and continued intermittently until September 1947. During this eruption a large lava flow spilled out from below the crest on the southeast side of the cone, and flowing down a broad avalanche valley, bifurcated near its outer extremity and overran the coast in two places". "Thick deposits of ash and lapilli covered most of the island, rendering unproductive many of the native gardens".

1956-66. Strombolian activity occurred in every year during this period, although there were distinct phases of activity which could be described as separate "eruptions", particularly in 1957-58 and 1960. During these two periods the most violent events recorded at Manam took place.

Reynolds (1957) described the first phase of activity, between about 8 December 1956 and February 1957. During January, ash-free strombolian ejections took place at the southern crater to heights of about 300 m, at an average rate of six or seven per minute. One or more lava flows were extruded, and avalanche phenomena are described, some of which can be interpreted as small nuées ardentes. The explosive activity was strongest in late December 1956 when strong aerial concussions were occurring. In an unpublished account G.A.M. Taylor (1958) gave a summary of the course of events following the resumption of similar activity in mid-1957. A phase of activity in May-July 1957 "was almost identical in nature" with the previous phase, with the exception that the main crater became involved, producing incandescent lava (presumably as lava jets, and not flows) and a little ash. "The next major phase in the eruption began early in October with rapidly fluctuating spasms of explosive activity which quickly culminated on the 18th in a strong eruption. Heavy nuées ardentes swept down the southern flanks..." After this, "gas and ash emission greatly increased, coarse cinder fragments fell on coastal areas for the first time and a more mobile lava flow descended the southern slopes". "On 4th December the volcano began to rumble continuously and the sounds were louder than any heard before. Two days later villages on the eastern coast received a heavy fall of ash and cinder blocks. The largest blocks, measured 2" by 3" - some of them penetrated the roofs of houses. More powerful activity occurred on the next day, 7th December, when a number of nuées ardentes descended the south-eastern valley, and dust and lapilli fell over the northern and western sectors of the island. The eruption reached a peak on the next day, 8th December. The early morning rumbling and ash emission changed at 10 a.m. to spasms of continuous roaring and the voluminous ejection of ash and cinders. For the greater part of the afternoon the northern and western sectors of the island were blacked out by heavy clouds of ash". "On this day further nuées descended the south-eastern valley", and "one of them entered the sea".

The whole population of Manam was evacuated from 10 December. Activity continued at a reduced level after the 8th, although "rhythmical explosive jets continued from the summit vents until the 13th, and nuées continued to be expelled from the southern vent", the last of them on Christmas eve. Fluctuations occurred in the intensity of activity over the next month.

"On the night of January 9th [1958] brilliant lava fountaining originated from the main crater". "At 1600 hours on the 10th of January an enormous cloud rose more than 20,000 feet above the summit with loud roaring and rumbling. This outburst expelled a heavy nuée ardente into the north-east valley", after which "a lava flow then poured over the north-eastern rim". "Heavy ash emission followed this eruption" and "on the 14th a more powerful eruption poured a much larger lava flow into the northwestern [sic, northeastern] valley", which reached to "within half a mile of the coast". Heavy falls of ash and scoria followed, and a lull in activity commenced on the 19th. At 0605 hours on 25 January, the inactive southern vent started silently emitting ash and vapour. An hour and a half later the main crater opened up. At 0800 hours nuées ardentes were expelled onto the southeastern flank" and "For the next five hours the column was fed by incessant explosions roaring and rumbling from the summit vents". "Heavy nuées ardentes were expelled" apparently into all four valleys. "Most of the western side of the island received a deposit of coarse scoria which was 5 to 6 inches thick along the coast." Another major phase occurred on 3 February, which also laid down about 15 cm of scoria on the west coast. During this period "voluminous quantities of fragmental and molten material were discharged and powerful nuées swept down the southeastern flanks". Although some observations suggested "that this eruption was more intense than its predecessor", "no evidence was found that nuées had descended into the other three radial valleys".

Yet another major phase commenced at 1415 hours on 4 March and was the most sustained event on record. It lasted for more than 24 hours and was similar to the previous one although "effusive activity was greatly emphasized and explosive phenomena were correspondingly reduced. Some nuées descended the southeastern flanks during the initial stages" and heavy scoria falls occurred on the eastern side of the island. "The most prominent feature of the activity was a heavy lava flow which descended from the southern crater", and entered the sea on 6 March.

Only occasional minor activity occurred after this, until strombolian activity recurred at the southern crater at the end of June 1958, lasting into August. Brief strombolian ejections also took place at the southern vent in June and July 1959 (data file, Rabaul Observatory).

Taylor (1963) summarised the next main eruptive phase, most of which occurred at the main crater between December 1959 and August 1960. He stated that "a new phase of eruptive activity began from the main crater with Strombolian explosions of increasing intensity", and that "This rising trend of activity reached a climax on 17th March, 1960". "At 0900 hours an explosive outburst greatly augmented the size and height of the eruption column and at the same time nuées ardentes were discharged from the crater. For the next two and half hours gas-charged masses of hot fragmental material were successively poured over the low northeastern rim of the crater to descend at high velocity into the valley below". "Heavy lava outpouring immediately followed this explosive activity with incessant lava fountaining from the crater". "This effusive activity continued with only short intermissions until the end of May".

Strombolian activity occurred again at the southern crater from September 1960, and data files at Rabaul Observatory show that explosive activity occurred frequently at the southern crater during the next few years (e.g. Fig. 3), with minor strombolian activity and probable lava flows occurring in every year from 1960 to 1966 (c.f., Branch, 1967). However, no more activity of this kind occurred at the main crater.

Fig. 3

<u>1974-75</u>. Only minor activity occurred at either crater after April 1966, but in 1974, on 23 May, a brief spasm of lava jetting at the southern crater took place. A few days of lava jetting at the main crater from 1 June 1974 led into a prolonged phase of southern crater strombolian activity which lasted until about mid-August. During this period, nuées ardentes occurred at two periods, 5-6 June and 29 June, followed in each case by lava flow, in the southeast valley. Five short bursts of large-scale activity, each lasting 1-3 hours, took place between 29 June and 10 July. Several heavy falls of ash and scoria were dropped on the west coast during these accumulating to a depth of about 5 cm.

A brief recurrence of southern crater lava jetting in September, and a more prolonged recurrence of intermittent activity from late October, followed. A lava flow again entered the southeast valley at the commencement of the latter phase. Activity died away again early in January 1975. The 1974-75 activity is summarized in greater detail by Cooke et al., (1975; this volume).

## DISCUSSION

### Nuées Ardentes and Lava Flows

Nuées ardentes of basaltic composition (Table 1; Taylor, 1963) appear to be a regular part of the eruptive phenomena at Manam. A nuée can be recognized in the description by Stanley (1923) of the 1919 eruption, and nuées were identified during the 1936, 1957-58, 1960, and 1974 eruptions. Although they are less clearly identifiable in the descriptions of other eruptions, nuées and not lava flows may have been the phenomena observed overnight on 27-28 June 1887 and in 1904, and very probably also in December 1920. The implication of high speed, of rolling motion, and the general destruction, all suggest nuées ardentes. Although there is little information available, it is probable that at least the larger nuées at Manam would be incandescent at night, creating the illusion of a lava flow; J. Holbrook (pers. comm. 1974) observed daylight incandescence at the base of the 29 June 1974, nuée from a vessel offshore. Certainly, the only historical lava flows to be seen on Manam which have entered the sea are those of 1947 and 1958.

Table 1

### Paroxysmal Pattern

It is clear from the foregoing descriptions that the strongest activity occurs only in brief spurts. If observations from passing vessels are excluded, and only reports derived from longer periods of observations are examined, specific dates are often nominated as eruptive peaks - for example 27-28 June 1887; 26 October 1904; 11 August 1919. In 1936, Höltker (1942) describes the typical form of "numerous large paroxysms" which "followed one another at intervals of 5 to 7 days". Each of these lasted, by implication, for less than one night. Taylor (1958) described a similar pattern for the eruption of 1957/58. Furthermore, Cooke et al. (this volume) point out that there were five striking pulses of stronger activity, on 29 June and 1, 3, 6, and 10 July 1974, each of only short duration (1-3 hours).

Although there is some tendency for activity to appear to cluster about the middle and end of the year, the paroxysms seem to have occurred in almost every month.

## Identification of Active Craters

Many of the earlier descriptions are not sufficiently detailed to allow the active crater to be identified. However, where identification is possible the active crater is usually the southern one. Apart from the notable main crater activity in January 1958 and in 1960, the only occasion on which large-scale main crater activity seems to be indicated is in 1898-1900, when "the glow of the fresh lave streams"could be seen on the northern slopes (von Hesse-Wartegg, 1902). Pending more detailed study of eruptive activity since the 1950s it is not possible to deduce any clear-cut pattern of interplay between the southern and main craters.

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Table 1. Chemical analyses of rocks produced by post-1957 eruptions of Manam volcano*.

| | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| $SiO_2$ | 53.90 | 53.05 | 52.70 | 51.10 | 53.08 |
| $TiO_2$ | 0.38 | 0.35 | 0.37 | 0.58 | 0.35 |
| $Al_2O_3$ | 16.60 | 13.90 | 17.00 | 16.65 | 14.84 |
| $Fe_2O_3$ | 4.90 | 6.07 | 4.45 | 4.98 | 5.96 |
| FeO | 3.90 | 3.99 | 4.40 | 4.00 | 3.81 |
| MnO | 0.17 | 0.16 | 0.17 | 0.16 | 0.16 |
| MgO | 6.70 | 8.60 | 6.80 | 6.80 | 7.37 |
| CaO | 10.50 | 10.68 | 10.75 | 10.60 | 10.70 |
| $Na_2O$ | 2.40 | 2.57 | 2.50 | 4.25 | 2.61 |
| $K_2O$ | 0.69 | 0.90 | 0.66 | 0.88 | 0.83 |
| $P_2O_5$ | 0.12 | 0.25 | 0.11 | 0.19 | 0.19 |
| $H_2O^+$ | 0.06 | - | 0.08 | - | 0.07 |
| $H_2O^-$ | 0.07 | - | - | - | 0.03 |
| $CO_2$ | 0.05 | - | 0.02 | - | - |
| TOTAL | 100.44 | 100.52 | 100.01 | 100.19 | 100.00 |

1 Lava from southern crater, January 1957 (Morgan, 1966, specimen 6)

2 Lava from main crater, 1957/58 (Morgan, 1966, specimen 9)

3 Lava from unknown crater, 1962 (Morgan, 1966, specimen 19)

4 Lava from southern crater, April 1964 (Branch, 1967)

5 Nuée deposit from southern crater, October 1957 (Morgan, 1966, specimen 8)

* For an analysis of a Manam lava flow erupted in 1974, see Cooke et al., this volume, Table 2

Figure captions

Fig. 1. North coast of mainland Papua New Guinea, showing Manam and other volcanic islands (stippled).

Fig. 2. Main topographic features of Manam volcano.

Fig. 3. View towards the southwest valley of Manam, showing a small explosion column rising from the southern crater, and the surface of 1958 nuée ardente deposits (foreground). Photograph taken 8 May 1963 by G.A.M.Taylor.

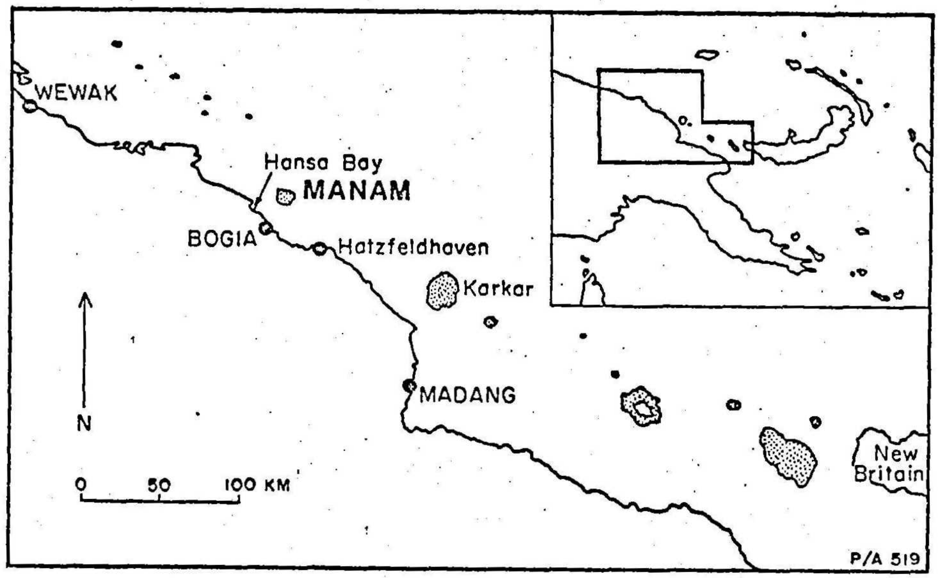
WEWAK
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Fig 1

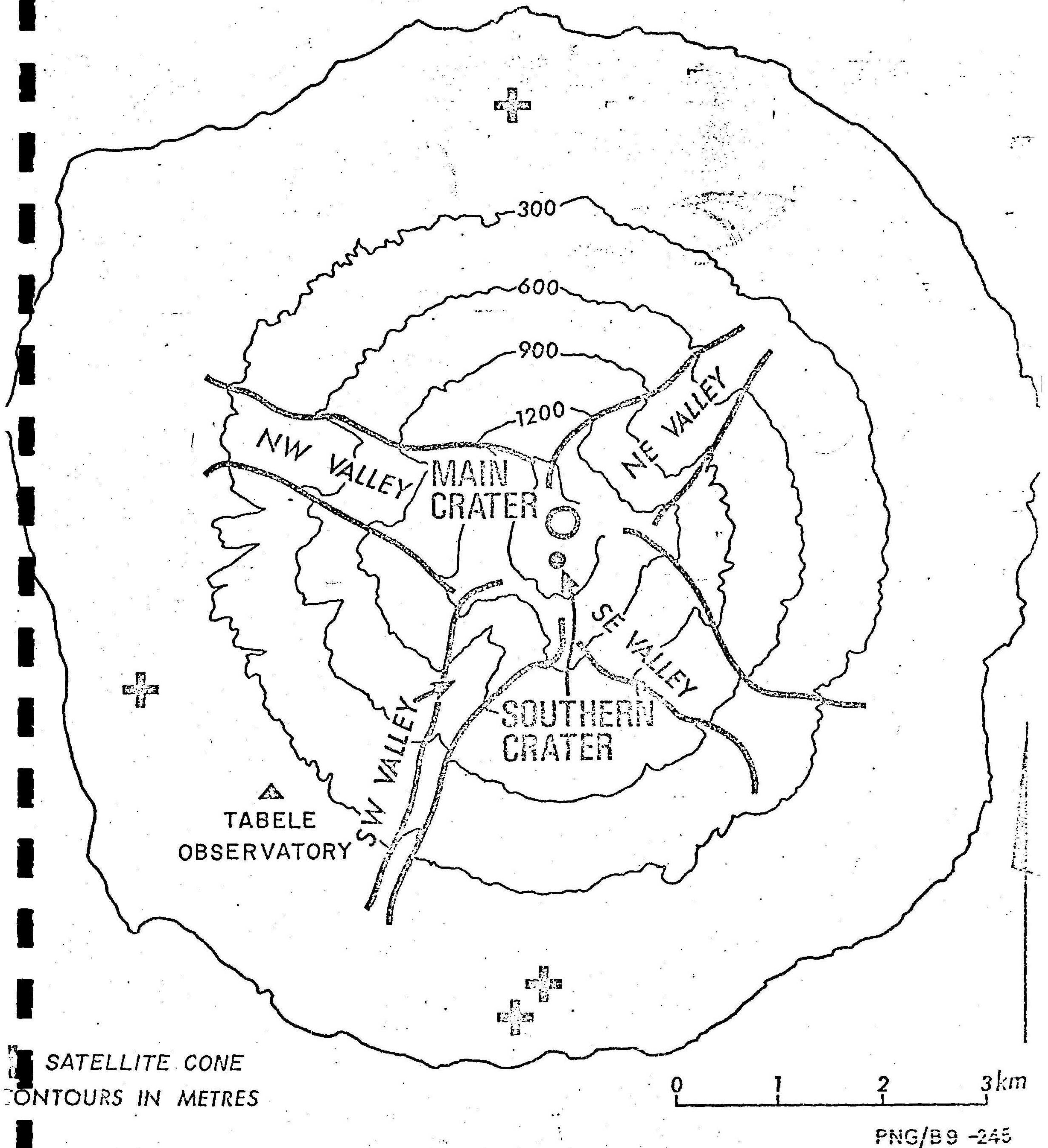


Fig 2
(to be reduced

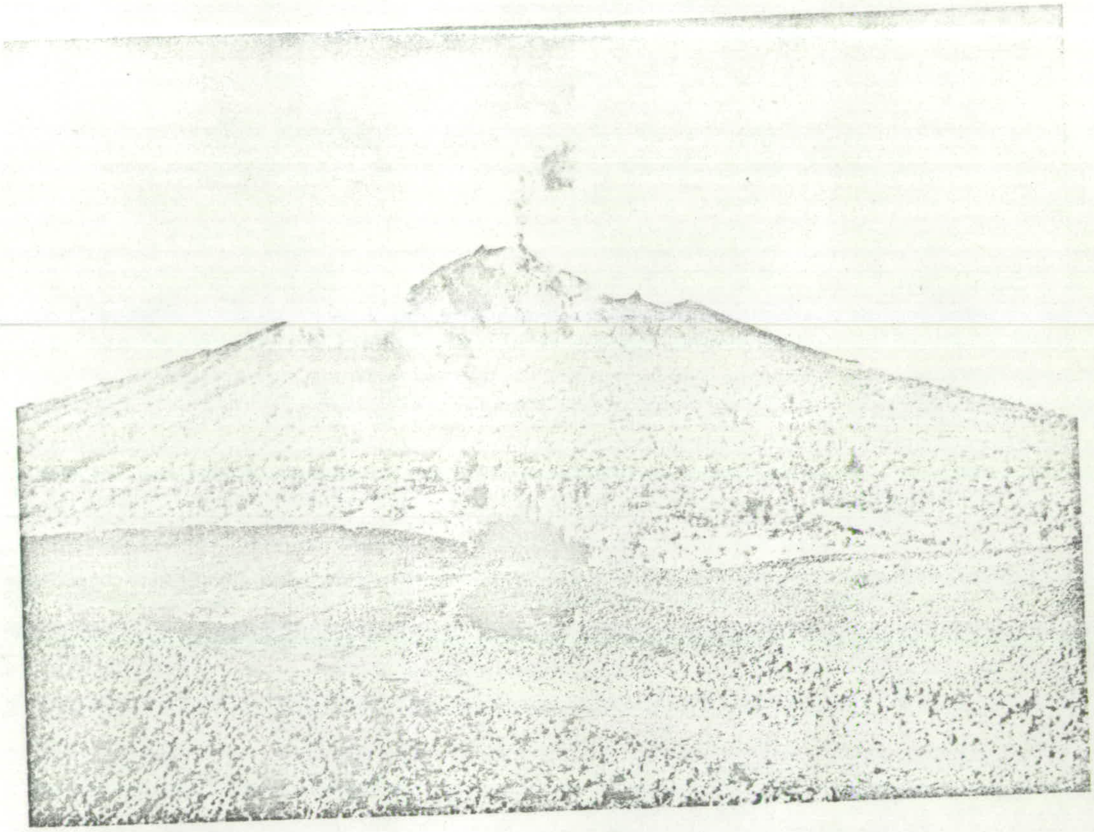

Fig. 3 - Neg. No. G/5500

VOLCANIC HISTORY OF LONG ISLAND, PAPUA NEW GUINEA

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ABSTRACT

Long is a Quaternary volcanic island at the southern margin of the Bismarck Sea. It consists of two stratovolcanoes at either end of a caldera complex containing a fresh-water lake, Lake Wisdom. The outer flanks of the island are covered by a thick pyroclastic mantle which, from geological evidence, the accounts of early explorers, and stories of the islanders, is thought likely to have been deposited during a major eruption sometime between 1700 and 1827 A.D. (and most likely, during the first half of this period). This eruption probably coincided with a period of cauldron subsidence. Subsequent volcanic activity in the southern part of Lake Wisdom built up Motmot Island whose eruptions have been recorded in 1953-55, 1968, and 1973-74.

#9

## INTRODUCTION

Long Island is a Quaternary volcanic complex off the north coast of mainland Papua New Guinea (Fig. 1). It was named in 1700 A.D. by the explorer William Dampier who described it as a "Long Island, with a high Hill at each End". The island consists of two extinct stratovolcanoes at either end of a fresh-water caldera lake, Lake Wisdom, and an active volcanic centre whose eruptions since the Second World have built up an island, Motmot, in the southern part of the lake.

Fig. 1

Long, also known as Arop (or Ahrup), is one of several volcanic islands off the north coast of mainland Papua New Guinea (Fig. 1). These islands form part of a 1000km-long chain of volcanoes along the southern margin of the Bismarck Sea, between the Schouten Islands in the west and Rabaul in the east (see Johnson, this volume). Long Island, Crown Island, and Hankow Reef form a line which intersects the general trend of the volcanic chain at about 30°, and Long and Crown rise from the same northwest-trending submarine ridge (Fig. 1). The western part of the volcanic chain is associated with a belt of seismicity that defines the boundary between the Indo-Australian plate in the south and the "South Bismarck" plate in the north (Johnson and Molnar, 1972). During the last 15-20 years, intermediate-focus earthquakes (between 150 and 230 km deep) have been especially frequent immediately south of Long Island, where they define an almost vertical zone (Denham, 1969; Johnson et al., 1971; Curtis, 1973).

Except for some information given by Fisher (1939, 1957), Taylor (1953, 1956), and Best (1956), little was previously known about the geological history of Long Island. However, geologists from the Central Volcanological Observatory (Rabaul) who have visited the island to investigate the post-war volcanic activity of Motmot, have been struck by the evidence for a major recent eruption on Long Island. →

G.A.M. Taylor, in particular, noticed the young, widespread, volcaniclastic deposits on the island and, hearing stories told by the islanders which were consistent with the geology, concluded that a castatrophic volcanic eruption has devastated the island about one or two centuries ago (Taylor 1953, pers. comm. 1971).

## GENERAL INFORMATION

Fig. 2

Long Island is, in fact, nearly round (Fig. 2). However, when viewed from the sea from many directions, the island appears to consist of two peaks connected by a long, low-lying isthmus, and it is therefore understandable why Dampier named the island as he did (see Fig. 5 and later discussion). The maximum diameter of Long Island is about 30 km, and the total area is about 425 $km^2$.

Mt Reaumur*, which dominates the north end of the island, is an eroded stratovolcano with three main peaks, the highest of which reaches 1304 m above sea level (Fig. 3). The other stratovolcano is Cerisy Peak which rises to 914 m above sea-level at the south end of the island. Both volcanoes flank a larger, central volcanic complex whose centre collapsed to form the steep-sided caldera now occupied by Lake Wisdom. The two stratovolcanoes are older than the last major eruption from the central volcanic complex, which produced an extensive pyroclastic mantle on Long Island. In addition to Mt Reaumur and Cerisy Peak, smaller satellite hills, which were probably minor eruptive centres, are present immediately east of Mt Reaumur, 6 km northwest of Cerisy Peak, and south of Bok village (Fig. 2).

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* Although the peak was named after the French physicist R.A.F. de Reaumur, recent maps show the spelling "Reamur" or "Reumur"

The surface of Lake Wisdom is about 190 m above sea-level, and most of the lake is surrounded by caldera wall cliffs between 200 and 300 m high (Fig. 3). The lake is nearly circular; it has a maximum width of about 11.5 km, and covers about 95 $km^2$. Using a fathometer, Ball and Glucksman (in prep.) found that much of Lake Wisdom has depths greater than 300 m. Repeated soundings in one of the deeper parts of the lake gave depths in the vicinity of 360 m. A water sample from just above lake bottom at this depth was fresh, indicating that the caldera is probably sealed from the sea. The lake level shows annual fluctuations of 1-1.5 m.

Fig. 3

Lake Wisdom has no surface drainage, but it probably feeds springs which emerge from the outer flanks of the island. The most important springs join to form a stream which originates below a low point on the eastern rim of the caldera, and which is the only one on the island that flows to the sea year-round. In 1972, following a period of six months without rain this stream was still running. According to the islanders, it started flowing during the 1930's (Taylor, 1953). Another stream, about halfway between the first one and Malala village, also carries substantial volumes of water, but disappears underground before reaching the sea. Other springs have been reported to emerge at, or near, the coast on the south and west sides of the island (D.A. Wallace, I. Hughes, pers. comms. 1974, 1975).

The recent major volcanic eruption greatly affected the flora, fauna, and human population of the island. Today, Long Island is fully vegetated, but the forest below 900 m has the appearance of being quite young. It is certainly more open than the forests on many of the nearby islands. However, judging from the accounts of visitors to Long Island throughout this century (Coultas, 1933-35; Evans, 1939; Taylor, 1953), who also commented on the youthful appearance of the forest, it appears likely that the lowland vegetation has reached a stunted steady-state caused by the extreme porosity of the soil (cf. Taylor, 1953). The montane forest above 900 m, however, which is almost always in clouds, is similar to that found on the peaks of adjacent volcanic islands (Diamond, 1974).

The faunas of Long and the adjacent islands are in general so poorly known that it is difficult to say whether Long is missing species that would be expected to be present. However, Diamond (1974) has found that Long has far fewer montane bird species, and more "supertramp" species, that would be predicted on the basis of its size and elevation, and this may indicate that the island was devastated in the not-too-distant past. The human population presents a similar picture, as Long has been either uninhabited, or only sparsely inhabited, since 1700 (Dampier). Since about 1930, however, the population has been increasing rapidly, reaching 697 in 1972.

## CENTRAL VOLCANIC COMPLEX

Rocks produced from the central volcanic complex, and deposited before the major eruption that formed the pyroclastic mantle of Long Island, are exposed in the walls of the Lake Wisdom caldera and, to a much lesser extent, on the northeastern coast of the island. The base of the pyroclastic mantle has not been identified in the caldera wall,but it seems that at least the lower half of the wall consists of materials laid down by earlier eruptions of the central complex.

Parts of the eastern and western sides of the caldera wall have been examined, and consist mainly of volcaniclastic deposits, particularly air-fall deposits, but also minor lava flows and some possible nuées ardentes deposits. Many of the deposits appear to have been reworked.

On the northeast and west sides of the island two prominent ridges extend outwards from the northeast and west sides of the caldera, and maintain a more-or-less constant elevation before dropping steeply to the coast (Fig. 2). Both ridges may represent outlying centres of volcanism which have been modified by deposition of later volcaniclastic deposits.

It is difficult to assess the form of the central volcanic complex prior to the formation of the present-day caldera. The caldera has obliterated the centres of eruption, which could have had any, or a combination, of the following forms: a single central crater; a widespread group of craters; a row of craters, or a fissure (perhaps on the line between Mt Reaumur and Cerisy Peak); a caldera, or calderas, smaller than the present-day caldera; etc.. However, the complex was probably unlike the large stratovolcanoes found elsewhere along the southern margin of the Bismarck Sea, whose flanks slope at up to about $35^{\circ}$ (e.g. Manam, Fisher, 1957; Ulawun, Johnson et al., 1972). Because lava flows appear to be extremely rare in the Lake Wisdom caldera walls, it is thought the central complex probably had the form of a large pyroclastic volcano, whose flanks sloped at perhaps no more than about $10^{\circ}$ (c.f. "ash flow volcanoes" described by Macdonald, 1972). In profile, therefore, the complex may have been low-lying and, depending on the form and distribution of its eruptive centres, and on the rate at which reworked clastic deposits were redeposited at its periphery, the profile could have been more-or-less flat-topped.

Five rocks from the central complex have been chemically analysed (Morgan, 1966; Johnson, in prep.). Three are basalts (53% $SiO_2$, or less), and two are low-silica andesites (more than 53, and up to 57, % $SiO_2$).

MT REAUMUR AND CERISY PEAK VOLCANOES

Mt Reaumur is an eroded, steep-sided volcano, and forms the highest point on Long Island (Fig. 3). Its base is surrounded by thick pumiceous materials deposited mainly by nuées ardentes and lahars (see below). The upper part of the volcano consists of a series of ridges which define two watersheds, each drained by a stream valley which reaches the coast (Fig. 2). Porphyritic lava flows and intercalated scoriae and lapilli crop out in the eastern stream up to at least 720 m above sea-level.

Cerisy Peak is a similar, steep-sided and deeply dissected volcano. Stream courses radiate from a 2 km-long ridge at the summit of the volcano, and at the northern end of the ridge is the highest point which, when viewed from the south and southwest, has a dome-shaped profile. Loose boulders in a stream on the southwest side of the volcano are thought to have been derived from minor intrusions which may be exposed at higher levels.

The rocks of both volcanoes are basalts. Of four samples chemically analysed, three are olivine tholeiites and one is a quartz tholeiite (Johnson, in prep.).

## PYROCLASTIC MANTLE

Fig. 4

Poorly consolidated volcaniclastic deposits are exposed along most of the coastline of Long Island (Fig. 4). They cover the surface of the central volcanic complex, constitute the upper part of the caldera wall, and probably mantle most of Mt Reaumur and Cerisy Peak volcanoes.

The pyroclastic mantle shows a range of thickness and lithology, but overall it appears to consist of three parts: "basal" and "upper" well-bedded units separated by a "middle" unsorted unit containing tree trunks and branches. This sequence suggests that a series of explosive eruptions preceded a catastrophic eruptive phase, which was followed by a period of waning explosive events.

Basal unit. The lower bedded unit crops out at the base of cliffs along many parts of the coast, but the bottom of the unit has not been observed. The maximum measured thickness is about 4 m at a point of the west coast, where the sequence consists of internally bedded and unbedded lapilli and ash layers, one of which contains uncharred leaf remains. Taylor (1953) described a similar sequence from the east coast of the island: "Numerous plant fossils and imprints left by standing trees" were contained in the basal part of this sequence, which was overlain by a layer of accretionary lapilli, a lapillus layer, another accretionary lapilli layer, and a "chocolate tuff containing marine fossils". This sequence suggests initial subaerial deposition followed by subsidence and submarine deposition and later uplift. Layers of accretionary lapilli and coral fragments are present in the basal unit at other coastal exposures.

Middle unit. The unit is between 3 and 5 m thick on the coast (Fig. 4). Unlike the basal unit, it is unsorted and unbedded, and in places shows crude columnar jointing. The middle unit consists mainly of pumice lapilli, accretionary lapilli, glassy angular lava chips, and fragments of crystalline lava; these are enclosed in a light-coloured earthy matrix. At numerous localities, both carbonised and uncarbonised fallen tree trunks and branches are present, particularly near the base.

The middle unit was almost certainly deposited by pyroclastic flows, but at several localities it is difficult to judge whether the deposits are those of nuées ardentes or of lahars. However, the deposits containing accretionary lapilli and uncarbonised wood, were probably laid down by cold lahars, or slurries, rather than by hot nuées ardentes, which are more likely to disintegrate accretionary lapilli, and char vegetation. Exposures of the pyroclastic mantle in the western stream draining Mt. Reaumur (Fig. 2) show thicknesses of over 30 m (similar values were reported by Taylor (1953) from the east flank of Long Island), and provide evidence for at least two superincumbent flow-units.

Deposits of the middle unit are found only on the lower slopes of Long Island. They appear to be absent from the upper parts of Mt Reaumur and Cerisy Peak volcanoes, and possibly also from other high points on the island. During their movement down the outer flanks of the central volcanic complex, the pyroclastic flows moved around obstacles. Mt. Reaumur, for example, is completely surrounded by middle-unit deposits; and the satellite hill southwest of Bok, and the small hills east of it, may have been off-shore islands and sea stacks before the deposits were laid around them. Unless immediately removed by the sea, the pyroclastic mantle, and especially deposits of the middle unit, may have extended the coastline of Long Island, particularly on its southeastern and northwestern sides. However, if the east coast was extended later erosion and subsidence must have taken place because the remains of coastal settlements buried by the middle unit are now below mean sea-level and have been almost entirely removed by the sea (I. Hughes, pers. comm., 1975).

Upper unit. The upper bedded unit occupies the high part of many coastal cliff sections (e.g. Fig. 4), but is absent in other places presumably having been removed by erosion. At most exposures the unit is only about 1 m thick, or less, but at a few localities the thickness is more than 2.5 m. The upper unit consists of well-bedded lapilli, and commonly contains well-formed accretionary lapilli, up to 1.5 cm in diameter. These accretionary lapilli are also found in the soil cover which extends up to the rim of the caldera. No wood fragments appear to be present in the upper unit.

The following evidence indicates that rain storms accompanied deposition of the pyroclastic mantle; cross-bedding and infilled stream channels, particularly in the basal and upper units; the presence of accretionary lapilli, formed when fine volcanic ash aggregated during rain storms; and the presence of mud flow deposits.

Two vesicular clasts from the pyroclastic mantle have been chemically analysed. They have the highest silica contents (56.7 and 58.5%) of all the analysed rocks from Long Island, and although not necessarily representative of the entire pyroclastic mantle, they suggest that the major eruption produced the most fractionated rocks on Long Island.

CALDERA AND ITS ORIGIN

Lake Wisdom caldera is 13.5 km from northwest to southeast, 11 km from southwest to northeast, and has an area of about 120 $km^2$ (Fig. 2). It is slightly larger than, for example, Hargy and Dakataua calderas in New Britain (Fisher, 1957; Branch, 1967; Lowder and Carmichael, 1970), and Crater Lake, Oregon (Williams, 1942). The highest parts of the caldera wall are in the north and west, where the rim is at least 490 m above sea-level (Fig. 3). The lowest point on the caldera rim is a breach in the eastern wall, which is at least 30 m above lake level.

The caldera has two prominent embayments on its western and eastern sides (Fig 2). The western embayment is defined by two peninsulas which project into the lake (Fig. 3). The northern peninsula has a steep escarpment on its eastern side, and the western part appears to be the southwest-dipping constructional surface of the central volcanic complex. The embayment seems to have formed by partial collapse of a separate arcuate segment, the northern end of which remained attached to the northern caldera wall. Water depths in this embayment are mostly less than 150 m (Ball and Glucksman, in prep.). In contrast, the eastern embayment is not flooded by the lake (Fig. 2); its forest-covered base dips southwards, and no high cliffs are preserved at lake-level.

Volcaniclastic beds in the southern peninsula of the western embayment, and in cliffs west and northwest of the eastern embayment, dip inwards towards the caldera. These inward dips could be due to mantling of parts of the caldera escarpment, but it is more likely they are the result of downwarping which accompanied collapse of both embayments. Elsewhere in the caldera wall similar beds appear to dip away from the caldera.

The pyroclastic mantle of Long Island closely resembles in lithology, thickness, and extent, the pyroclastic mantles associated with many calderas, throughout the world, in which a close genetic relationships is implied, or can be proven, between cauldron subsidence and the eruption of copious pyroclastic materials (cf. Williams, 1941; McBirney and Williams, 1969). In the field, however, it is extremely difficult to identify the precise geological relationship between the eruptions and subsidence. It is impossible to say, for example, if the eruptions took place from central vents, from caldera ring faults, or from ring fractures associated with the caldera; if cauldron subsidence took place before, during or after the eruptions; or if the subsidence was caused by deep-seated withdrawal of magma, by gravitational collapse into a less-dense body of magma, by subsidence into a chamber evacuated by a catastrophic eruption, or by the initial formation of cone-sheet fractures (by doming) and subsequent subsidence. On Long Island it is also not certain that the Lake Wisdom caldera was formed entirely during the period of deposition of the pyroclastic mantle. Indeed, the abundance of volcaniclastic materials in the lower parts of the caldera wall could be taken to indicate the contrary - that a series of subsidence events took place, each associated with a major pyroclastic eruption, and that only the last of these produced the caldera as it is seen today.

AGE OF THE LAST MAJOR ERUPTION

Several lines of evidence can be used in an attempt to date cauldron subsidence and deposition of the pyroclastic mantle. These include direct geological observations and carbon-14 dating, the accounts of early explorers, and stories told by the islanders.

Geological evidence. The young age of the pyroclastic mantle is suggested by the poor consolidation of the deposits, and by the fresh appearance of the uncarbonized wood remains, which seem to have undergone little alteration since their burial. G.A.M. Taylor also obtained carbon-14 dates from two samples of wood collected in the early 1950's - one from a carbonized tree trunk from the base of the middle unit on the east coast, the other, an uncarbonized sample from a clastic layer in the eastern part of the caldera wall. Both samples gave dates of not older than the beginning of the 19th century (unpublished BMR data). More recently, however, I. Hughes (pers. comm., 1975) obtained two carbon-14 dates of 1720 ±75 and 1750 ±65 A.D. on separate charcoal and charred wood samples from the middle unit on the northwest coast. Three carbon-14 dates from coastal habitation sites showed that the island had previously been inhabited for at least 800 years, though not necessarily continuously.

Accounts of explorers. The earliest recroded description of Long Island was by William Dampier in 1700*.

> The 31st in the Forenoon we shot in between 2 Islands, lying about 4 Leagues asunder; with Intention to pass between them. The Southermost is a long island, with a high Hill at each End; this I named Long Island. The Northermost is a round high Island towering up with several Heads or Tops, something resembling a Crown; this I named Crown-Isle, from its Form. Both these Islands appear'd very pleasant, having Spots of green Savannahs mixt among the Wood-land: The Trees appeared very green and flourishing, and some of them looked white and full of Blossoms. We past close by Crown-Isle, saw many Coco-nut-Trees on the Bays and the Sides of the Hills; and one Boat was coming off from the Shore, but return'd again. We saw no Smoaks on either of the Islands, neither did we see any Plantations; and it is probable that they are not very well peopled. We saw many Shoals near Crown-Island and Riffs of Rocks running off from the Points, a Mile or more into the Sea. My Boat was once over-board, with Design to have sent her ashore; but having little Wind, and seeing some Shoals, I hoisted her in again, and stood off out of Danger.

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* Dampier's "Voyage to New Holland" was originally published in two parts - the first in 1703, and the second in 1709. These works have since been republished many times, under different editorships. The editions consulted for this paper are those of Masefield (1906) and Williamson (1939).

Dampier also made profile drawings of the islands he saw along the north coast of mainland Papua New Guinea. Although the drawings are unlabelled, and no cross-reference is made with the text*, analyses by Reche (1914) and R.J.S. Cooke (pers. comm., 1974), indicate that Long Island should be shown in Table 13, sketches 4 and 5, of Dampier's book. In sketch 4 (Fig. 5b) the bearings are not consistent with the way the islands are drawn. However, when the sketch is redrawn according to the bearings (Fig. 5c), the new arrangement is consistent with the islands being Long and Tolokiwa (R.J.S. Cooke, pers. comm., 1974).

Fig 5

Dampier did not land on Long Island, and his sketch map (Fig. 5a, showing no caldera) is simply a coastal outline. The profile illustrated in Figure 5b & c, however, shows the interior as low-lying and flat-topped, which, as previously discussed, does not necessarily indicate the presence of a central caldera at that time, but does exclude the presence of any large central cones.

Dumont D'Urville (1833) sailed past Long Island in August, 1827, named the two main peaks, and observed that "The ground in the vicinity of the shore appeared more arid than all the other islands....we saw neither coconut trees nor any trace of inhabitants".

In 1884, Finsch (1888) visited Long Island, which he described as being heavily wooded throughout, although showing more undergrowth-covered areas than Karkar Island to the west (Fig. 1). He noted two or three small settlements along the coast of Long. Later reports also agree that the island was fully vegetated, although three observers - Coultas (1933-1935), Evans (1939), and Taylor (1953) - all suggested that the vegetation indicated the possibility of a recent eruption.

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* According to Williamson (1939, viii), this absence of a close correlation between the illustrations and the text is due, "to the fact that Dampier was at sea when the second part of his book was printed and was dead when the second edition appeared in 1729".

<u>Stories of islanders</u>. Stories concerning the eruption of Long Island have been collected from the local people on many different occasions. Unfortunately, the stories are not consistent. Coultas (1933-1935), for example, stated:

> According to native legends, Ahrup was at one time a large active volcano, much higher than Tolokiwa, and with a large population. Eventually an eruption occurred which blew the cone completely out of the center of the island, throwing out hot stones and lava and killing the people, with the exception of one woman who escaped in a canoe to the mainland of New Guinea where her descendants are supposed to be living now.

Taylor (1953) wrote:

> Recent investigation by A.D.O. [Assistant District Officer] Parish suggests that the eruption was of comparatively recent origin as stories of the escape from Arop are still current among natives of the surrounding islands. It seems evident that some very alarming warning phenomena preceded this eruption as a considerable number of natives appear to have escaped from the island before the catastrophic eruption took place.
>
> Mr. Parish believes that the Siassi Island people originally come from Long Island, and has found, on the harsher parts of the neighbouring New Guinea coast, settlements of natives who are also evacuees. One group, he believes, settled near Lutheran Anchorage on northern Umboi but were subsequently wiped out by the 1888 eruption of Ritter Island.

Stories concerning the recolonization of the island are another potentially useful source for dating the eruption. Data collected by several investigators indicate that some of the people now living on the island are fifth and sixth generation islanders, and the islanders claim their ancestors returned to the island at a time when the vegetation was just starting to become re-established. However, as pointed out by T. Harding (pers. comm., 1972), dating by this technique can be uncertain, because many Melanesian cultures tend to maintain their genealogies to only a certain number of generations, and then drop out additional generations between themselves and the founding ancestors.

Conclusions. The above evidence indicates that the major eruption took place sometime during the first half of the eighteenth century. To our knowledge no observations of Long Island were recorded between 1700 (Dampier) and 1827 (Dumont D'Urville) and this is the only record-free interval in the post-1700 period long enough for revegetation to have taken place following an eruption without the post-eruption devastation being noted.

The relationship of this major eruption to the time of formation of the Lake Wisdom caldera is less clear. It is possible that cauldron subsidence was a short-lived event that took place at the same time as the eruption; but the possibility cannot be excluded that the caldera has had a long and complex history consisting of several periods of subsidence and associated eruptions. In either case, the central volcanic complex of Long Island may never have had a high central peak; or if it did, the peak had disappeared by 1700 A.D.

## POST-CALDERA ACTIVITY

The first well-documented records of volcanic activity in Lake Wisdom are aerial photographs taken in 1943, which show a horseshoe-shaped, low-lying island crater about 3 km from the southern shore of the lake. The photographs show material from the active crater discharging into the lake. However, after eruptions in 1953, islanders who had formerly been unable to provide information about previous eruptions, informed Best (1956) that there had been minor eruptions within the lake in 1933, 1938 and 1943. According to N.H. Fisher (pers. comm., 1972) no island was visible from the caldera rim in 1938 (cf. Fisher, 1939). Taylor (1953) and J.G. Best visited Lake Wisdom in August, 1952, but saw no sign of volcanic activity, and from lake-level were unable to see traces of any island.

In May, 1953, volcanic activity was reported from the same site in Lake Wisdom (Best, 1956). An island was created consisting of two contiguous craters which joined to form a ridge about 400 m long, 100 m wide, and 30 m high. According to Taylor (1956), volcanic activity continued periodically until January 7, 1954. Incandescent ash was emitted from the same site on June 5, 1955, and activity is said to have reached a climax on June 13, before terminating abruptly (Fisher, 1957).

Except for fumarolic activity observed in 1961 (unpublished data, Central Volcanological Observatory), the island appears to have remained quiescent until 1968. During this interval, wind and waves eroded the island until only three small islets remained. In March, 1968, a new period of explosive activity began, and created an island about 300 m long by 180 m wide (G.W. D'Addario, pers. comm., 1972). Further activity took place, and by

Fig. 6

November, 1969, a second small island could be seen just above water-level, separated from the main island by a channel about 1 m deep (Fig. 6). On all subsequent visits this second island could not be seen although a shoal was still present below lake-level.

Conditions on Motmot in November, 1969, were described by Bassot and Ball (1972) and conditions up to November 1972 have been summarized by Ball and Glucksman (in press). The situation in November, 1972, is described in Figure 7. The bathymetric map of Ball and Glucksman (in prep.) suggests that Motmot caps a slightly elongate cone whose northeast-trending axis extends southwestwards towards a submarine ridge running out from the caldera wall. Another submarine ridge trends east-southeastwards from the peninsula on the

Fig. 7

west side of the lake (Fig. 2) as far as the Motmot cone. This ridge has a high area about halfway between the caldera wall and Motmot. It is unknown if these submarine ridges are the result of post-caldera volcanic activity, or if they represent submerged parts of the pre-caldera central volcanic complex.

Ball and Glucksman (in prep., b) measured water temperatures at various points around the circumference of Motmot in 1969, 1971, and 1972. Shifts in the position of the thermometer by only a few centimetres caused a change of more than 20°C in the recorded temperature, but the following generalized trends could be inferred from the data: (1) a general cooling during the interval 1969-1972; (2) rising temperatures in the crater pond, and on the east side of the island north of the saddle, in 1972. A decrease in the island's circumference at the water line, from 892 m (in 1969) to 845 m (1971) to 814 m (1972) was also noted.

The only activity observed between 1969 and 1972 was the opening of a vent, about 1.5 m in diameter, in the crater wall on the eastern shore of the pond between October, 1971, and November, 1972. This hole was probably caused by a gas or steam explosion; no molten material appears to have been expelled. Motmot was again intermittently active between May, 1972, and February, 1974, producing lava flows (see Cooke et al.,this volume).

Two rocks formed during the 1968 eruption of Motmot (Johnson, in prep.) and four formed in 1973-74 (Cooke et al, this volume) have been chemically analysed. All are tholeiitic basalts.

## CONCLUSION AND ACKNOWLEDGEMENTS

This paper provides a broad overview of the volcanic evolution of Long Island. It is a preliminary account which we hope will stimulate further, more detailed, geological work on the island.

We thank I. Hughes and R.J.S. Cooke for their criticisms of the draft manuscript and for supplying some of the factual data, and gratefully acknowledge reviews by referees D.H. Blake and J.G. Jones. The paper is published with the permission of the Director of Bureau of Mineral Resources, Canberra.

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## Figure captions

Fig. 1. Locality map. Dashed line is 1000 m isobath.

Fig. 2. Long Island in 1972, adapted from United States Army 1 inch to 1 mile map (1943). Contours in feet.

Fig. 3. Southern flank of Mt. Reaumur and northeastern wall of the Lake Wisdom caldera, viewed northwards from the western shore of Lake Wisdom, showing part of the western embayment (cf. Fig. 2).

Fig. 4. Deposits of the pyroclastic mantle on the northeast coast of Long Island. The lower two-thirds of the cliff are made up of deposits of the "middle unit", which are overlaid by bedded "upper unit" deposits containing accretionary lapilli; the "lower unit is not exposed here. Cliffs are about 5 m high.

Fig. 5. Facsimiles of diagrams drawn by William Dampier during his voyage in 1700 A.D. (Williamson, 1939). (a) Detail from chart showing Dampier's course (cf. Fig. 1); note in particular the elongate ("long") outline of Long Island (cf. Fig. 2). (b) Profiles of islands in Table 13, sketch no. 4, of Dampier. (c) The islands of b transposed to give a consistent change in bearings across the diagram; the left-hand island is Long, and the right-hand one is identified as Tolokiwa.

Fig. 6. Motmot Island from the southwest in November, 1969, showing the smaller, un-named island west of its southern end.

Fig. 7. Sketch-map of Motmot Island, Lake Wisdom, in November, 1972. The southern part of the island consists of a northeast-southwest ridge, about 23 m high, made up of poorly consolidated ash and lapilli beds that dip southwards at about 30°. Bedding is especially prominent in those parts of the ridge marked by the stippling. The ridge is surrounded by scree slopes. The northern part of the island is a pyroclastic cone, containing a crater, which is partly filled by a pond. The cone consists of bedded ash, lapilli, bombs, and blocks. The crater rim was continuous as late as October, 1971, but aerial photographs taken in March, 1972, showed that the crater wall had been

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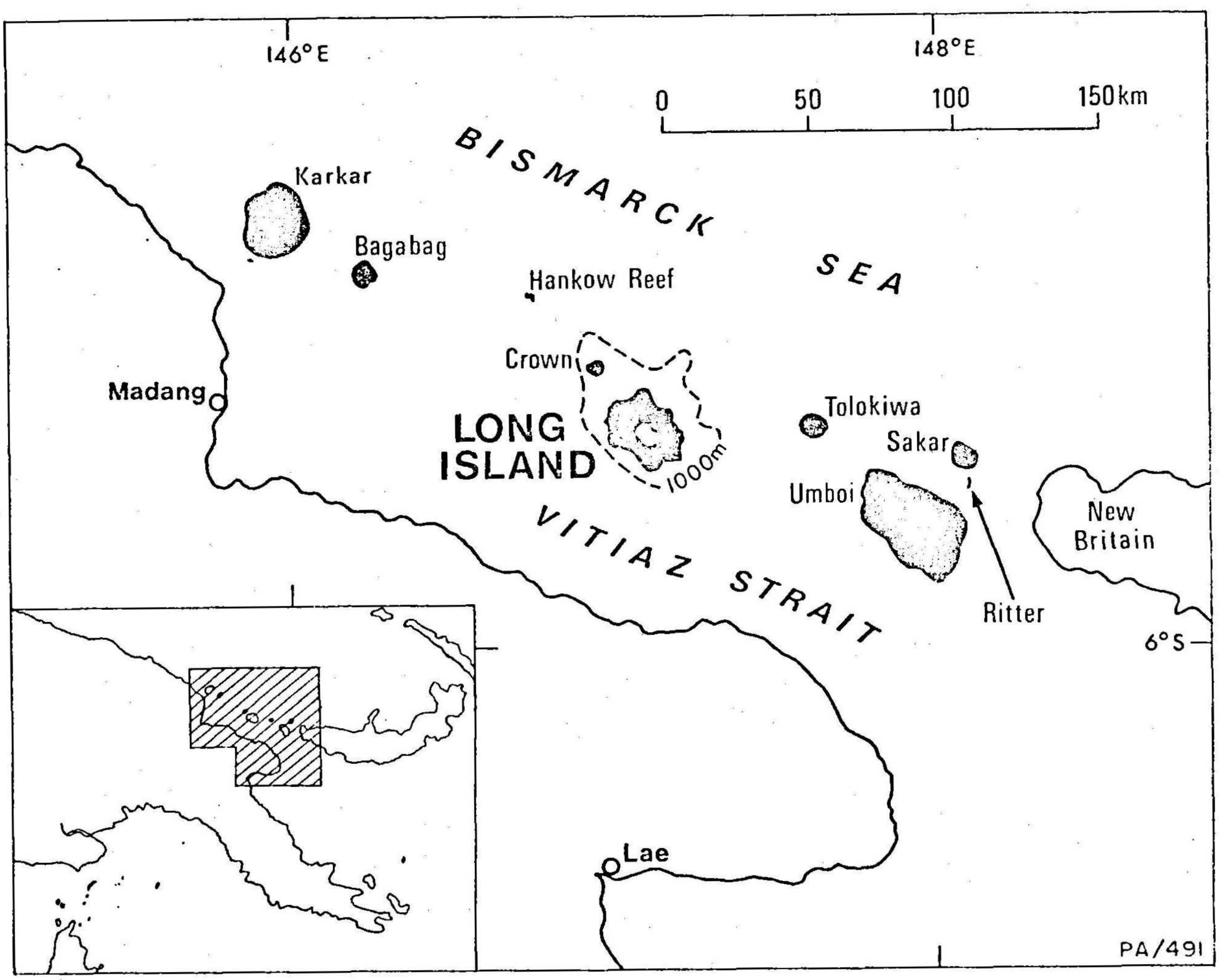
breached, and that the pond was connected with Lake Wisdom. This sketch shows that by November, 1972, the breach had been partly sealed again by beach deposits that form a fan at the edge of the crater pond.

Fig 1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LONG
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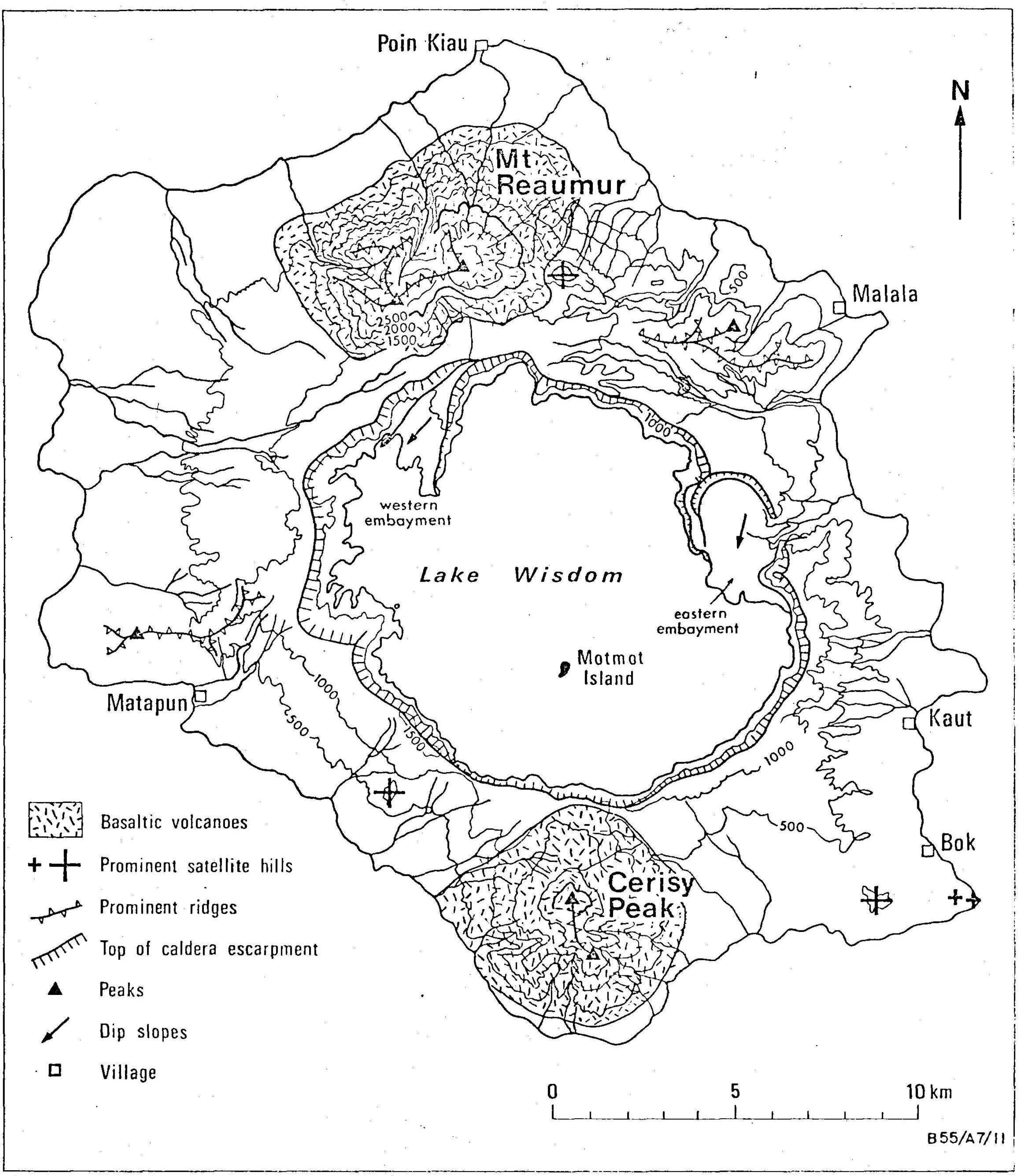
Fig 2



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Mt Reaumur
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Malala
2500
2000
1500
1000
western embayment
Lake Wisdom
eastern embayment
Motmot Island
Matapun
1000
1500
1500
Kaut
1000
500
Bok
Cerisy Peak
Basaltic volcanoes
Prominent satellite hills
Prominent ridges
Top of caldera escarpment
Peaks
Dip slopes
Village
0
5
10 km
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Fig. 3

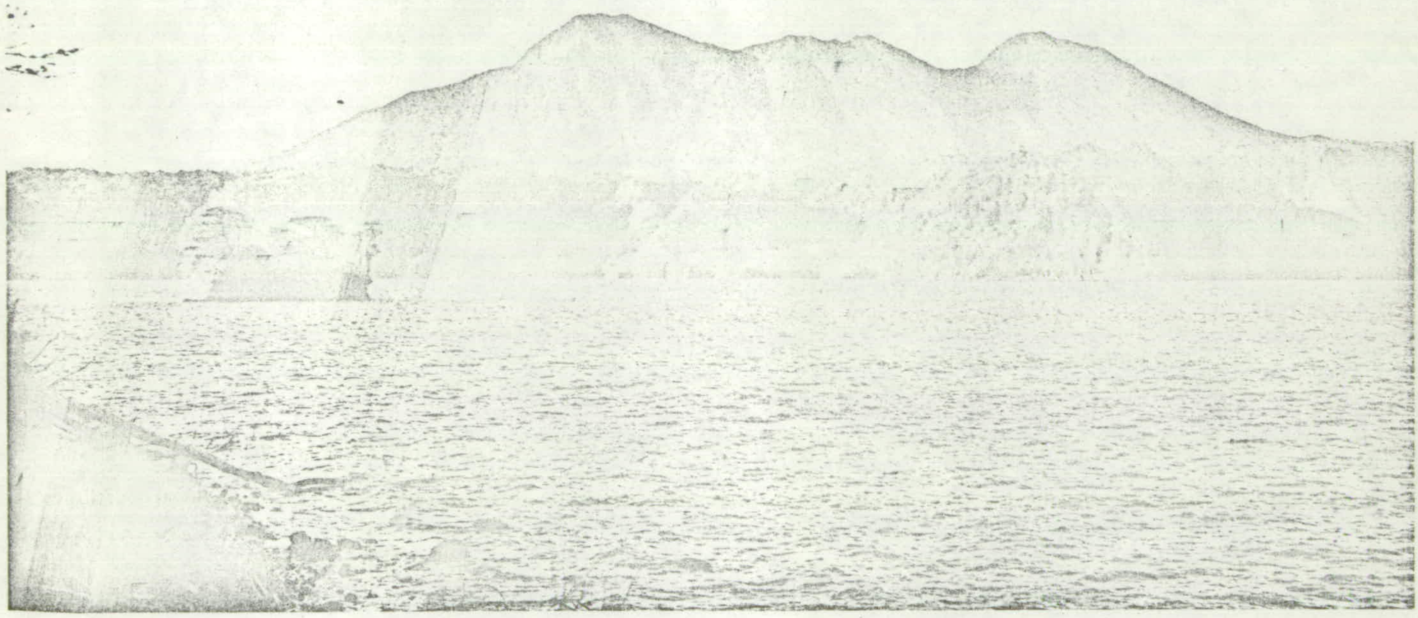


Fig 4

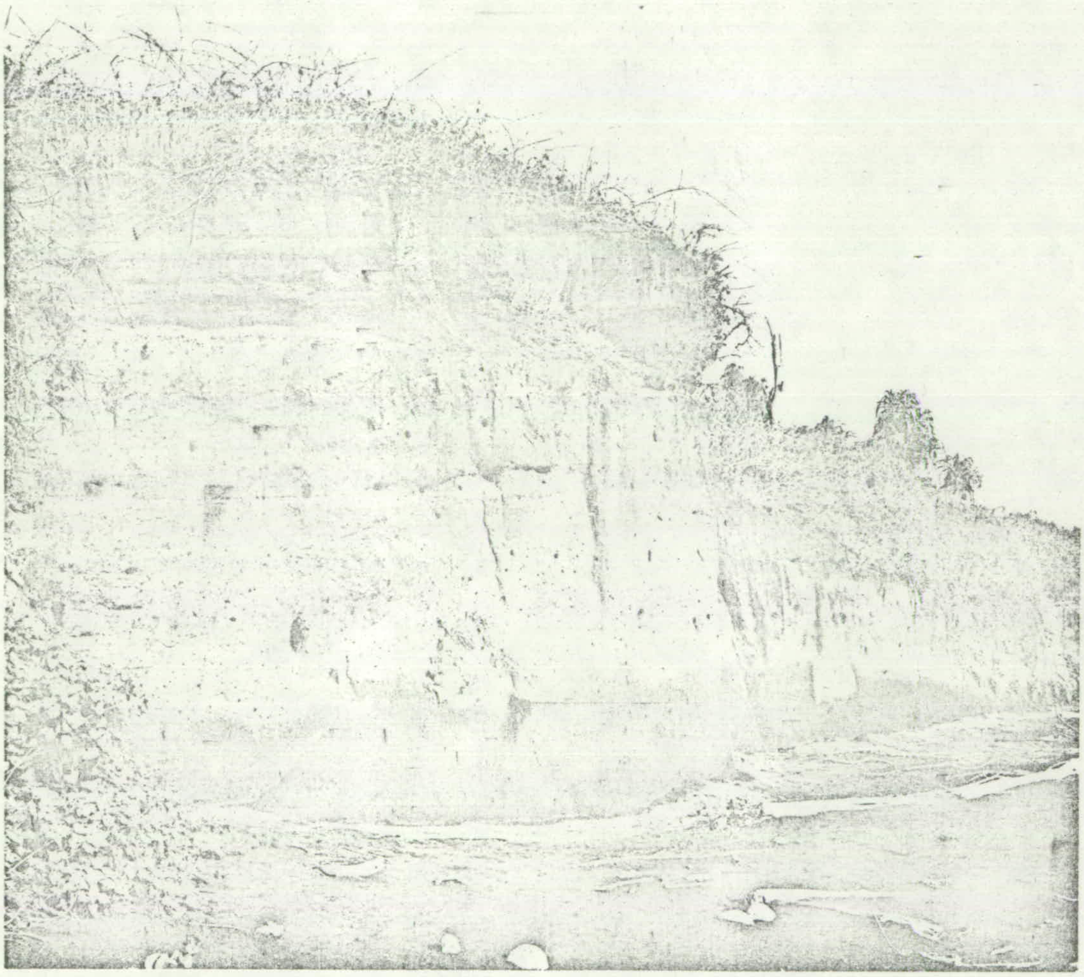
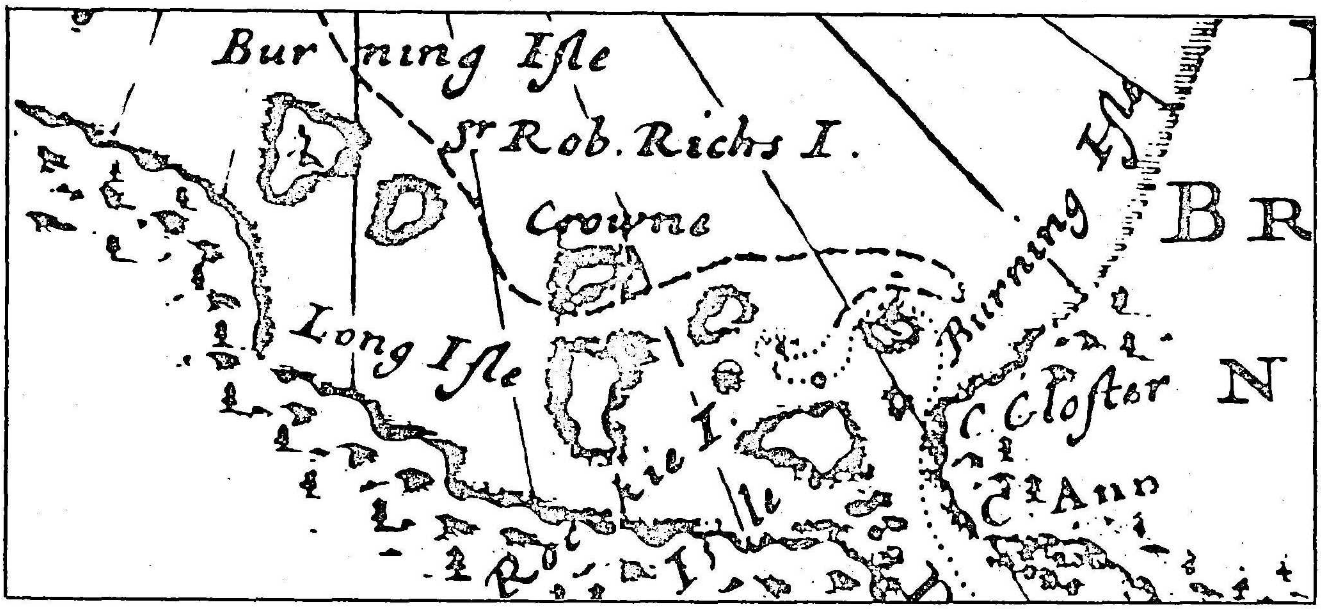


Fig. 5


Burning Isle
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Crowne
Long Isle
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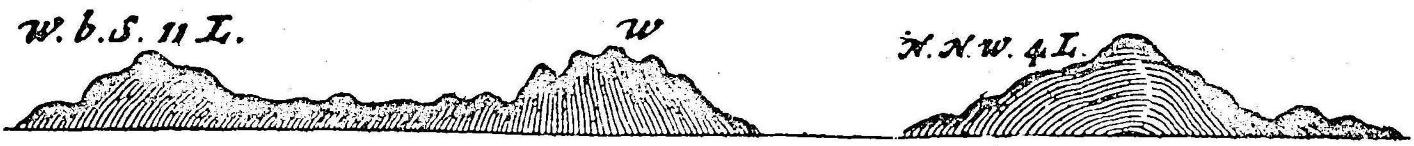
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Fig 6

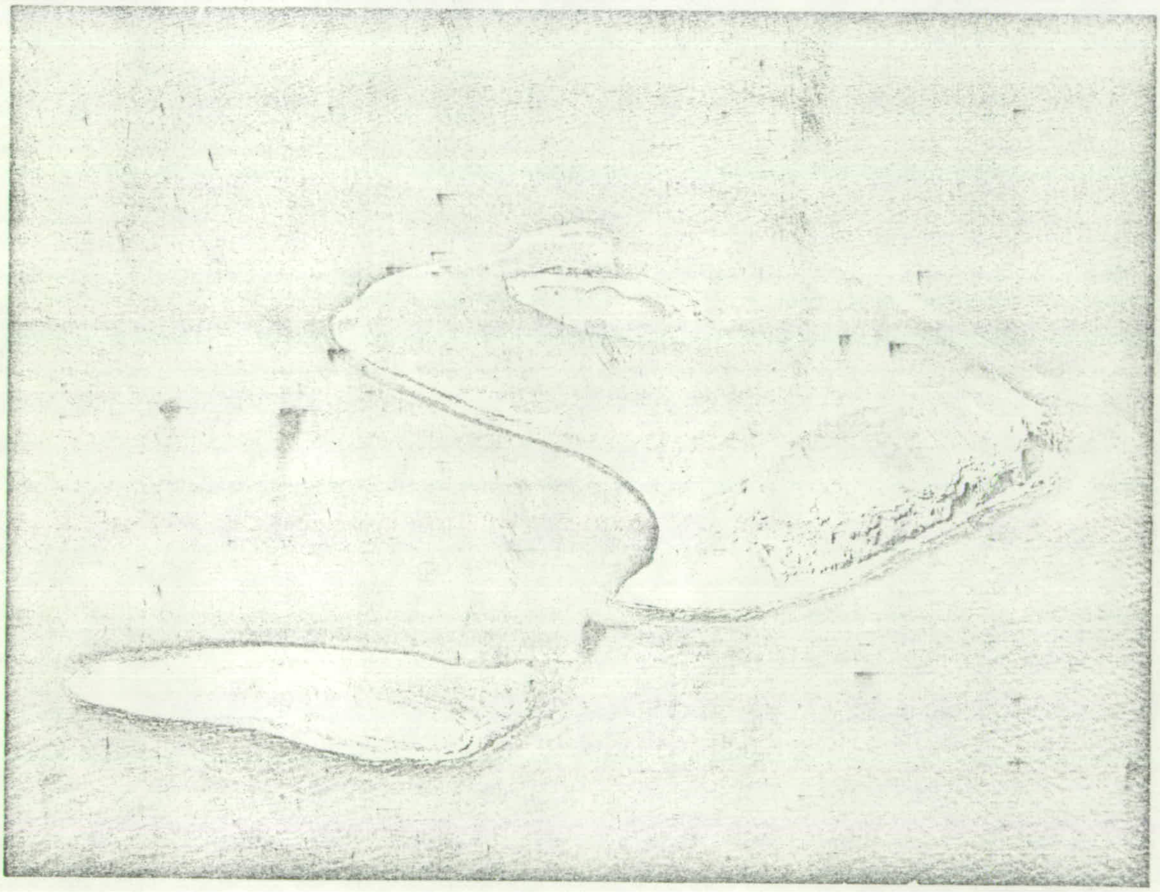

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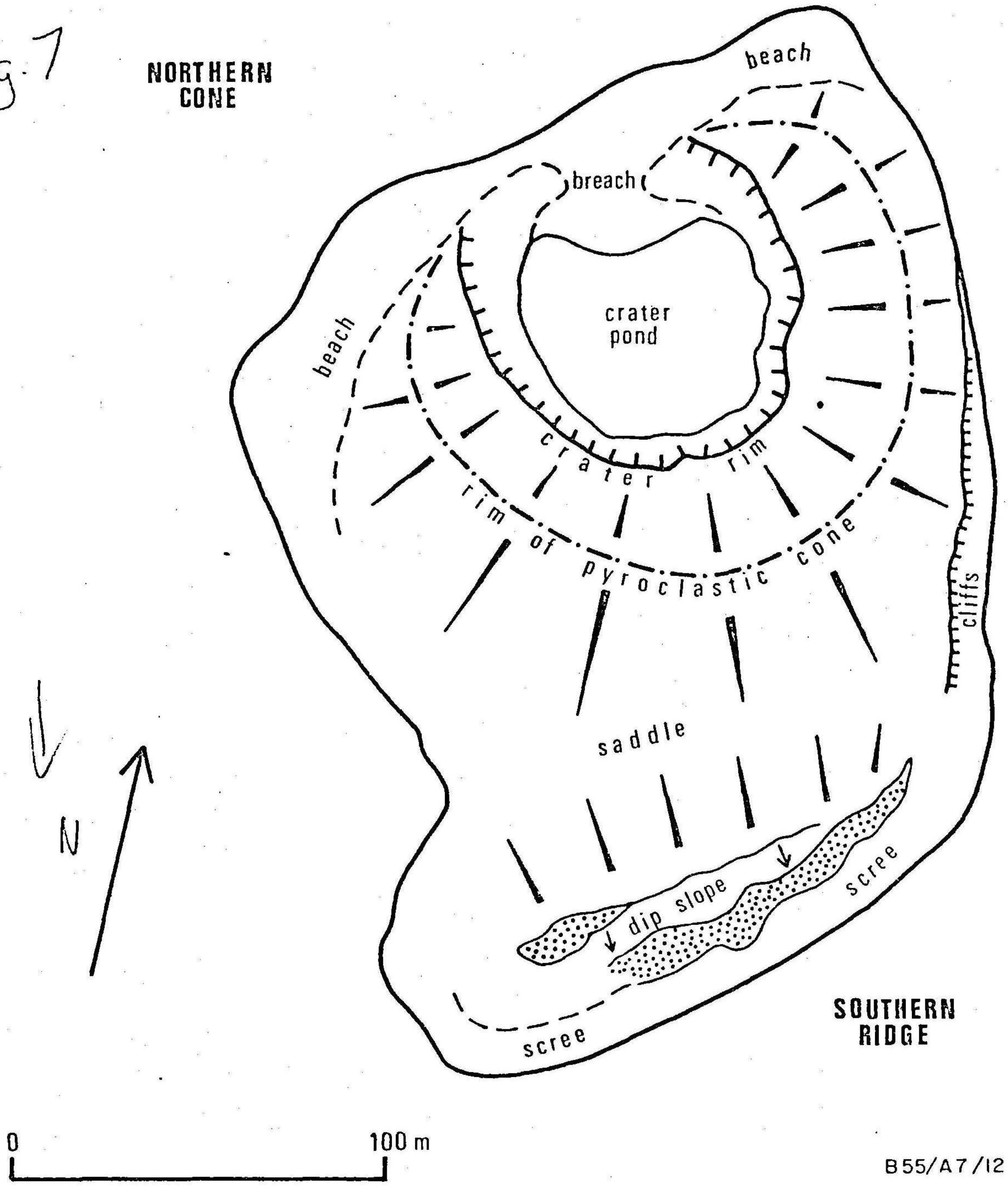
Fig. 7




B55/A7/12

# 1941-42 ERUPTION OF TAVURVUR VOLCANO, RABAUL, PAPUA NEW GUINEA

N.H. FISHER
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ABSTRACT

The eruption of Tavurvur between 6 June 1941 and March 1942 was characterized by a long build-up period which began after ~~with~~ the violent explosive outburst of Tavurvur that accompanied the Vulcan eruption of 1937. Pre-eruption phenomena included: a general but intermittent increase in size of the gas and vapour cloud given off from the central crater; increases in the contents of sulphur dioxide, hydrogen chloride and, later, other high temperature gases; development of new active fumaroles in the 1878 crater and on the outer southeastern slopes of the volcano; and some minor explosive activity at the southwestern end of the main crater between February and May, 1940. The temperatures of the gases given off by the fumaroles in the 1878 crater first rose above 100°C in November 1940, and had risen to nearly 400°C just before the eruption started on 6 June 1941. Pre-eruption seismic effects were too small to be identified on the Observatory seismograph (7.2 km away, with a magnification of 1:150), or on a similar instrument installed in April 1941 at an observation post 1 km from the crater. After the eruption began, individual outbursts were preceded and accompanied by distinctive tremors. From June to September 1941, eruptions were almost continuous, except for some minor breaks of a few days. After September, the breaks became longer, though individual eruptions were commonly violent. No detailed record is available of the eruptions after 23 January 1942. However, reports indicate that the main eruptive period terminated about March 1942, though there is evidence of small later eruptions, probably in October 1943.

## INTRODUCTION

The eruption of Tavurvur, or Matupi, volcano that began on 6 June 1941, provides an interesting example of the phenomena that accompany the approach of an eruptive period, corresponding to a slow rise of the lava column in the conduit of the volcano. The build-up of this eruption really began with the 1937 Blanche Bay eruption (Fisher 1939) where the main activity took place at Vulcan, 6 km distant, on the southwestern side of Blanche Bay. Tavurvur is situated on the eastern side of Matupi Harbour, a subsidiary harbour on the eastern side of Blanche Bay (Fig. 1).

In 1937, a violent eruption at Vulcan built up a cone of pumice 226 m high and about 2 km in diameter at the base. An eruption at Tavurvur began when the Vulcan eruption had been in progress for 21 hours, and consisted of a violent steam explosion that blasted out parts of the super-structure of the volcano along a southwest-northeast fissure. Before 1937, Tavurvur had been in a quiet solfataric stage with water vapour and other gases, probably mostly carbon dioxide and sulphuretted hydrogen, gently escaping from several vents on and around the volcano. After 1937, Tavurvur did not return to this quiet solfataric condition, but continued to produce a large vapour cloud, which issued from fairly vigorous vents within the blasted-out main central crater, and commonly containing obvious sulphur dioxide.

Sources of information for this paper, in addition to the author's own recollections of direct observations, include: weekly volcanological reports for the period 1937 to 1942 prepared by the author, L.E. Clout, C.L. Knight, and L.C. Noakes, and published in the Rabaul Times; daily reports on the 1941 eruption from 6 June to 23 July by C.L. Knight; an incomplete unpublished description of the eruption by C.L. Knight written about 1946; and aerial photographs taken during and after the Second

World War. Pre-eruption seismograph records were lost in the war, but the results were recorded in the weekly volcanological reports.

## PRE-ERUPTION PHENOMENA

### Temperature

After the 1937 eruption a number of observation points were established at which temperatures were recorded weekly; these temperatures were normally 100°C or less. In the later part of August 1940, however, a new area of solfataric activity was noticed on the northern side of the 1878 crater, where only one or two patches of hot ground had previously existed, about 20 m~~etres~~ from the top of the steep eastern wall of the 1937 crater (Fig. 2). A temperature observation point was immediately established here, but the temperature of the gases stayed at 100°C until early November, when 110.8°C was recorded. The temperature remained about 110°C until 8 December, when a spectacular rise up to 224°C took place over a 10 day period (Fig. 3). By 7 January, 1941, the temperature had reached 260°C, and it continued to rise steadily but spasmodically until by 20 March the temperature had passed 300°C. From 15 April, the temperature rose more rapidly and by about 10 May had reached 380°C. The last temperature recorded, 3 days before the eruption started, was 392°C.

Only in the fumaroles within the 1878 crater were temperatures above 100°C recorded. The bottom of the 1937 crater (Fig. 2) was at that time extremely difficult to reach because of vertical walls exuding hot gases. When increased activity in the 1878 crater was first noticed, a somewhat precarious descent was made with the aid of a rope, and temperatures of the gases in the crater measured, but nothing above 100°C was recorded. The experience was not repeated, but no doubt the gases here, at a lower elevation and closer to the source of heat, also participated in the subsequent rise in temperature. Although there were signs of increased activity at other areas around the volcano, temperatures remained at 100°C, and there was no visible increase in activity in the solfataric areas along the Matupi Harbour waterfront.

Volume and distribution of gas emission

After the 1937 eruption emission of hot gases was confined to (1) the 1937 crater, mainly at the southwestern end and near the foot of the central lava plug (Fig. 2; see also Fisher 1939, p.36 & 62), (2) quietly steaming vents in the blow-out craterlet on the southwestern flank of Tavurvur, (3) patches of hot ground on the southeastern and northern outer slopes of the volcano, and (4) the solfataras at the western foot of the volcano along the harbour front (Fig. 2). The last two areas were not affected by the build-up to the 1941 eruption. In the 1937 crater, minor changes in the relative strength and even in the location of the various gas vents took place from time to time. For about a year after the 1937 eruption no significant change was recorded in the total volume of gas emitted, although there was possibly a slight decrease in the first few months, with minor variations thereafter. In June 1938 it was observed that a slight increase had taken place in the previous two months, and again, in October 1938, an increase in gas pressure and volume was noted. Thereafter, up to the time the eruption began, increases were frequently noted, but there was no way of measuring these quantitatively.

Until August 1940 the only signs of activity in the 1878 crater were one or two patches of hot ground in the northern and northeastern inner slope mentioned above. In the latter part of August it was noticed that the new series of quite vigorous vents which had developed in this area, were giving off a strong smell of sulphur dioxide. At the same time it was observed that in what had previously been just an area of hot ground on the outer southeastern slope of Tavurvur, a number of active fumaroles now existed, also emitting sulphur dioxide, but not as strongly as those in the crater area. Temperatures then and later did not exceed 100°C. Many vents developed along and near a gully running from this area off the southeastern slope of Tavurvur, and this became quite a large area of solfataric activity. New vents also appeared about this time (August-October 1940) on the southern and southwestern outer slopes, but these did not individually show much subsequent increase in activity, although white incrustations developed over a considerable area of the southern slope.

When, in November 1940, the temperatures of the gases in the 1878 crater fumaroles began to rise above 100°C, an increase in the vigour of the gas emission within the 1937 crater was noticeable, and the solfataric areas on the southeastern outer slope increased in size, extending onto the flat land at the foot of the mountain and giving off increasing amounts of sulphur dioxide, particularly from the higher vents. The active area in the 1878 crater continually increased in size, number of vents, vigour of gas emission, and amount of sulphur dioxide and other 'high temperature gases'. By 22 April 1941 the main fumarole, now at a temperature of about 350°C, was no longer accessible, but several others in the vicinity were at temperatures equally or nearly as high. At this time the production of gas within the 1937 crater was increasing and on 20 May the largest discharge was noted from the fumarole at the bottom of the central basalt plug, the subsequent eruption point. New vents were observed nearby on the southern wall of the 1937 crater. On 27 May, gas emission had further increased and was noted to be issuing with a 'roar like a train' and the steam cloud was visibly larger. This increase was maintained until the eruption began at 7.40 a.m. on 6 June.

Composition of gases

Before the 1937 eruption the only gas, apart from water vapour, that was evident in the solfataric discharges, was hydrogen sulphide. When conditions settled down after the eruption, again only hydrogen sulphide was noticeable in the gases. In September and October 1938, gas with a pungent odour and irritating to the eyes - probably hydrogen chloride - was noticed, and this was recorded again the following March when sulphur dioxide was noticed for the first time. From then on, sulphur dioxide was noticed more and more frequently, and it increased intermittently in quantity. By January 1940, the smell of sulphur dioxide was so strong that it was rarely possible to detect any hydrogen sulphide that might have been present in the gas, though it was recorded in May and again, for the last time, in the gases from the 1937 crater, in October 1940. Hydrogen chloride, on the other hand, was observed quite frequently.

No gas analyses were carried out of samples from the 1937 crater. Tests on the gas from the Matupi Harbour beach front solfataras indicated that the gas associated with the water vapour (the main constituent) consisted of carbon dioxide and hydrogen sulphide in the proportion of at least 5 to 1. No apparent change occurred in the composition of the gas from these solfataras at any time.

The fumaroles in the 1878 crater were rich in sulphur dioxide from the time of their inception in August 1940. Hydrogen chloride was noticed for the first time in early December, and the sulphur dioxide content continued to increase both from these vents and from those on the outer southeastern slope. As the temperature rose, so did the quantity of sulphur dioxide, hydrogen chloride, and other acid gases, and they became so corrosive that it was very difficult to maintain the pyrometers that were used for taking the temperatures. Mercury thermometers reading to 360°C had been obtained but the limit of these was passed in early May 1941.

Seismic effects

In 1940, seismographs were installed at the Rabaul Observatory, on the north side of Blanche Bay 7.2 km northwest of Tavurvur (Fig. 1), and recordings were maintained fairly continuously until January 1942. The instruments were locally built smoked-paper-recording horizontal seismographs with a magnification about 150:1. In early April 1941, a similar installation was completed at an Observation Post on the caldera wall 1 km north of Tavurvur crater (Fig. 2). The sensitivity of this was increased on 20 July.

Because of the low sensitivity of the seismographs and the gradual nature of the build-up towards the eruption, little significant information on seismic events was recorded before the eruption began. All identifiable events on the seismograms were earth tremors of tectonic origin. After the eruption began tremors of volcanic origin preceded each eruptive outburst by at least one hour, commonly by several hours, and each explosion also registered its signature on the seismograms.

Tiltmeters at the Observatory and at the observation post, reading to about 1 second of arc, were not sufficiently sensitive to register any tilting that may have occurred at Tavurvur before the 1941 eruption, although a tilt aggregating about 30 seconds of arc was recorded just before a severe regional earthquake of January 14, 1941 (Fisher 1944). An analysis of tide gauge records at Rabaul compared with tide markers that were read regularly every week indicated that Tavurvur sank gradually about 7.5 cm relative to Rabaul over the two years prior to the eruption.

Minor explosive activity

Although the catchment area of the Tavurvur main crater was small, after the 1937 eruption a certain amount of siltation occurred in the bottom of the crater, mainly from material washed down from the northwestern wall. Also, depending on the season, a pool of water up to about 2 m deep generally occupied the southwestern end of the crater (one of the main centres of gas emission). In February and March 1940 (at the end of the wet season) a series of small steam explosions took place in this part of the crater, which were attributed to the clearing of vents that had been blocked by mud and water. In early February fountains of muddy water a few metres high were observed, and in early March quite substantial explosions occurred. On 3 March there were four such explosions which hurled mud and small stones up to about the height of the crater rim. These explosions were followed by greater than usual emissions of water vapour and sulphur dioxide, and a geyser action developed in the muddy pool, with fountains up to 10-12 m high at 10-minute intervals. On 12 and 16 March even more severe explosions took place; mud, stones, and later dry dust were thrown out to a height of a hundred metres or more above the crater rim, depositing 15 to 20 cm of fine volcanic dust around the upper part of the crater. Hydrogen chloride was noted in the gases during the next few weeks, and on 17 and 18 May 1940 mild explosions similar to those on 3 March occurred, but these were the last recorded.

1941 ERUPTION

Although it provided at times spectacular pyrotechnic displays and, while the southeast season continued, was a great nuisance to the town of Rabaul because of the dust cloud swept across by the wind (Fig. 4), the 1941 eruption was not at any time of major proportions. There were no lava flows or anything resembling a nuée ardente; the largest individual explosion reached only a height of about 1200 m above sea-level ? (although dust was carried much higher by the wind and convection currents), and only a small crater was built up, in the northeastern part of the bottom of the 1937 crater (Fig. 2). The eruption point was near the base of the lava plug in the 1937 crater, and because the southern wall of the 1937 crater was practically vertical, it was possible, with care, to approach the southern side of the crater even when an eruption was in progress. The series of active fumaroles that occupied the 1878 crater before the eruption subsided completely soon after it began.

The general course of the eruption was an almost continuous period of eruptive activity for four months, with minor breaks, and a particularly active period during August and September 1941. After September the breaks between eruptive periods became longer, although some individual eruptions were more violent than any that had gone before. When the Japanese invasion force captured Rabaul on 23 January 1942 the volcano was in eruption, but it appears to have become quiescent about March the same year.

The first eruptive period lasted from 6 to 29 June and consisted of irregularly spaced explosions of varying intensity, the largest of which occurred on 17 June when ejecta were hurled beyond the foot of the outer slopes of the volcano. During this period, activity was more or less continuous for the first 16 days and intermittent for the next eight. For the first two days of the eruption only dust and rocks were ejected, but on the evening of the second day red hot rocks were thrown out. The initial eruption was followed by 'puffs' up to 60-120 m above sea-level (the crater?) ? at five-minute intervals. A 10-hour lull occurred after the eruption had been in progress for 14 hours, and 10 hours later came a large explosion to 300 m with red-hot rocks. The eruption continued irregularly, with strong explosions on 15 and 17 June. The tremor accompanying the latter was recorded at both the Observatory and the observation post; it produced the largest dust cloud up to that time, and rocks, including bombs up to 1 m in diameter, were hurled beyond the foot of the outer slopes of the volcano. Similar eruptions occurred over the next few days, but activity declined on 21 and 22 June and ceased from 2 a.m. on 24 to 7.40 p.m. on 26 June. In the 10 hours before the eruption resumed, eight large and many small local tremors were recorded on the seismograph at the observation post. At 1.15 p.m. on 27 and 4.25 p.m. on 28 June very violent explosions took place, rising to 1200 m above sea level with spectacular ejection of red-hot rocks. This eruptive period ceased at 8 p.m. on 29 June. By this time a crater wall about 20 m high had been built up to the west of the vent, which was 6-10 m wide.

Activity resumed on 3 July at 9. a.m. and was fairly continuous but not severe until 11 July. From then, until 17 July there was a marked reduction, and after a renewed burst of activity on 16 and 17, the volcano lapsed into quiescence. On 27 July eruptions took place between 1.30 and 2.00 p.m., but activity was very mild until 9 August, increasing gradually over the next two days, and inaugurating the longest and most continuous period of activity in the whole eruption. Billowing dust clouds rose every 3 or 4 minutes, and red-hot bombs were thrown out. Particularly violent explosions occurred on 22 and 25 August, and on 27 August it was observed that the central section of the crater was occupied by solidified lava. Water vapour and gases escaped through small vents in the lava and gave rise to a continuous roaring sound, which ceased when the lava was removed by an eruption late in the afternoon of 27 August. The week of eruption ending

27 August was easily the most violent since commencement of activity in June (Knight, vulcanological report for the week ending 27 August, 1941).

A short break in activity occurred between 30 August and 1 September, and eruptions continued, but generally at reduced severity, until 7 October. This break lasted until 22nd, when powerful explosions occurred, hurling hot rocks beyond the foot of the slopes of the volcano - still hot enough to set fire to dry grass 1 km from the crater. Another lull followed from 27 October to 8 November, and another from 13 to 15 November, followed by strong activity until 25 November with powerful explosive outbursts, some of which, on 19 November, hurled bombs the farthest yet. After 25 November, the volcano was quiet except for short bursts of activity in the afternoon of 4 December, and on 9 January 1942, until the week beginning 19 January, when another eruptive period began.

No record is available of details of the volcano's activity after the Japanese occupied Rabaul on 23rd January. However, the main eruption probably came to an end about March, 1942, although there are reports of some small later outbursts of activity. T. Kizawa (pers. comm., 1961) referred to an eruption in October 1943, and comparison of aerial photographs taken during or before March 1943 with those taken in March 1944 and later, show that in this period a small crater had developed at the northern end of the 1941-42 crater adjacent to the northern wall of the 1937 crater and the central basalt plug referred to earlier.

The whole course of the eruption is consistent with the slow rise of a small column of lava underneath the Tavurvur crater, activated in the first place by the major disturbance of the 1937 eruption, until the point of eruption balance was reached; thereafter explosive release of gases and fragmentation of the upper part of the lava column took place, with the formation of some pumice and glass, followed by a gradual settling down after the main pressure was released, with longer and longer periods between eruptions. Some of these later eruptions, however, must have been particularly violent because of the energy required to re-open the vent.

PETROLOGY

Petrological information on the products of the eruptions of Tavurvur is very limited. No comprehensive study has been made and the precise localities of some of the specimens examined has not been recorded.

C.L. Knight (in Fisher 1939, p 51) described specimens from the central basalt plug and from the ejecta of the 1878 eruption. Heming (1974) referred to the Tavurvur rocks and gave a chemical analysis of a glassy andesite bomb from Tavurvur. Miyake and Sugiura (1953) recorded chemical analyses of three samples from Tavurvur; these analyses, while very similar to each other, are markedly different from the one recorded by Heming (e.g. silica content 47.82 percent compared to Heming's 62.26 percent). In the absence of detailed information about the location and nature of the samples it is not possible to comment further on these differences, although it might be expected that the composition of the bombs thrown out by Tavurvur in 1941-42 would be similar in composition to the material of the 1878 eruption. Further study is needed of the detailed petrology and chemistry of the Tavurvur lavas and ejecta.

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## FIGURE CAPTIONS

Fig. 1. Locality plan.

Fig. 2. Aerial photograph of Tavurvur taken 17 March, 1943. A: 1878 crater. B: 1937 crater. C: 1941-42 crater. D: south-eastern solfataric area 1941. E: solfataras on northern slope. F: solfataric areas along Matupi Harbour beach front. G: central lava plug. H: Matupi observation post.

Fig. 3. Pre-eruption rise of temperature in the fumaroles in the 1878 crater.

Fig. 4. Tavurvur in eruption sometime during the southeast season (about April to November) of 1941. Photograph taken from Matupi Harbour looking east-northeast, and showing South Daughter volcano in right background. The southeasterly wind is blowing the eruption dust-cloud towards Rabaul.

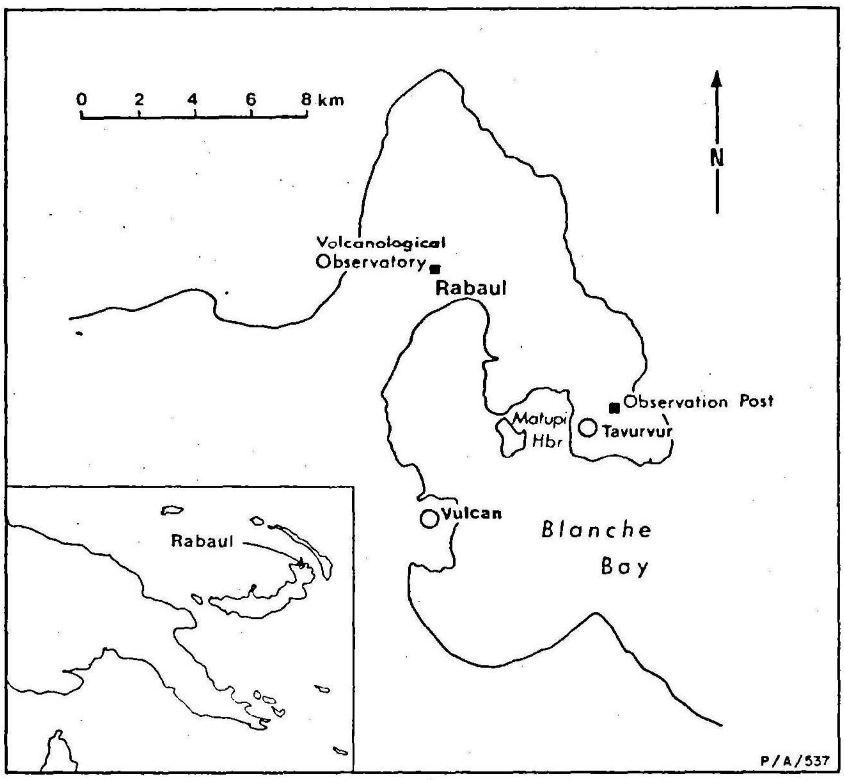


Figure 1.

Locality Plan

Fig. 2

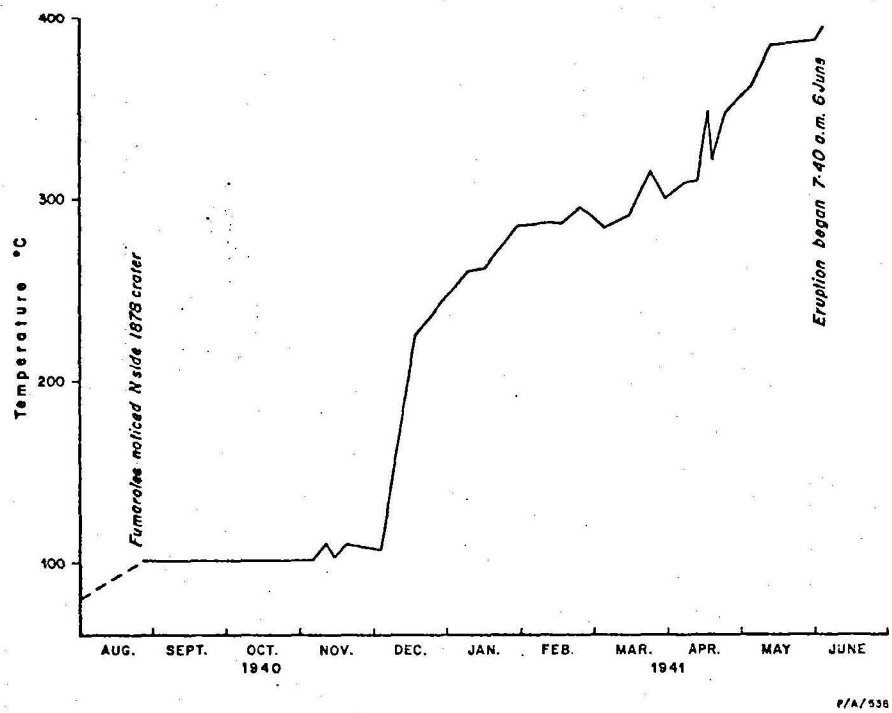


Figure 3.

Pre-eruption rise of temperature in the fumaroles in the 1878 Crater.

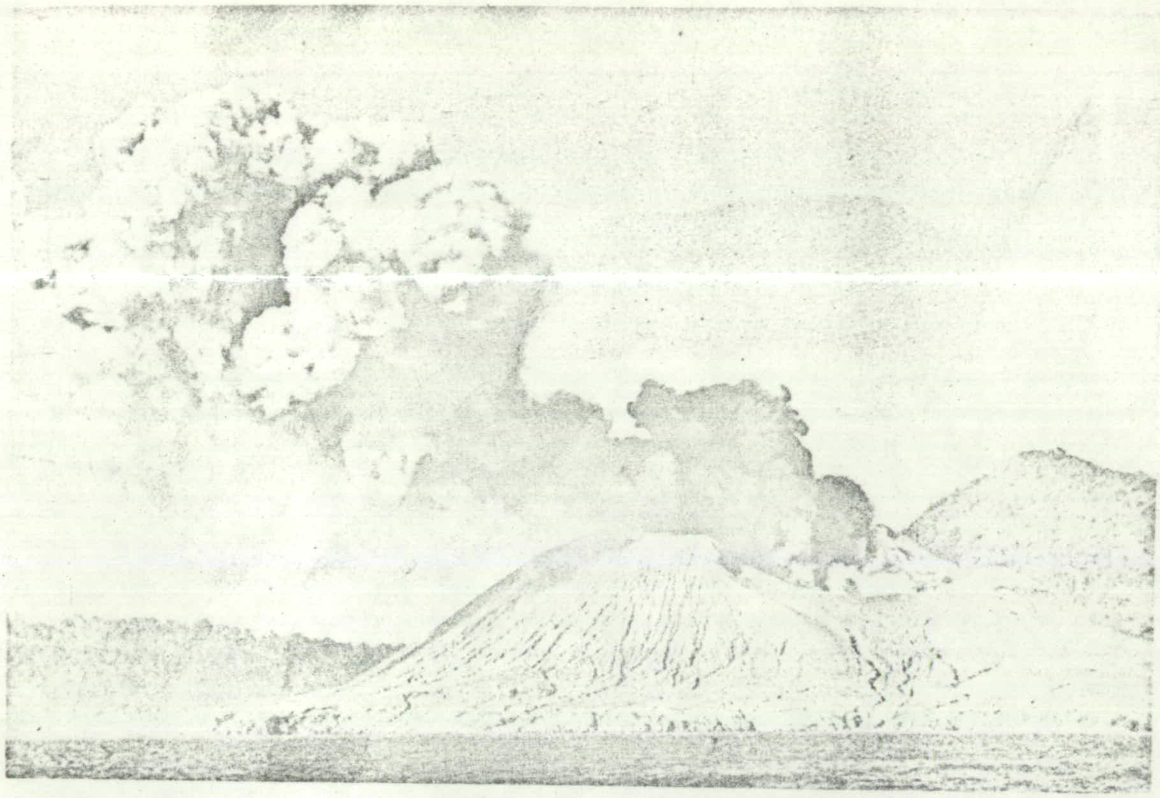

Figure 4. Tavurvur in eruption sometime during the southeast season (about April to November) of 1941. Photograph taken from Matupi Harbour, looking east-northeast, and showing South Daughter volcano in right background. The southeasterly wind is blowing the eruption dust-cloud towards Rabaul.

AERIAL THERMAL INFRARED SURVEY, RABAUL AREA, PAPUA NEW GUINEA, 1973

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ABSTRACT

An airborne thermal infrared survey of the Rabaul caldera was carried out in May 1973 to try to detect previously unknown hot spots, both on land and within the waters of the harbour. All imagery was acquired before sunrise at wavelength intervals of 3.5 - 5.5 micrometres and 8 - 14 micrometres (µm). Both sets of imagery showed the location and extent of all known thermal areas, but no new areas were discovered. Because of its greater contrast, the longer wavelength displayed the details of the terrain better than the shorter wavelength record. An airborne radiometer provided general information on the temperatures of sea and land, its broad field of view (2 degrees) combined with its response time (50 millisec) resulting in averaging of the surface temperatures of the small thermal areas.

## INTRODUCTION

The town of Rabaul, on the northeastern end of the island of New Britain, New Guinea, is built on the shores of Simpson Harbour (Fig. 1), within a breached volcanic caldera.

The latitude is 4 degrees south, and the climate humid tropical. The Fig 1. northwest monsoon season generally begins in December and continues to March, followed by a period of relatively calm weather until the south-east season starts in June. The average annual rainfall is approximately 2000 mm.

Two important volcanic centres in the caldera have been active in recent times. Vulcan erupted in 1937 with simultaneous minor activity at Tavurvur, and Taburvur erupted again between 1941 and 1943 (Fisher, this volume; Taylor, 1973). A volcanological observatory was established on the caldera rim in 1939, when a program of volcanic surveillance was begun (Fisher, 1940). During the Second World War the observatory was destroyed, but it was ~~rebuilt~~ reestablished in ~~1952~~ 1950, and since then the surveillance instrumentation has been undergoing development and is now of a high standard (Taylor, 1973)*. The temperatures of vapours from fumaroles and water from thermal springs in the vicinity of Tavurvur are recorded each week. The thermal infrared survey was undertaken to find out whether previously unknown hot spots could be detected, particularly springs that might issue from the sea-bed within Simpson Harbour and the shallower waters of Blanche Bay.

## KNOWN THERMAL ACTIVITY

Thermal activity within the Rabaul caldera is confined to mild fumarolic emission in the craters and on the slopes of Tavurvur and Rabalankaia and to thermal waters which flow into the bay mostly from areas adjacent to and below the high tide mark along the shoreline of Matupi Fig2 Harbour, the northern shoreline of Vulcan and in Sulphur Creek (Fig. 2).

Fumarolic emission in Rabalankaia crater is confined mainly to two concentric rings open to the southeast which are probably related to ring

*See Myers (this volume). Ed.

fractures. The outer ring of fumaroles coincides with the inner slopes and base of the crater wall whereas the smaller circle of fumaroles lies centrally on the flat floor of the crater. Heming (1973) suggested that the smaller circle of fumaroles may have been formed along a fault around a collapsed central plug. Two small thermal springs, which flow intermittently, are located in a gully half way up the western slope of Tavurvur. Minor gas ebullition occurs in several thermal springs along the northern shoreline of Matupi Harbour.

Over the past twenty years fumarolic temperatures have rarely risen much above 100°C. Thermographs of fumarolic temperatures show little variation except for one small fumarole in one of Tavurvur's craters which exhibits a regular rise and fall of as much as 37°C over a twelve hour period, a phenomenon which may be related to tidal rhythms. Heavy rainfall can produce sharp falls in fumarolic temperatures but recovery to original levels occurs within several hours of cessation of rain. This factor did not influence infrared imaging of the fumarolic areas as no rain fell for at least 48 hours before imaging commenced or during the period imaging took place. Temperatures of the fumaroles at the time of imaging ranged from 87°C to 99°C with the majority of fumaroles recording a temperature of 99°C.

Shoreline thermal spring temperatures fluctuate widely depending on tidal conditions and on the amount of recent rainfall and consequent height of the water table.

INFRARED SURVEY

The area was surveyed with a Daedalus optical-mechanical infrared line scanner operating in both the 8 – 14 μm and the 3.5 – 5.5 μm intervals. The detectors had an instantaneous field of view of 2.5 milliradians, and a thermal resolution of 0.3°C (C. Ellyett, pers. comm.). Their

* For details the reader is referred to an unpublished account by Perry and Crick in Bureau of Mineral Resources Record 1974/97, "Aerial thermal infrared survey, Rabaul, Papua New Guinea, 1973."

output was recorded on magnetic tape and the data later converted to film records. From the survey altitude of 914 m (3000 ft) the detectors viewed a square of terrain 2.28 m (7.5 ft) on a side at any instant. Preferred flying time was pre-dawn to avoid the daily heating effect of the sun, so that thermal anomalies would show up most strongly against the background, but local restrictions limited flying to the period between first light and sunrise. The scanner was fitted with a roll stabilizer that provided electronic compensation for aircraft roll to the scanner imagery by reference to a gyroscope.

The aircraft, an Aero Commander 680F, also carried a Barnes PRT 5 radiometer with a field of view of 2 degrees which recorded surface temperatures with a specified accuracy of $\pm$ 0.5$^{\circ}$C on a strip chart. During daylight hours the location of the radiometer trace could be determined by reference to photographs taken with a 35 mm tracking camera. Because flying took place after first light, navigation was by visual reference to ground features, and the coastlines in the area made this a relatively simple task. A Doppler navigation system did not provide useful information from most survey runs because they were partly over land and partly over calm water. During the period of the infrared survey (May 3 - 11, 1973) 1:15 000 scale vertical aerial colour photographs of the caldera were obtained for comparison with the infrared imagery.

## CORRELATIONS BETWEEN THERMAL AREAS AND INFRARED ANOMALIES

The grey tones of the imagery (Figs 3 and 4) give an indication of relative temperature of the terrain. The hotter areas are light toned and the cooler areas relatively dark.

### 8–14 μm band (Fig. 3)

The most noticeable features of the imagery are light and dark bands at right angles to the flight lines, which are caused by an instrumental effect due to amplification of the detector output practically to the maximum to extract as much information as possible on temperature variations of the water surface (C. Ellyett, pers. comm.). Neglecting these it can be seen that the sea is generally warmer than the land at this time of day.

Fig 3

Bright areas within the water, principally in Matupi Harbour, are caused by hot water from thermal springs. The main activity is at the head of Matupi Harbour, with other hot zones at the southeast end of the airstrip, and west of Tavurvur; the water in Sulphur Creek is slightly warmer than the harbour water, and north of Vulcan Crater a small body of water isolated from the harbour by a narrow sand bar is also warmer than the harbour water.

On the land Rabalankaia is outlined by concentric bright rings (Fig. ~~6~~ 5) that can be seen on the aerial photographs to correspond with areas of bare ground on the crater floor and walls. It is concluded that the bright returns on the imagery are produced by warm ground. Similarly, small areas of bare ground between Rabalankaia and Matupi Harbour show as hot spots on the imagery.

The bright, rather complicated pattern at Tavurvur corresponds in part to bare ground, but exceptions are dark toned bare areas that image cool, for example the northeast slope and the upper southwest flank. Bare areas on the lower southwest flank are warm in only a few places. Also imaging cool are the floors of the four individual craters within the main crater rim, and the floor of one small parasitic cone on the southwest flank of Tavurvur.

μ

<u>3.5–5.5 μm band (Fig. 4)</u>

In general the contrast is less in this band than in the 8–14 μm interval, and as a result details of the terrain are less apparent. The obvious features of the imagery are the dark zones along one side of each strip which were flown from south to north. Four runs not shown on the figure were flown round the coastline from west to east, and it was apparent that

Fig 4

the dark zone remained on the starboard side of each strip irrespective of aircraft heading, therefore it is concluded that the dark zones are caused by some instrumental factor.

All the 'hot spots' visible on the long wavelength record can be seen on the original 3.5-5.5 µm imagery, but on the latter the details of the background are less well shown because of the lack of contrast.

µ

## Radiometer

It is evident from a comparison of radiometer readings with the infrared imagery of Tavurvur ~~[illegible]~~ and Rabalankaia (Fig. ~~6~~ 5), that the recorded temperature variations are less than one would expect from the tonal variations on the imagery. These differences are ascribed to the larger field of view of the radiometer (nearly 35 milliradians as compared to the scanner's 2.5 milliradians) and to its response time (50 millisec), and it is concluded that in surveys of this type the radiometer should have a field of view comparable with that of the scanner, or alternatively a calibrated thermal scanner should be used.

Fig 5 (see note on page 10)

## OTHER USES FOR AERIAL INFRARED TECHNIQUES IN PAPUA NEW GUINEA

Airborne infrared imaging systems have been used in the thermal surveillance of more than 23 volcanoes around the world (Friedman and Williams, 1968; Moxham, 1971). Use of such systems over volcanoes in Papua New Guinea could provide information on the location and extent of thermal areas on active and dormant volcanoes to facilitate ground surveillance studies, and in the case of calibrated systems could give an estimate of temperatures. Infrared imaging could prove valuable in monitoring rapid changes in thermal activity, particularly effusive eruptions. If a calibrated system were used, volcanic thermal energy yield and partition estimates could be determined in conjunction with volcanologic ground observations.

Prediction of an impending eruption using airborne infrared imaging systems is possible but not yet demonstrated. Thermal activity increased prior to the eruption of Tavurvur in 1941 (Fisher, this volume) and at some other volcanoes around the world (Moxham, 1971). Regular flights would need to be maintained over the volcano in question to detect changes in thermal activity using preferably a calibrated device, and the results analysed in conjunction with other surveillance techniques.

A potential major use for infrared imaging in Papua New Guinea is in the mapping of possible sources of geothermal power. Several possible sites exist, and although thermal activity in these areas is usually indicated on normal aerial photography by the lack of vegetation, in some places, as for example, along the Num River, Ambitle Island (off New Ireland; see Johnson et al., this volume) thermal activity occurs along stream and river courses and may be overlooked in the initial mapping if normal aerial photography is used.

## CONCLUSIONS

The thermal infrared scanner successfully mapped the location and extent of most of the known thermal areas with the exception of a few springs below high water level, and it is concluded that there are no other major 'hot spots' within the surveyed area. Of the two infrared records, the 8–14 μm imagery is preferred for this sort of general volcanological work, because of its better contrast, although had only the 3.5–5.5 μm detector been available the mapping of thermal areas could still have been satisfactorily achieved.

The radiometer contributed little useful information to the study of the thermal areas; in the writers' opinion the use of a radiometer of this type in similar aerial surveys in future would be worthwhile only in situations in which the thermal areas are larger than the instantaneous field of view of the instrument at the operational altitude. Small or patchy fumarolic areas commonly found on slopes and in craters of volcanoes would not be suitable, but the temperature of large areas of warm surface water produced by subaqueous thermal activity could probably be accurately measured.

ACKNOWLEDGEMENTS

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## CAPTIONS FOR FIGURES AND DIAGRAMS

Figure 1. Rabaul and environs

M Figure 2. Thermal anomalies of Matupi Harbour from infrared imagery (8 - 14 µm)

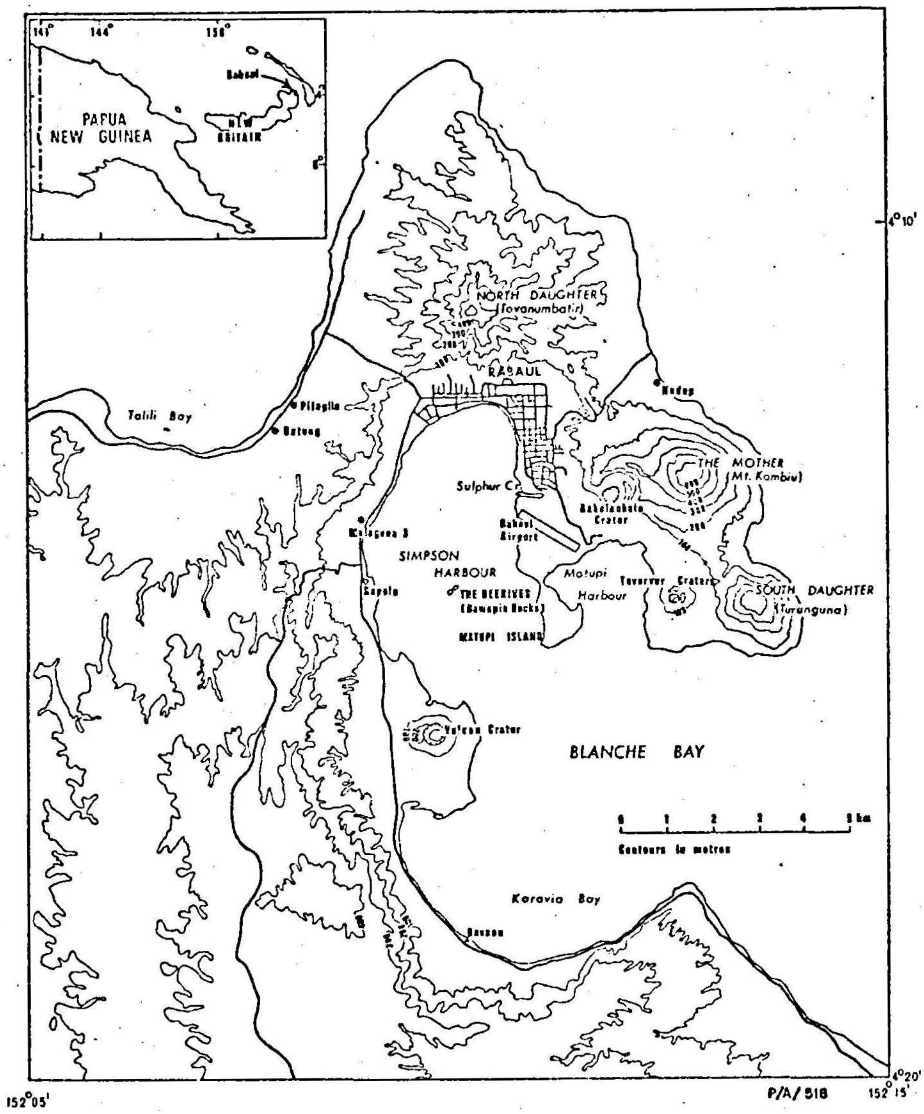
M Figure 3. Mosaic of thermal infrared runs, 8 - 14 µm, Rabaul Caldera

M Figure 4. Mosaic of thermal infrared runs, 3.5 - 5.5 µm, Rabaul Caldera

Figure 5. Infrared imagery (8 - 14 µm) of Tavurvur (upper) and Rabalanakaia (lower) matched with radiometer traces. Both runs flown north-south.

Please ensure Figure 5 occupies a right-hand page, and that its caption is at the bottom of of the opposite left-hand page

Fig. 1


PAPUA
NEW GUINEA
NEW BRITAIN
NORTH DAUGHTER
(Tovanumbatir)
RABAUL
Talili Bay
THE MOTHER
(Mt. Kombiu)
Sulphur Cr.
SIMPSON
HARBOUR
Motupi
Harbour
Tavurvur Crater
SOUTH DAUGHTER
(Turanguna)
THE BEEHIVES
(Dawapia Rocks)
MATUPI ISLAND
Rabaul
Airport
Vulcan Crater
BLANCHE BAY
Contours in metres
Karavia Bay
4°10'
4°20'
152°05'
152°15'
P/A/518

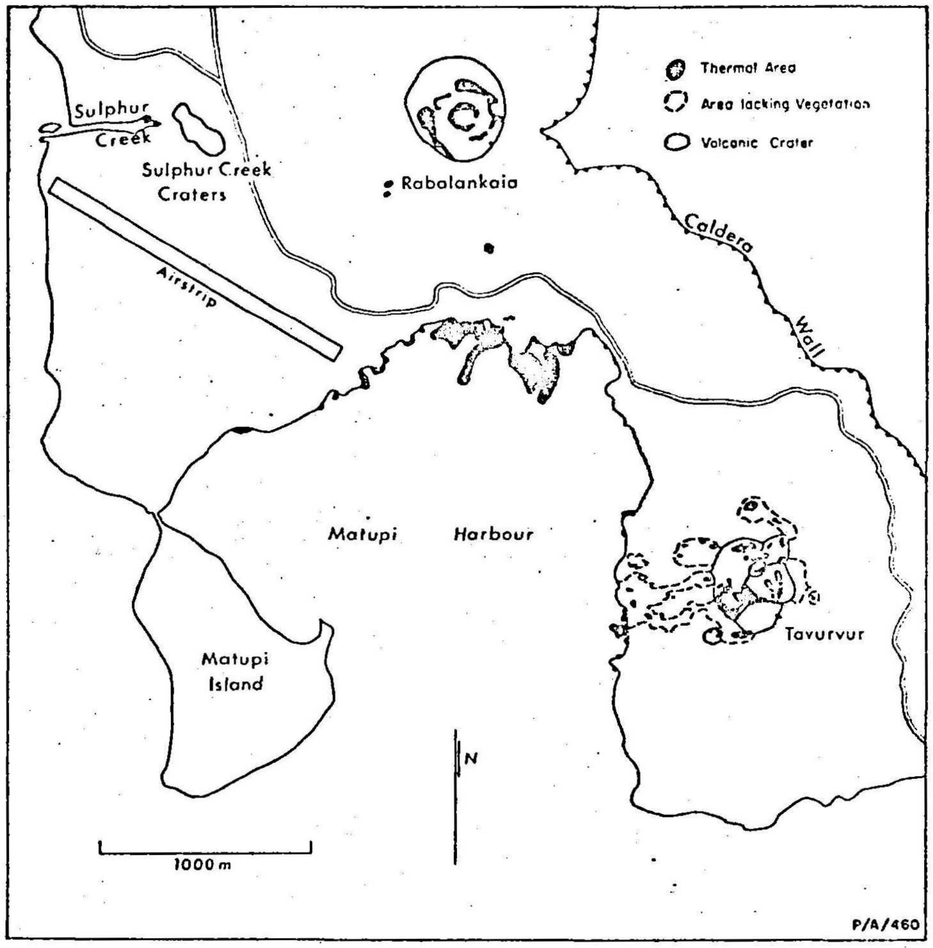

Thermal Area
Area lacking Vegetation
Volcanic Crater
Sulphur Creek
Sulphur Creek Craters
Rabalankaia
Caldera Wall
Airstrip
Matupi Harbour
Tavurvur
Matupi Island
N
1000 m
P/A/460

Fig. 2

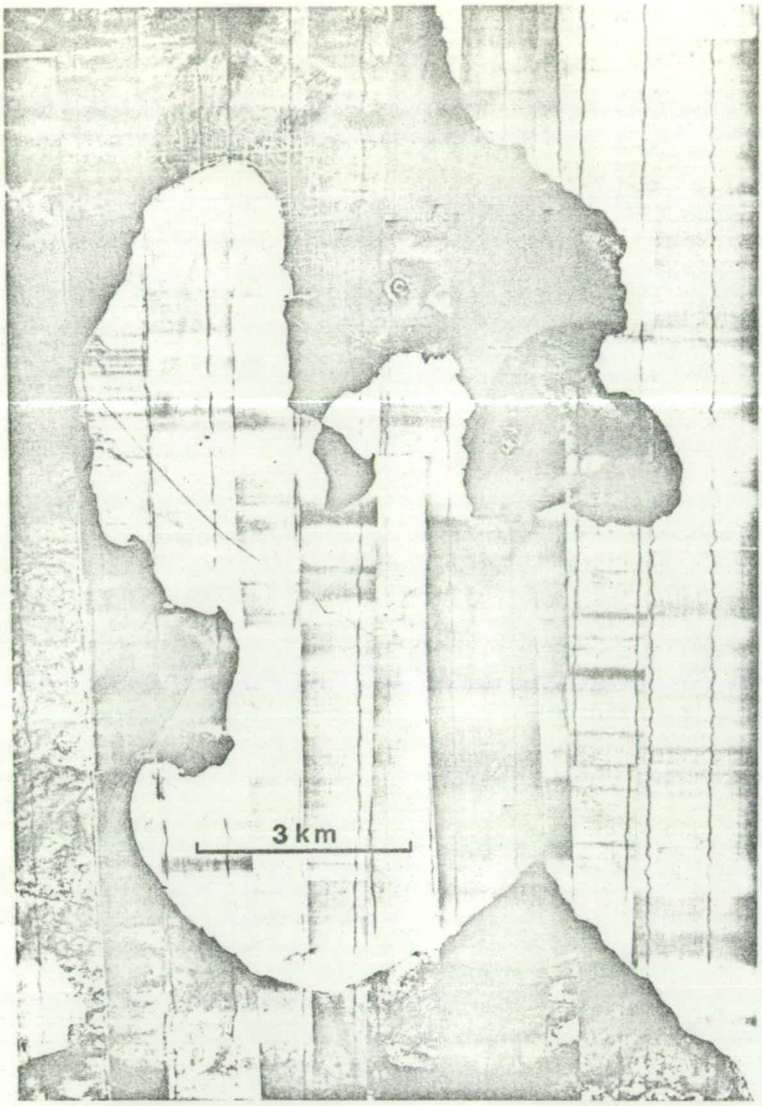



Fig 3

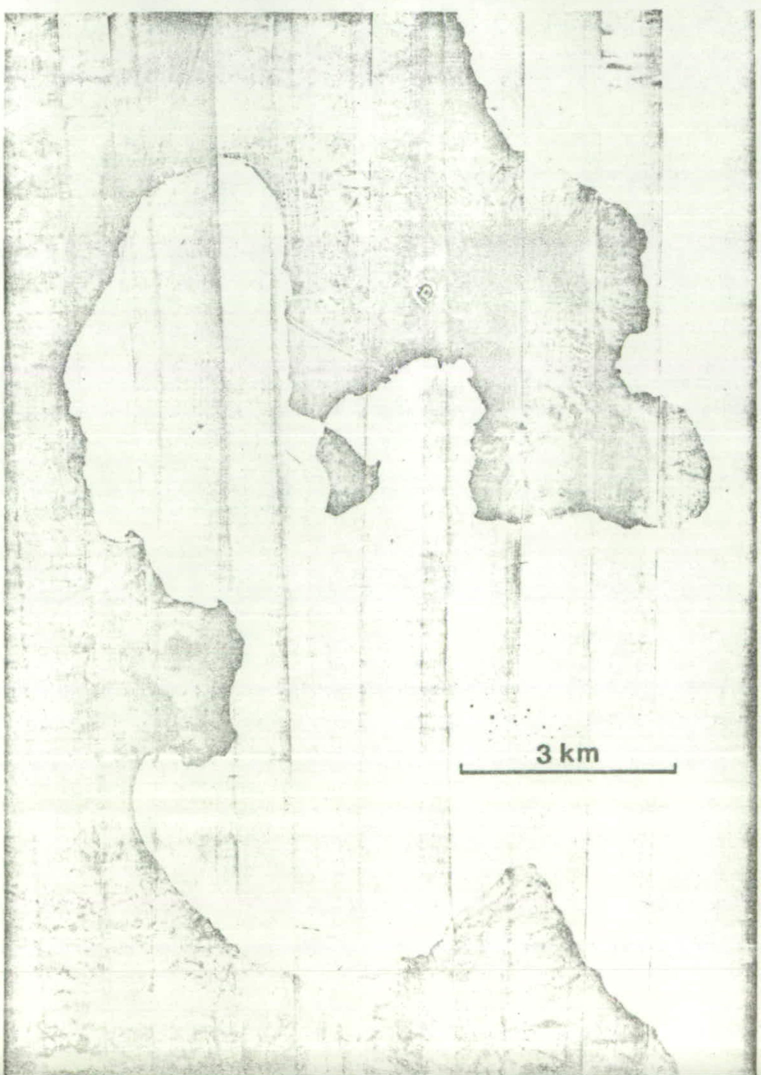
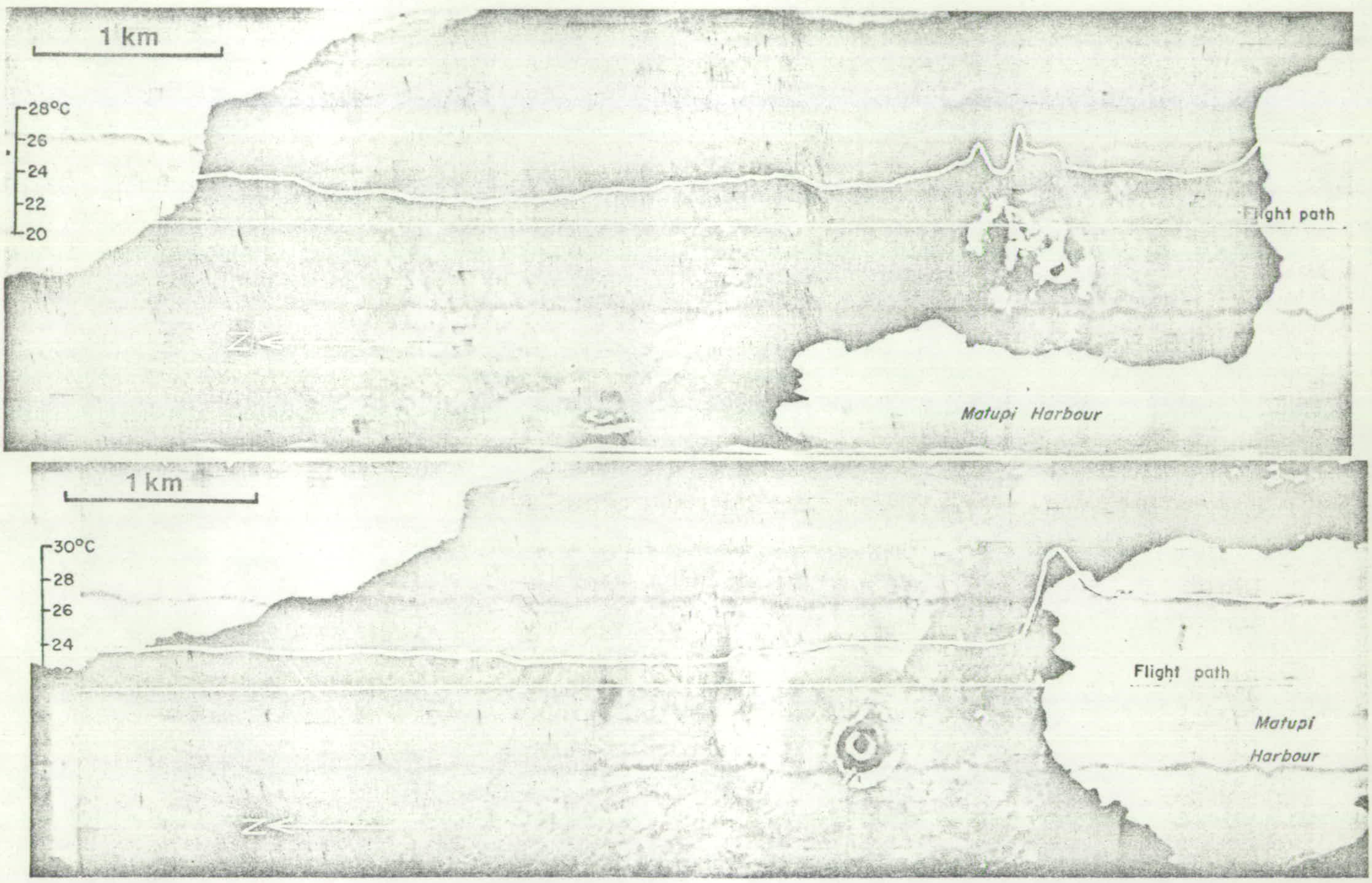



Fig 4.

Fig 5



1 km
28°C
26
24
22
20
Flight path
Matupi Harbour
1 km
30°C
28
26
24
Flight path
Matupi
Harbour

LATE CAINOZOIC VOLCANISM AND PLATE TECTONICS AT THE SOUTHERN MARGIN OF THE BISMARCK SEA, PAPUA NEW GUINEA.

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ABSTRACT

Two Late Cainozoic volcanic arcs can be recognised at the southern margin of the Bismarck Sea, Papua New Guinea. Both arcs provide striking examples of the geodynamic complexity to be expected in regions characterised by small plates whose instantaneous poles of rotation are nearby. A western arc is associated with the boundary between the South Bismarck and Indo-Australian plates. The chemical compositions of its volcanic rocks change along the arc - i.e., in a direction parallel to the strike of a postulated subducted lithospheric slab. These changes can be explained by identifying Late Cainozoic poles of rotation in the north-western part of mainland Papua New Guinea, and by postulating eastwardly increasing rates of plate convergence. An eastern volcanic arc is associated with the boundary between the South Bismarck and Solomon Sea plates. The volcanoes are arranged in an unusual zig-zag pattern, and the compositions of the volcanic rocks change with increasing depths to the northward dipping New Britain Benioff zone - i.e., in a direction at right angles to the strike of the Benioff zone, and to the axis of the New Britain submarine trench. The existence of a thrust slice in the northwestern corner of the Solomon Sea is postulated to account for the distribution pattern of the eastern-arc volcanoes.

#13

## INTRODUCTION

Petrological studies of island-arc volcanoes have been strongly influenced by the theory of plate tectonics which proposes that lithospheric plates are subducted beneath the arcs (e.g. Wyllie, 1973). Recent studies favour the concept that island-arc magmas are formed above the downgoing slabs by partial melting of upper mantle peridotite under hydrous conditions. Partial melting of the peridotite is thought to take place through access of volatiles produced by dehydration of the crustal portions of the slabs, or by volatile-rich siliceous melts generated by partial fusion of the slabs (e.g. McBirney, 1969; Green and Ringwood, 1972; Wyllie, 1973; Stern and Wyllie, 1973; Nicholls and Ringwood, 1973; Ringwood, 1974). An understanding of the occurrence, periodicity, and chemistry of volcanism in any island-arc would therefore appear to be critically dependent upon the formulation of a geodynamic plate model.

This paper proposes a plate model for the 1000 km-long chain of volcanoes that border the southern margin of the Bismarck Sea (Fig. 1). It summarises conclusions reached in another report, which includes a more complete presentation and discussion of petrological data (Johnson, in prep.)

Fig 1

On a global scale, Papua New Guinea straddles part of the boundary between the Indo-Australian and Pacific plates (Le Pichon, 1968, 1970; Denham, 1973). In detail, however the configurations of plate boundaries are more complicated, and three independent studies suggested that the Late Cainozoic geodynamics of the region has been greatly influenced by at least two minor plates, trapped between the two larger ones (Johnson and Molnar, 1972; Curtis, 1973; Krause, 1973).

## VOLCANO DISTRIBUTION

The Late Cainozoic volcanoes of the south Bismarck Sea region are associated with the southern margin of the well-defined minor plate that underlies the southern part of the Bismarck Sea. Following the terminology of Johnson and Molnar (1973), this plate will be called "South Bismarck" (Fig. 2).

Its southern margin is adjacent to the Indo-Australian plate in the west, and to the "Solomon Sea" plate in the east. The south Bismarck Sea volcanoes are therefore associated with two plate boundaries, and although it is convenient to refer to all the volcanoes as the "Bismarck volcanic arc" (e.g. Johnson et al., 1973; Cooke et al., this volume), it is proposed here that two volcanic arcs should be recognised - each one corresponding to a different plate boundary, and to different geodynamic events during the Late Cainozoic. On the basis of chemical composition (see later) and distribution, the two volcanic arcs are:

Fig 2

1. A western arc, extending from the Schouten Islands, in the west, to the Cape Gloucester area of west New Britain, in the east (Fig. 3). These volcanoes are considered to be associated with a contact zone between the South Bismarck and Indo-Australian plates, whose nature is discussed below.

2. An eastern arc, which includes the volcanoes of the Witu Islands, Willaumez Peninsula, and north coast of central New Britain (Fig. 4). Discussion below proposes that these volcanoes are associated with the South Bismarck/Solomon Sea plate boundary although this apparently simple correlation is shown to be complicated by a possible thrust slice in the northwestern corner of the Solomon Sea.

The volcanoes of the Rabaul area (Fig. 1) may also be associated with the South Bismarck/Solomon Sea plate boundary, but they are considered to be distinct from the eastern arc and are therefore not considered here (see Heming, 1974). Neither does this paper deal with the volcanic centres of Andewa and Schrader in west New Britain (Fig. 4), and with Uvo in the Adelbert Range (Fig. 3), as they appear to be older (at least in part) than the other volcanoes, and are not sufficiently well surveyed. It is emphasised that a study of the petrology and ages of rocks from Andewa, Schrader, and Uvo could necessitate some revision to the geodynamic interpretations given below.

## GEOLOGY OF THE WESTERN ARC

### Description

Fig 3

The volcanoes of the western arc form an extremely narrow zone (Fig. 3). The volcanoes are close together at the eastern end of the arc, but further west they are more widely separated: the largest gap is 120 km - between Manam and Karkar Islands. In addition, a comparison of the volumes of the volcanoes shows that the Schouten Islands, in the extreme west, are smaller than most other western arc volcanoes. These relationships suggest therefore that, in broad terms, the volume of volcanic rocks decreases from east to west along the western arc.

Three other features shown in Figure 3 are also noteworthy: (1) the volcanoes of the Schouten Islands form a chain which is off-set about 25 km northeastwards from the extension of the line that can be drawn between Aris, Manam, Karkar, and Bagabag Islands; (2) Long Island, Crown Island and Hankow Reef, and Tolokiwa Island and the long axis of Umboi Island, constitute two parallel lines which intersect the general trend of the western arc at about $30^{\circ}$; and (3) in the Cape Gloucester area of west New Britain, Tangi, Aimaga, and Talawe volcanoes form a north-south line. As shown in Figure 3, six subaerial volcanoes and one submarine centre, have had eruptions which have been recorded by literate observers during the last one-hundred years.

About 500 rocks from the western arc volcanoes have been studied in thin section by the writer. In addition, 143 major element chemical analyses have been obtained, including those published by Morgan (1966). These data indicate that the most common rock types in the western arc are hypersthene-normative basalts and low-silica andesites (containing less than 57wt% $SiO_2$). High-silica andesites are also present, but dacites (~~more than~~ $>$ 62% $SiO_2$) are rare, and rhyolites (~~more than~~ $>$ 70% $SiO_2$) have not been found. Basalts appear to be absent from the Schouten Islands.

The chemical analyses show that pronounced changes in chemistry take place along the western arc. For example, in Figure 5, alkali contents are shown to increase from zone A, in the extreme west, to zone B, comparing rocks of the same silica content; they increase abruptly to maximum values in the rocks of zone C, but decrease in those of zone D most of which, nevertheless, contain higher total-alkali contents than most zone B rocks. In other words, the lowest total-alkali contents are shown by rocks from the Schouten Islands (comparing rocks with the same silica content), and the highest values are those of rocks from Long and Tolokiwa Islands. Similar maxima and minima are shown when the following oxides (and oxide ratio) are plotted against silica - $K_2O$, $Na_2O$, $K_2O/Na_2O$, $TiO_2$, $P_2O_5$.

The compositions of these rocks are the same as those of many hypersthene-normative volcanic rock suites associated with active island-arcs in many parts of the world. By analogy, and because of the high frequency of associated intermediate-focus earthquakes, it can be inferred that the western arc has developed at a convergent plate boundary where subduction has taken place. The supporting evidence, however, is incomplete as a Benioff-zone-like feature is associated with only part of the arc, and because a submarine trench appears to be absent.

Studies of intermediate-depth earthquakes recorded during the last 10 to 20 years show: a pronounced concentration of epicentres in the strait between Long Island and the north coast of mainland Papua New Guinea; a decrease in the density of these epicentres to the east and west of Long Island; and an apparent absence of intermediate earthquakes beneath the volcanoes west of Karkar Island (e.g. Denham, 1969; Johnson et al., 1971; Curtis, 1973). The intermediate-depth earthquake foci east of Karkar Island define a steeply inclined northward-dipping seismic zone, which extends from depths of about 120 km down to more than 200 km, and whose dip increases down-dip to more than $80^{\circ}$ beneath the arc. Focal mechanism solutions for a few of these earthquakes give steep, dip-slip underthrust senses of motion (see Ripper, in press).

Many shallow- and intermediate-depth earthquakes take place beneath the coastal ranges of mainland Papua New Guinea to the south but these do not appear to define a Benioff zone (Johnson et al., 1971). Rare intermediate-focus earthquakes beneath the Highlands-Wau region (Fig. 1) suggest the presence of lithospheric material at these depths, but records are required for a much longer time span before these earthquakes become sufficiently numerous to clearly define any possible inclined seismic zone (see also Fig. 6 and discussion below). There is a pronounced deficiency of intermediate-depth events beneath the Adelbert Range, opposite the gap in the volcanic arc between Manam and Karkar Islands (R.J.S. Cooke, pers. comm., 1974).

In contrast to the eastern arc, which is associated with the New Britain submarine trench (see below), the available bathymetry suggests that no trench-like feature is associated with the western arc. Unpublished seismic reflection profiles obtained by the Bureau of Mineral Resources in 1970 revealed the presence of extremely thick sedimentary accumulations between the western arc and the north coast of the mainland and north of the western arc, but it is doubtful if either of these can be regarded as an infilled submarine trench (see also Connelly, 1974).

## Discussion

Dewey and Bird (1970) described the mainland of Papua New Guinea south of the western arc as a type example of a region where a continent had collided with an island-arc. They suggested that this collision caused a subduction zone to change its polarity from an original northward dip beneath the island-arc, to a southward one - i.e., the Benioff zone "flipped" (cf. McKenzie, 1969). Johnson and Molnar (1972), Karig (1972), and Hamilton (1972) all supported this interpretation. The geology of the region lends support to the interpretation that the Adelbert and Finisterre Ranges and Huon Peninsula (i.e. the coastal ranges, Figs 1 and 2) represent a welt of island-arc rocks on the northern edge of the continental mass beneath the Highlands of Papua New Guinea (Thompson and Fisher, 1957; BMR, 1972; Bain, 1973; Robinson et al., 1974). However, there is little support for the concept of a reversal of arc polarity.

Firstly, as discussed above, it has yet to be clearly established that mainland Papua New Guinea is underlaid by a single, southward-dipping Benioff zone at the present day. In addition, it is unlikely that a southward-dipping seismic zone existed beneath the coastal ranges and Ramu-Markham Valleys (Figs 1 and 2) in earlier parts of the Quaternary. World-wide observations of many island-arcs and continental margins indicate that volcanoes overlie those parts of inclined Benioff zones which are deeper than 70-100 km (e.g. Dickinson, 1970). If a southward-dipping Benioff zone existed beneath the northern part of mainland Papua New Guinea, Quaternary volcanoes would be expected in, and immediately south of, the Ramu and Markham Valleys, and not in a narrow zone north of the coastal ranges. (The Quaternary volcanoes of the Highlands region, Fig. 1, do not appear to be directly related to a downgoing slab - see Mackenzie, this volume).

The claim of Johnson and Molnar (1972), Curtis (1973), and Krause (1973) that the Ramu and Markham Valleys represent the present-day boundary between the Indo-Australian and South Bismarck plates is also open to question. Firstly, there is no concentration of earthquakes beneath the valleys. Rather, the epicentres are spread more or less evenly over a wide zone covering the coastal ranges, the Ramu and Markham Valleys, and the northern foothills of the Highlands region (see, for example, Denham, 1969). Secondly, extensive Holocene faulting has not been reported from the Ramu and Markham Valleys, and there is no evidence that Holocene faults are more common in these valleys than they are in other parts of the wide zone of present-day earthquake activity (BMR, 1972). The Ramu and Markham Valleys may represent the site of a former plate boundary which was destroyed by a continent/island-arc collision (see below), but the seismic and geological evidence does not demonstrate that the valleys necessarily represent a present-day plate boundary.

## GEOLOGY OF THE EASTERN ARC

### Description

Unlike the western arc, the eastern arc of the New Britain region appears, on first inspection, to be part of the more familiar kind of island-arc. The arcuate mountainous axis of New Britain island is a geanticlinal ridge composed mainly of Tertiary volcanic and intrusive rocks and limestones similar in type and age to those of the mainland coastal ranges (Ryburn et al., in prep.; BMR, 1972). A submarine trench showing a maximum depth of more than 8000 m parallels the island off the south coast (Fig. 1). A Benioff zone dips northwards at about 70° beneath New Britain, and shows earthquake foci up to 565 km deep (Denham, 1969; Johnson et al., 1971; Curtis, 1973); the Quaternary volcanoes have been built up on the northern concave side of the island over the deeper parts of the Benioff zone (Fig. 4). The thickness of crust beneath New Britain ranges from about 25 km in the north to about 45 km along the south central coast (Finlayson and Cull, 1973; Wiebenga, 1973).

Fig 4

Despite these familiar island-arc characteristics, two features of the geology are unusual. Firstly, the distribution of the volcanoes is anomalous. Instead of forming continuous chains parallel to the entire lengths of the axes of New Britain and the submarine trench, the volcanoes show a pronounced zig-zag distribution pattern (Fig. 4). The zig-zag consists of (1) the east-west zone of the Witu Islands (excluding Unea Island), (2) the north-south chain of volcanoes on Willaumez Peninsula, and (3) the zone of prominent volcanoes along the north central coast of New Britain, between Willaumez Peninsula and Likuruanga (Fig. 4).

Secondly, the bathymetry of the northwestern corner of the Solomon Sea shows that west of 150°E the New Britain trench shallows and splits into two branches (e.g. Krause et al., 1970; Mammerickx et al., 1971). A northern branch extends into Vitiaz Strait, and a southern one trends towards Huon Gulf. An unpublished BMR seismic reflection profile run south from near Kandrian (Fig. 1), shows that the southern branch is deeper than the northern one, and contains a thicker sedimentary sequence.

These two anomalous features show some noteworthy correlations (Fig. 2). Firstly, the Witu Islands zone lies north of the double-trench feature west of 150°E. Secondly, the trend of Willaumez Peninsula passes southwards through the point where the New Britain trench splits into two. Lastly, the north central coast volcanoes lie north of the deeper, single-axis part of the New Britain trench.

About 700 volcanic rocks from the eastern arc have been examined in thin section by the writer. Including the chemical analyses published by Lowder and Carmichael (1970), Johnson et al. (1972), and Blake and Ewart (1974), a total of 205 eastern arc rocks have been analysed for major elements. Like the western arc, low-silica andesite is common in the eastern arc. However, unlike the western arc rocks basalts are much less common, high-silica andesites and dacites are more common, and rhyolites are present, although rare. Except for these differences in the relative abundance of rock-types, the rocks of both the eastern and western arcs are chemically similar; all of them are hypersthene-normative.

Whereas the western arc shows systematic changes in rock compositions along the volcanic chain, the eastern arc volcanoes show changes across the arc - that is, in a direction at right-angles to the axis of the submarine trench, and concomitant with changing depths to the underlying Benioff zone. In Figure 5, it can be seen that in rocks with the same silica content, total-alkali contents increase progressively northwards

Fig 5

between zone E and zone H through the intermediate zones F and G (note also that rocks from the northern part of Willaumez Peninsula are different in composition from those in the southern part). This progressive increase in alkali contents is the same as those identified in many island-arcs as distance from the submarine trench and depths to the Benioff zone increase (e.g. Rittmann, 1953; Kuno, 1959). Johnson (in prep.) shows that $K_2O$ and $Na_2O$ (independantly), and $TiO_2$ and $P_2O_5$, also generally increase northwards in the eastern arc.

Discussion

The New Britain island-arc is widely accepted as a region where the Solomon Sea and South Bismarck plates converge, the former being consumed along the New Britain trench (Isacks and Molnar, 1971; Johnson and Molnar, 1972; Curtis, 1973; Krause, 1973)*. However, if plate convergence is taking place in this region, and the Solomon Sea plate is disappearing beneath New Britain, the branching of the submarine trench west of 150°E indicates that subduction cannot be regarded as a simple case of one plate disappearing beneath another along a single line. In addition, the correlation proposed above between the submarine features south of New Britain island and volcano distribution north of it suggests that the "double trench" feature west of 150°E has exerted a pronounced effect on magma genesis beneath New Britain.

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* Wiebenga (1973) presented an alternative interpretation that the New Britain trench is a tensional feature.

## SPECULATIONS ON PLATE KINEMATICS

### Rates of slab descent

In this section, some speculative interpretations are presented on the development of the western and eastern arcs, based on the initial conclusion that each one is underlaid by a subducted lithospheric slab. It is envisaged that these slabs are colder than the mantle material through which they have descended, and that they are capable of extracting heat from the surrounding mantle (e.g., Oxburgh and Turcotte, 1970; Minear and Toksöz, 1970, a.b; Griggs, 1972; Le Pichon et al., 1973). Furthermore, it is accepted that, all other factors being more or less equal, rates of descent will determine the effectiveness of the slabs as "heat sinks" - that slowly descending slabs will heat up more readily than quickly descending ones with the same composition and thickness. Thus, rates of slab descent will govern the thermal regimes (the shapes of isotherms) of the slabs and of those parts of the upper mantle above the slabs where magmas are believed to form. In addition, if intermediate and deep focus earthquakes take place because of mechanical failure in the cold interiors of the downgoing slabs, deeper earthquakes will take place more readily in those parts of the slabs which have remained colder to greater depths - that is, in slabs which have been subducted more rapidly.

### Development of the western arc

The events which are thought to have lead to the formation of the western arc are summarised in Figure 6. The most important points to note are: (1) that a Miocene continent/island-arc collision destroyed the plate boundary which had existed south of the Adelbert-Huon island-arc; (2) that post-Miocene plate convergence and forshortening caused the uplift of the Highlands and the associated welt of island-arc rocks (the coastal ranges); and (3) caused the steepening of the subducted slab which continued to descend beneath the South Bismarck plate, resulting in the Quaternary volcanism of

Fig 6

the western arc. It is also important to note that no single line at the Earth's surface is identified as the present-day plate boundary in this region; rather, the contact between the Indo-Australian and South Bismarck plates is visualised as a broad zone of foreshortening characterised by earthquakes which show compressional focal mechanism solutions, but which beneath mainland Papua New Guinea are not concentrated in a single, well-defined inclined seismic zone.

The 3-dimensional form of the subducted slab beneath the entire western arc cannot be deduced directly because of the apparent absence of intermediate- and deep-focus earthquakes west of Karkar Island. However, the distribution pattern of the volcanoes may be taken as circumstantial evidence for establishing the general shape of the slab throughout the Late Cainozoic.

The three closely-spaced parallel lines in Figure 2 are hypothetical strike-lines for the upper surface of the subducted slab. They are drawn parallel to the general trend of the western arc, and are dimensionless, although they might correspond with depths between about 150 and 250 km. Two features are noteworthy. Firstly, the lines are straight between the Schouten Islands and Long Island, but east of Long they curve towards the east (Fig. 2). Secondly, the offset of the Schouten Islands line suggests a comparable displacement may exist in the underlying slab.

Using this plate configuration, and accepting the proposal of Krause (1973) that the instantaneous poles of rotation for the Indo-Australian and South Bismarck plates could be in the northwestern part of mainland Papua New Guinea, it can be shown that for any finite period of time the magnitudes and directions of the relative velocity vectors change progressively along the contact of the two plates at depth. For example, using the pole position in Figure 2, rates of plate convergence increase regularly between the

Schouten Islands and the Cape Gloucester area because of the increasing distance from the pole. The direction of plate convergence will not, however, be at right-angles to the strike lines given in Figure 2, but will contain transcurrent components of motion. West of Long Island this transcurrent component will be right-lateral, but east of Long Island, it will change to left-lateral, as distance from the pole of rotation increases, because of the change in orientation of the strike lines. Similar changes in the rate of plate convergence (and the sense of transcurrent motion) can be proposed using many other pole positions in the northwestern part of mainland Papua New Guinea.

This interpretation is consistent with the known distribution of earthquakes beneath the western-arc. Intermediate-focus earthquakes are common east of Karkar Island perhaps because rates of plate convergence are greater there, and cold lithosphere has been subducted to greater depths than in the west. In addition, because the concentration of earthquakes south of Long Island coincides with the change in orientation of the strike lines, it is possible these earthquakes are due to high stresses set up where the slab is strongly bent. Rates of plate convergence are lowest in the west, where subduction may be so slow that the downgoing slab is no longer sufficiently cold to produce intermediate focus earthquakes when it reaches these depths.

If rate of subduction is a primary influence of the thermal regimes of the slab and the overlying upper mantle (see above), the conclusion that rates of plate convergence change progressively along the western-arc has important consequences for theories of magma genesis. Because of the changes in convergence rates, the patterns of isotherms may be different in all sections drawn perpendicular to the western arc. Slab-derived water and melts may therefore rise from the downgoing slab at different depths in each section, and may lead to partial melting at different depths in the overlying mantle. Magmas beneath each part of the arc may therefore be generated under unique sets of conditions of pressure, temperature, contents of volatiles, composition and volume of the slab melts (which may mix with the mantle-derived magmas), etc. The possible interactions between these variables are, of course, so complex that it is impossible to elucidate them satisfactorily at the present time. Indeed, additional factors may contribute to even greater complexities - e.g., compositional heterogeneities in the slab and upper mantle, lateral changes in thickness of the slab, lateral differences in thermal properties of the slab and overlying mantle, changes in the positions of the instantaneous poles of rotation throughout the Late Cainozoic, etc. However, it is proposed that in broad terms the changes in rock compositions along the western arc are functions of differences in the thermal regimes beneath each part of the arc, and that these, in turn, are dependent upon differences in rates of plate convergence and subduction along the western arc.

Finally, it is noted that these interpretations are consistent with the apparent increase in the volume of subaerial volcanic rocks from west to east along the western-arc. In the east, the volumes of slab-derived water and melts may be greater than those produced from the slab in the west because, during any finite period of time, a greater supply of subducted, water-rich crustal rocks will be made available for dehydration or fusion. In the east, these larger volumes are likely to induce more extensive partial melting of mantle peridotite compared to that in the west, and possibly therefore lead to the eruption of a proportionally greater volume of magma.

Fig 7

Development of the eastern arc

A cross-section through the eastern arc east of 150°E shows all the familiar features of an island-arc (Fig. 7A). The volcanoes (zones E and F of Fig. 5) overlie the shallow part of the New Britain Benioff zone, and their rocks are low in alkalies (Fig. 5), corresponding to rocks called "tholeiitic" by others (e.g. Tilley, 1950; Kuno, 1959; Jakeš and Gill, 1970). These rocks may have originated in the manner proposed by Nicholls and Ringwood (1973) - that sufficient volumes of water were released from the downgoing slab by subsolidus breakdown of amphibole, and rose to parts of the overlying upper mantle where temperatures were greater than those of the appropriate peridotite-water solidus (see Green, 1972). For example, in Figure 7A, water may be released from the slab at X and rise, and if sufficient water reaches Y at 50 kilometres depth it may cause melting there, because Y is above the temperature of the water-saturated peridotite-water solidus at 15 kilobars (Green 1972). Subsequent fractionation of these primary magmas may then lead to the spectrum of compositions represented by the rocks of zones E and F.

In contrast, the cross-section west of 150°E shows features which are markedly different from those in the eastern section (Fig. 7B): two trenches appear to be present; volcanoes equivalent to those of zones E and F are absent; and the Witu Islands represent the only known eastern arc volcanoes which overlie the deepest part of the New Britain Benioff zone. One possible interpretation of these features follows, but it is largely speculative, and should be regarded as tentative.

It is proposed that the branching of the submarine trench indicates that underthrusting may have taken place along two boundaries throughout the Late Cainozoic, and that the area of sea-floor between the two branches is the upper end of a thrust slice that dips northwards beneath ⟶

New Britain. This "Vitiaz" slice (Fig. 2) is envisaged as a sliver of crust and upper mantle - perhaps only 20 km thick - underlaid by the crust of the subducted Solomon Sea plate (Fig. 7B). Down-dip, the Vitiaz slice is visualized as thinning out, extending no further than about 70 or 80 km north of the north coast of New Britain, and no deeper than about 100 km. ~~(Fig. 8B).~~ The slice is presumed to have underthrust an unknown length of sea ~~ocean~~ floor (Fig. 7B).

In broad terms, it may be suggested that at depths less than 100 km the effect of the Vitiaz slice has been to complement the effect of the down-going Solomon Sea plate as a heat-sink; the combined effect of both the slice and the plate may have been to absorb more heat from the surrounding mantle than would have been absorbed by the Solomon Sea plate alone. In addition, if it is true that the present-day pole of rotation for the South Bismarck and Solomon Sea plates is west of New Britain (Krause, 1973), it follows that rates of subduction west of Willaumez Peninsula are lower than those east of it, and therefore at depths greater than about 100 km more heat may have been extracted by the slab in the east than in the west.

The speculative isotherms shown in Figure 7 reflect these proposed differences in the capacities for heat absorption by the slabs on either side of Willaumez Peninsula. The isotherms in the eastern section extend deeper within the slab than those in the west, and parts of the mantle above the slab in the west at depths less than about 100 km are cooler than equivalent parts in the east. Thus, in Figure 7B, water derived from X' (equivalent to X in Fig. 7A) will, on rising to Y' (equivalent to Y) at a depth of 50 km (15 kb) reach mantle whose temperature is lower than that of the water-saturated peridotite-water solidus (Green, 1972). No amount of water will cause melting, and therefore no volcanism will be expected along this part of the north coast of New Britain (Fig. 7B). To the north, however, water or water-rich melts are presumably derived from the slab at a few hundred kilometres depth. Experiments are, at present, unable to predict the composition of these melts and their effect on the overlying mantle, but mantle melting perhaps take place (to give rise to the volcanism of the Witu Islands) because the temperatures of the slab melts generated from the Solomon Sea plate west of Willaumez Peninsula are higher than those of the melts which might have been generated from the plate east of the Peninsula (compare the position of the 1350°C isotherm in Fig. 7).

The Vitiaz slice may also provide a possible explanation for the existence of Willaumez Peninsula whose southward extension intersects the point where the New Britain trench divides (Fig. 2). If underthrusting has taken place along both branches of the trench west of 150°E, rates of underthrusting at the Vitiaz/South Bismarck boundary will have been much less than rates at the Solomon Sea/South Bismarck plate boundary, because the combined rates of convergence for the two boundaries west of 150°E are thought to be less than the rate east of 150°E (see above). Thus, the region beneath Willaumez Peninsula, and beneath the South Bismarck plate, can be visualised as one in which the downgoing Solomon Sea plate east of 150°E may be dragged past the underthrust Vitiaz slice. If so, releases in total-pressure caused by shearing could have taken place in the "drag-zone" (Fig. 2) beneath Willaumez Peninsula, producing melting of subducted oceanic crust in a comparatively narrow, north-south zone. Perhaps, too, shear-strain heating (cf. Minear and Toksoz, 1970 a, b) was enhanced in this zone. Rise of the slab melts into the overlying upper mantle might have caused partial melting of peridotite, and the formation of primary magmas, which rose, fractionated, and produced the volcanoes of Willaumez Peninsula. In this interpretation, therefore, Willaumez Peninsula does not represent a fault in the South Bismarck plate, but rather is the surface expression of differential movements <u>beneath</u> the South Bismarck plate.

## CONCLUDING REMARKS

The volcanoes of the south Bismarck Sea are considered to be related to convergent plate boundaries which, throughout the Late Cainozoic, have been characterised by unusually complex relative movements. This paper has attempted to provide an explanation for the distribution of the volcanoes using orthodox principles of plate theory, but the interpretations should be regarded as tentative, as considerably more geophysical and geological data should be collected before any plate model for this region can be regarded as secure.

## ACKNOWLEDGEMENTS

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Fig. 1. Principal Late Cainozoic volcanoes of Papua New Guinea (solid dots), including those of the "western" and "eastern" arcs shown in Figs 3 and 4, respectively. Squares are settlements (the site of Rabaul township is a volcano).

Fig. 2. Tectonic features at the southern margin of the Bismarck Sea. Open circles represent epicentres of earthquakes that define the northern margin of the South Bismarck plate (after Denham, 1973). This northern margin is thought to continue southeastwards and reach the Solomon Sea plate at a triple junction south of New Ireland (see: Johnson and Molnar, 1972; Curtis, 1973; Krause, 1973). The solid lines marking the margins of the Solomon Sea plate and Vitiaz slice are the axes of submarine trenches south of New Britain (cf. Fig. 1). The western ends of these trenches are visualised as terminating at a transcurrent plate boundary (curved dashed line) the northern end of which is thought to mark the eastern end of the near-vertical slab beneath the western arc, and which may have been responsible for the apparent off-set between the coastal ranges and New Britain in Tertiary times. This postulated plate boundary is drawn as the northward extension of the zone of minor seismicity found in eastern Papua (e.g. Milsom, 1970), but is located with some reservation as there is no direct evidence for its existence on the floor of the northwestern Solomon Sea (in fact, the boundary may be a <u>zone</u> of deformation, and its position could have changed throughout the Late Cainozoic). The three closely spaced lines north of the coastal ranges are hypothetical, dimensionless strike lines for the upper surface of the postulated, northward-dipping slab beneath the western arc. The set of arrows illustrates the effect of plate convergence about a single position for a series of instantaneous poles of rotation (other positions in the northwest part of mainland Papua New Guinea have similar effects) - i.e., that those points at the upper surface of the downgoing slab further from the pole move greater distances for any period of rotation about the pole; in other words, rates of plate convergence increase away from the pole. The straight lines accompanying each arrow are normal to the strike lines. Note that because of the differences in orientation of the strike lines on either side of Long Island, the transcurrent components of movement for this particular pole position change from right-lateral west of the island to left-lateral east of it. A-A' and B-B' show locations of cross-sections in Figs 7A and B, respectively. See text for explanation of "drag zone".

# 13

Fig~~ure~~ 3. Principal volcanoes of the western arc. Islands are shown stippled or solid, and volcanic islands are indicated by bold lettering. Filled circles are prominent volcanic centres on mainland Papua New Guinea and New Britain, and on Long and Umboi Islands. Stars indicate volcanoes which have erupted during the last one hundred years. Langila is a cluster of satellite craters on the eastern flank of Talawe. The outlines of Schrader, and of the coastal ranges of mainland Papua New Guinea, are shown by an approximate 330 m contour.

Fig~~ure~~ 4. Principal volcanoes of the eastern arc. Filled circles are volcanoes on the New Britain mainland. Island volcanoes are shown solid. Dakataua, Witori, and Hargy are volcanoes containing calderas within which the volcanoes Makalia, Pago, and Galloseulo, respectively, have been built. 330m-contour shows the outline of Schrader and Andewa volcanic complexes. Stars indicate volcanoes which have erupted during the last one hundred years (Makalia and Bamus have probably also erupted during this period).

Fig~~ure~~ 5. Weight percent $Na_2O + K_2O$ <u>versus</u> $SiO_2$ diagrams (upper) for rocks from western and eastern arcs, using the division of volcanoes into zones A-to-H shown in the lower diagram. The dotted line in the upper diagrams separates the "high-alumina basalt" (upper) and "tholeiitic" fields of Kuno (1959). The solid lines are drawn arbitrarily, and define compositional fields a-to-h which correspond to zones A-to-H, respectively. Note that Ulawun, a zone E volcano, has rocks of similar alkali content to those ~~that~~ from volcanoes in zones F and G.

Fig. 6. Simplified scheme for the development of the western arc. Sketches 1 and 2 are based largely on the geological history given by Robinson et al. (1974). The sketches are highly diagrammatic and for the sake of clarity they ignore the complex geological events that took place at the northern edge of, and within, the continental mass (heavy stipple) now represented by the Highlands - Wau region (cf. Fig. 1). These events include, for example, the emplacement of the Marum Basic Belt, extensive magmatism in Middle Miocene times, and widespread volcanism in the Quaternary (see, for example: Page and McDougall, 1970; BMR, 1972; Mackenzie and Chappell, 1972; Bain, 1973; Mackenzie, this volume). At least some of the events are possibly related to underthrusting at the northern edge of the continent, as shown in sketches 2 and 3 by the dotted line (cf. Johnson and Molnar, 1972, their Fig. 6; present-day earthquake activity gives little indication of the existence of underthrust lithosphere beneath the Highlands--Wau region, except for a few intermediate depth earthquakes in the Wau area).

1. During the Upper Eocene and Oligocene the leading edge of the northward-moving Australian continent, carried on the Indo-Australian plate, moved towards a zone of northward subduction where oceanic crust was consumed beneath an Adelbert-Huon island-arc (light stipple). The "Finisterre Volcanics" (Robinson et al., 1974) are believed to represent island-arc magmas related to the downgoing portion of the slab.

2. By Lower Miocene times it is envisaged that the leading edge of the continent had collided with the island-arc, and that subduction and island-arc volcanism had effectively ceased. Plate convergence continued, however, and movements were taken up on faults in the collision zone (represented by the present-day Ramu and Markham Valleys), on block-faults in the island-arc, and on faults at the northern edge of the continental mass. Limestones were deposited on fault blocks in the northern part of the island-arc in the Lower and Upper Miocene and, to a lesser extent, in the Upper Miocene.

3. Convergence continued throughout post-Miocene times, resulting in the uplift, warping, and foreshortening, of the continental mass and its new welt of island-arc rocks (the coastal ranges). The subducted slab beneath the old island-arc was steepened, and movement of the Indo-Australian plate beneath the South Bismarck plate continued, giving rise to Quaternary volcanism in the western arc (see text).

Figure 7. Cross-sections showing schematic configurations of downgoing slabs and speculative thermal regimes east (A) and west (B) of Willaumez Peninsula. Locations of cross-sections are shown in Figure 2. Light stippling shows hydrated crustal rocks in the upper parts of the slabs. Coarse stippling indicates thickness of crust in cross-section A which coincides with the crustal profile K-L given by Finlayson and Cull (1973, their Fig. 9b; crustal thicknesses beneath west New Britain are undetermined). WP and E-F represent the northern end of Willaumez Peninsula, and zones E and F (Fig. 5), respectively, projected from the east onto cross-section B. X-Y and X'-Y' are ascent paths of slab-derived water (see text for details). The flat parts of the isotherms are taken from the "undisturbed" mantle geotherm of Griggs (1972).

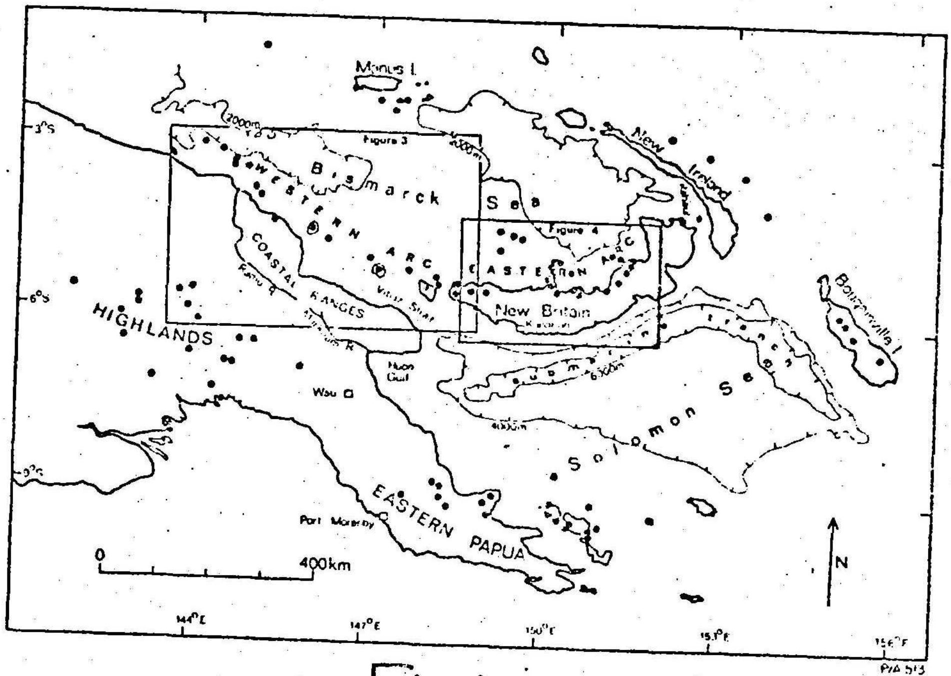

Manus I.
New Ireland
Bismarck Sea
WESTERN ARC
EASTERN ARC
Figure 3
Figure 4
COASTAL RANGES
Vitiaz Strait
New Britain
HIGHLANDS
Huon Gulf
Wau
Bougainville I.
Solomon Sea
EASTERN PAPUA
Port Moresby
2000m
4000m
6000m
3°S
6°S
9°S
144°E
147°E
150°E
153°E
156°E
0
400km
N
P/A 513

Fig 1

Fig 2

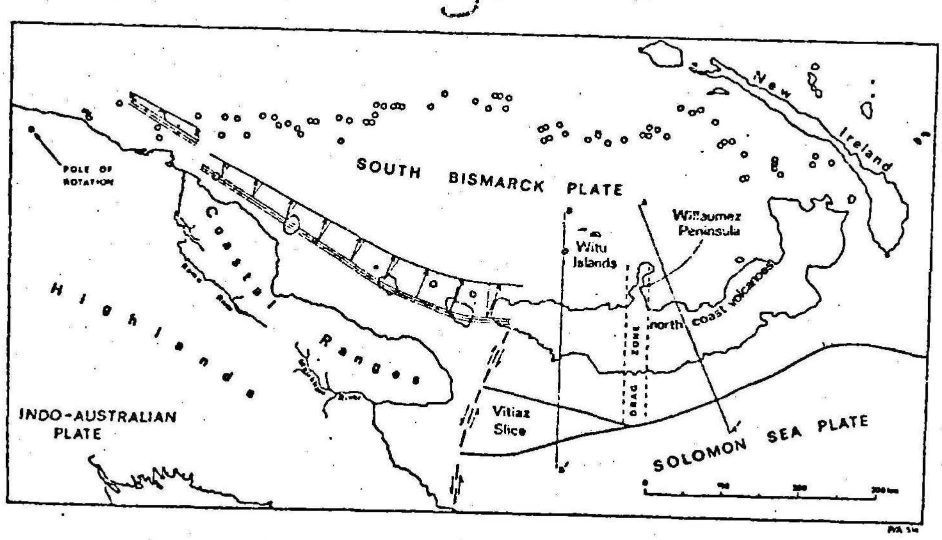
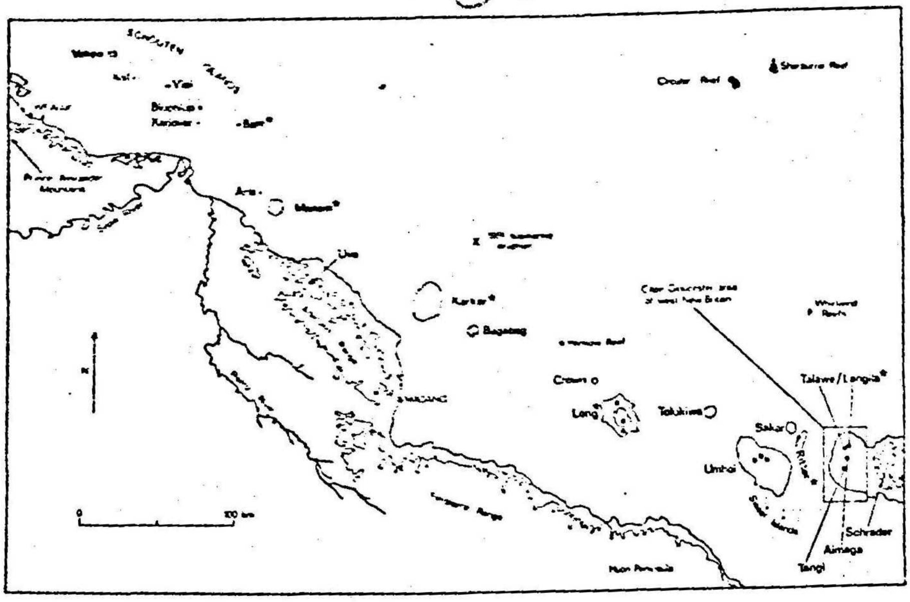
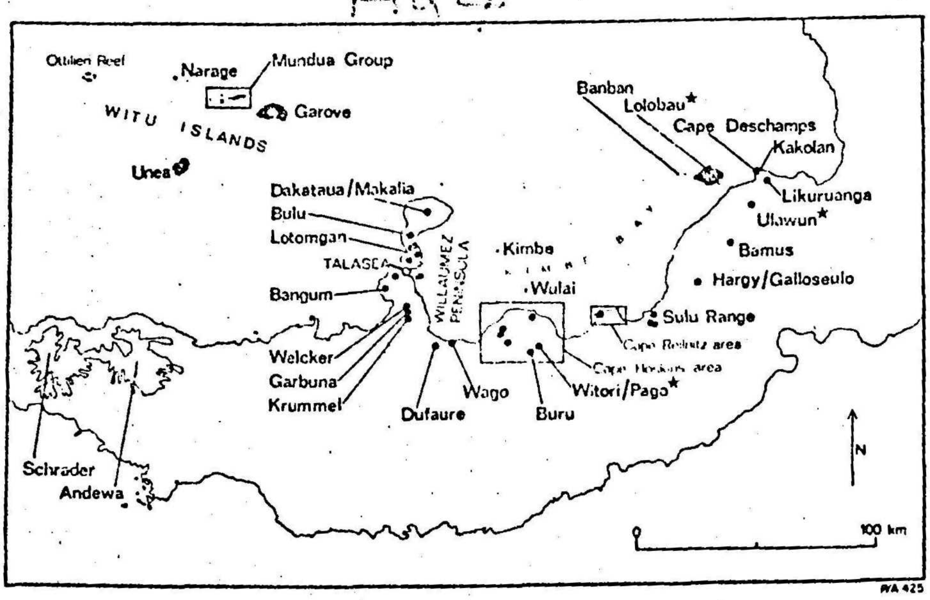

POLE OF ROTATION
SOUTH BISMARCK PLATE
New Ireland
Coastal Ranges
Highlands
Witu Islands
Willaumez Peninsula
north coast volcanoes
DRAG ZONE
Vitiaz Slice
INDO-AUSTRALIAN PLATE
SOLOMON SEA PLATE

Fig 3



Karkar*
Bagabag
Crown
Long
Tolokiwa
Sakar
Umboi
Talawe/Langila*
Schrader
Aimega
Tangi


Fig 4



Ottilien Reef
Narage
Mundua Group
Garove
WITU ISLANDS
Unea
Dakataua/Makalia
Bulu
Lotomgan
TALASEA
Bangum
Welcker
Garbuna
Krummel
WILLAUMEZ PENINSULA
Dufaure
Wago
Kimbe
Wulai
Buru
Witori/Pago*
Cape Hoskins area
Cape Reitutz area
Sulu Range
Hargy/Galloseulo
Bamus
Ulawun*
Likuruanga
Kakolan
Cape Deschamps
Lolobau*
Banban
Schrader
Andewa
N
0
100 km
WA 425

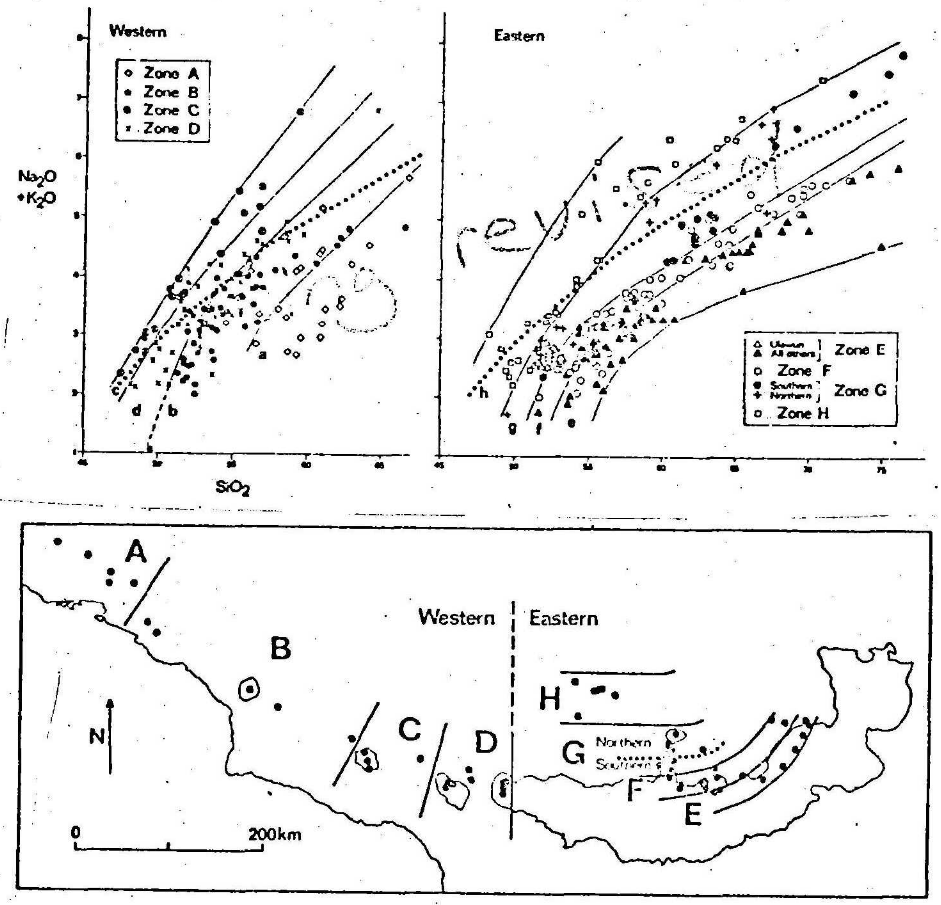

Western
Zone A
Zone B
Zone C
Zone D
$Na_2O$ $+K_2O$
$SiO_2$
Eastern
Zone E
Zone F
Zone G
Zone H
A
B
C
D
Western
Eastern
H
G
Northern
Southern
F
E
N
0
200km

Fig 5

Johnson, Taylor Volume paper

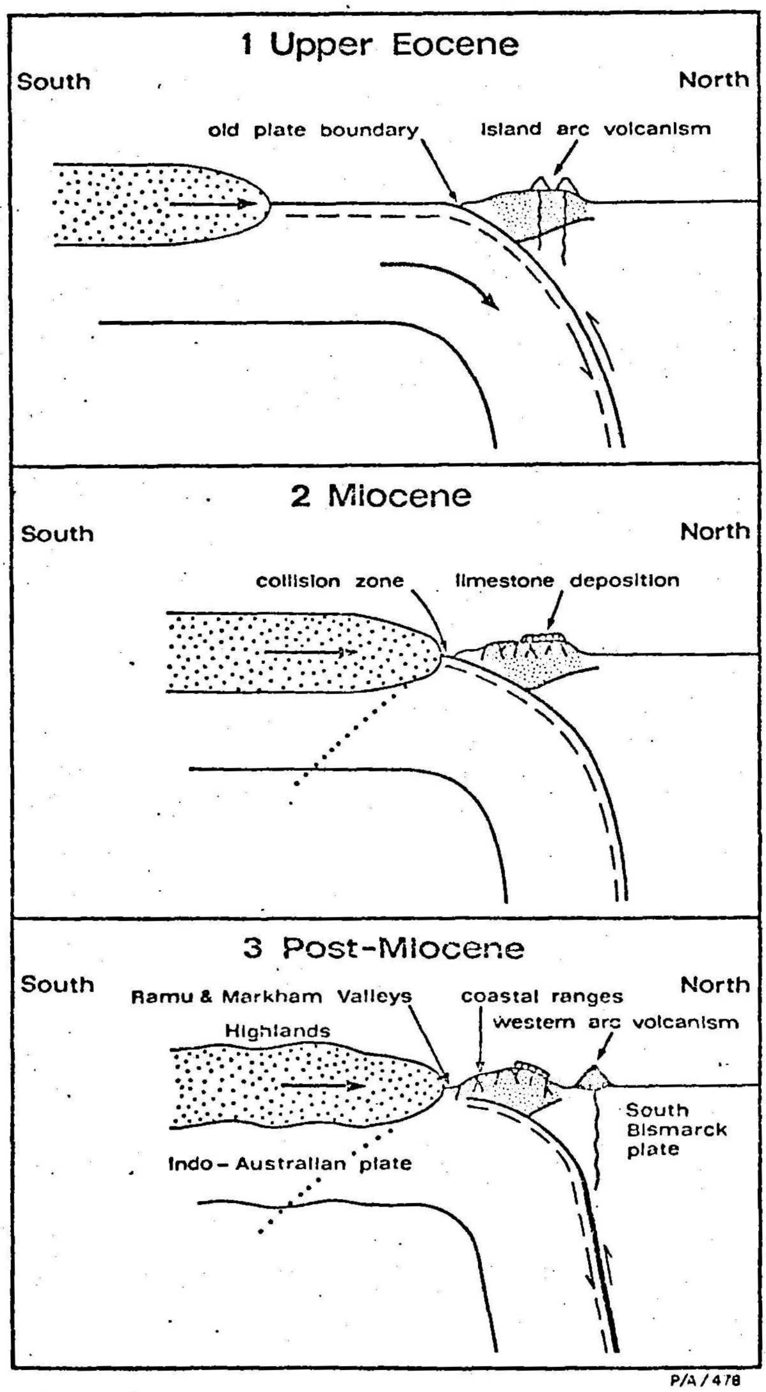
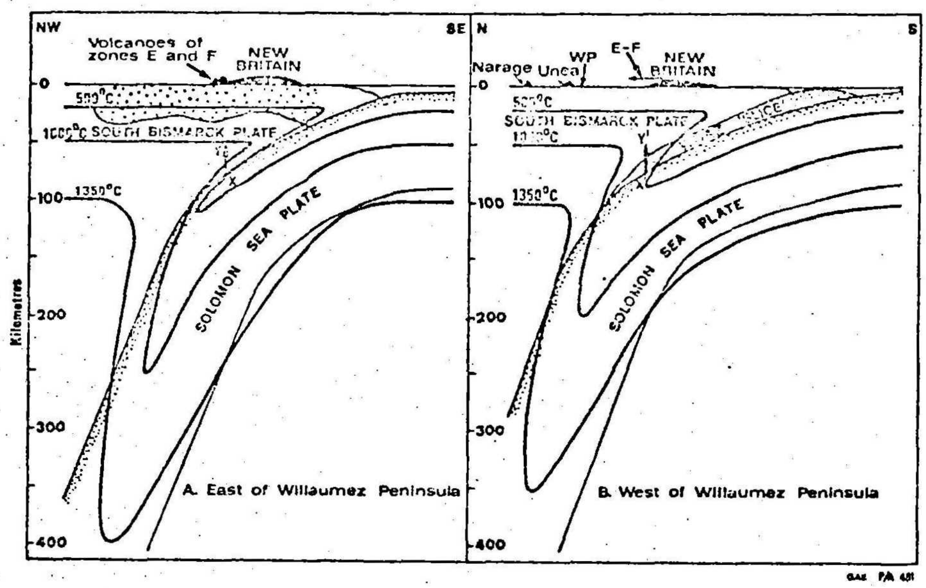



Fig. 6



NW
SE
Volcanoes of zones E and F
NEW BRITAIN
0
500°C
1000°C
SOUTH BISMARCK PLATE
100
1350°C
SOLOMON SEA PLATE
Kilometres
200
300
400
A. East of Willaumez Peninsula
N
S
Narage
Unea
WP
E-F
NEW BRITAIN
0
SOUTH BISMARCK PLATE
100
1350°C
SOLOMON SEA PLATE
200
300
400
B. West of Willaumez Peninsula


Fig 7

# NATURE AND ORIGIN OF LATE CAINOZOIC VOLCANOES, CENTRAL PAPUA NEW GUINEA

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ABSTRACT

Sixteen late Cainozoic volcanic centres in central Papua New Guinea rest on 25-30 km of Palaeozoic sialic crust and up to 10 km of post-Palaeozoic sedimentary and volcanic rocks. One rests on a 35 km-thick pile of eugeosynclinal sediments and volcanic rocks. Fourteen of the centres are deeply eroded stratovolcano, and the last a ring-like cluster of lava domes. Basaltic rocks are dominant in eleven centres, and these range from high-K, oversaturated (shoshonitic) types, through lower-K saturated types to low-K types just saturated to slightly undersaturated in silica. Other rock types in these centres, and the dominant rock types in the other six centres are low-Si andesites and andesites with high to moderate $K_2O$ contents. The magmas originated in the upper mantle low-velocity layer following plate collision in the middle to late Miocene and subsequent crustal warping and uplift continuing into the Pliocene. Complex chemical variations were produced by mantle inhomogeneity, various types and amounts of partial melting in rising diapirs at various levels, and various rates of ascent and degrees and types of crystal fractionation at the base of and within a mechanically inhomogeneous crust.

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1 15

## INTRODUCTION

The rocks discussed in this paper are the eruptive products of a widespread group of large Quaternary stratovolcanoes in central mainland Papua New Guinea (Fig. 1). These volcanoes first came to the attention of geologists in 1939 when L.C. Noakes, chief geologist of the Papuan Administration, visited Mounts Hagen, Giluwe, and Ialibu. Subsequently, Rickwood (1955) and the Australasian Petroleum Company (1961) mapped all of the volcanic centres in the area, except Mount Yelia which was recognized as volcanic in 1962 (Branch, 1967). In 1968, Bain *et* al. (1975) sampled Mounts Hagen, Giluwe, Ialibu, Suaru, and Karimui, and Crater Mountain, and Taylor (1971) visited Doma Peaks to investigate reported eruptive activity.

Following upon the work of several geologists (e.g. Dickinson & Hatherton, 1967), Jakeš & White (1969) related shoshonitic rocks from Mounts Hagen, Giluwe and Ialibu to a deep (ca. 200 km) portion of a supposed Benioff zone. Johnson *et al*.(1971) and Mackenzie & Chappell (1972), discussing chemical analyses of samples collected by Bain *et al*. (1975), showed that the shoshonites were genetically related to calcalkaline andesites and that there was no direct evidence, in particular intermediate and deep-focus seismic activity, to relate the volcanoes to a Benioff zone. Mackenzie & Chappell (1972) also pointed out that the volcanoes they described all rest on strongly buckled, but now stable, continental crust 30 to 40 km thick. They postulated partial fusion of eclogite detached from the base of downbuckled parts of the crust to explain the origin of the shoshonites.

New information on the volcanoes studied by Mackenzie & Chappell (1972), and on several other centres, reveals that basaltic rocks are dominant and range from silica-saturated or oversaturated, high-K types to low-K types just saturated or slightly under-saturated in silica. In all but three of the centres, andesite overlies and grades into basalt.

Such a combination of features is difficult to explain in terms of conventional models of magma generation in island-arc or continental margin areas involving downgoing lithospheric slabs, particularly as seismic evidence is lacking.

GEOLOGY

Geological and Tectonic Setting

All the volcanoes except Mount Yelia rest on up to 10 km of folded Mesozoic and Tertiary sedimentary rocks which, in turn, overlie 20-30 km (St John, 1970) of Palaeozoic continental crust. Part of this is exposed in the Kubor Range, the part expression of a 120 km-long anticline in which Permian granitic and older metamorphic rocks crop out at up to almost 4000 m above sea level. (Fig. 1). Mount Yelia rests on the New Guinea Mobile Belt (cf. Dow *et al.*, 1974), a thick, strongly folded and faulted sequence of Mesozoic and Tertiary rocks which includes a discontinuous 1000 km-long belt of Miocene volcanic rocks. The Miocene volcanics close to the northeastern corner of the Palaeozoic continental plate include high-potassium undersaturated rocks which are not found elsewhere in the belt. Beyond the mobile belt to the north and northeast are the old island arc-like features which make up the coastal ranges: the Torricelli 'block' and the Finisterre 'block'.

Gravity and seismic studies (e.g. St John, 1970; Jenkins, 1974) show that most of western Papua New Guinea has a crustal thickness of 30-35 km, and the presence of deep fault zones beneath many of the Quaternary volcanoes.

Seismic data (Denham, 1969; Johnson et al., 1971; Johnson & Molnar, 1972; Curtis, 1973; Dent, in prep.) show that present-day seismicity is confined to crustal and shallow (50 km) subcrustal depths beneath all of the volcanoes except Mount Yelia which is underlain by a small, pronounced pocket of intermediate-depth (175-190 km) seismicity. It is significant that some of the most frequent seismic activity in the area has occurred beneath Mount Yelia, and until the recent past, in the Doma Peaks area (Dent, in prep.); these are the only two volcanoes still showing signs of thermal activity. Branch (1967) recorded cold solfataras and gas seeps from Mount Yelia, and noted that local people talked of activity on the mountain. Taylor (1971) found warm and cold springs and solfataras on Doma Peaks and also commented on local legends of eruption such as those recorded by Glasse (1963).

Geological and geophysical evidence (e.g. as summarized by Jenkins, 1974) indicate that the crust beneath the volcanoes was strongly warped (up to 11 km of relief) during the late Miocene-early Pliocene, and uplifted (by up to 3 km) during the late Pliocene.

## Morphology

The volcanoes, except for Mount Giluwe which is shield-like in form, and Aird Hills which is a ring-like dome cluster, are all stratovolcanoes surrounded by extensive aprons. All except Mount Yelia, which shows only slight effects of erosion, are deeply eroded; basement is exposed in the erosion calderas of Mounts Hagen, Karimui, and Murray. Erosion is commonly asymmetric due to concentration of orographic rainfall on the sides facing the wet seasonal winds (Ollier & Mackenzie, 1974).

## Age

Volcanism was initiated in the early Pleistocene or perhaps late Pliocene, and has continued virtually until the present day in at least two centres. The oldest dates are 1.1 $\pm$ 0.4 m.y. for a plug near Mount Ialibu (Jenkins, 1974) and 0.85 $\pm$ 0.05 m.y. for a flow near Mount Kerewa (Williams et al., 1972). Three dates of around 0.21 m.y. for lavas from Mounts Hagen and Karimui were reported by Page & Johnson (1974).

INTRODUCTORY PETROLOGY

The system of nomenclature used here is the same as that of Mackenzie & Chappell (1972), except for a change in slope of the lines separating higher-K from lower-K rocks on the $K_2O/SiO_2$ diagram: the line separating shoshonitic from high-K calcalkaline rocks passes through 2 percent $K_2O$ at 50 percent $SiO_2$ and 3 percent $K_2O$ at 70 percent $SiO_2$. The high-K calcalkaline - calcalkaline dividing line has the same slope. This change brings the system more into conformity with nomenclature used by other geologists, such as Joplin (1968). The term 'transitional' applies to basalts with low $K_2O$ contents, saturated or just saturated in silica, which are neither shoshonitic nor alkaline. Those rocks which contain modal or normative nepheline and have low $K_2O$ contents are termed 'slightly alkaline'.

Most of the volcanoes are made up of olivine-bearing basalt, overlain and in places intercalated with, or grading into, basaltic (low-Si) andesite and amphibole-bearing andesite. The basalt occurs mainly as lava flows, with some interbedded ash, agglomerate, and lahar deposits; a larger proportion of the andesites is fragmental. The proportion of andesite to basalt rocks appears to be different from centre to centre. Andesitic rocks make up about 20 percent of Mount Hagen, about 5 percent of Mt Giluwe, 10 percent of Suaru, 20-30 percent of Crater Mt., less than 1 percent of Karimui, 5 percent of Mt Murray, 30-40 percent of Duau and Favenc, and about 30 percent of Mt Bosavi. However, because of their stratigraphic position and the more fragmental nature of the andesitic rocks, they may have been far more extensively eroded away than the basaltic rocks.

Doma Peaks, and Mounts Ne, Kerewa and Ialibu are made up largely of low-Si andesite and andesite. Olivine basalt occurs in Ialibu and Doma, but only in minor quantities. Mount Yelia is made up largely of hornblende-pyroxene andesite resting on minor olivine-bearing low-Si andesite. Aird Hills is made up of generally deeply weathered leucocratic andesite and dacite which are characterized by plagioclase phenocrysts up to 2 cm long.

PETROGRAPHY

The rocks from the central Papua New Guinea volcanoes may be divided into three general petrographic groups: basalts, basaltic (low-Si) andesites, and andesites and dacites.

The basalts of each volcanic centre or group of centres are petrographically distinct. Those of Mounts Hagen, Giluwe, and Karimui are similar to one another in most respects, and in many respects are also similar to the shoshonites of Wyoming (Nicholls & Carmichael, 1969). They differ in being fresher, generally holocrystalline and coarser-grained, and in commonly containing orthopyroxene instead of biotite. Most basalts of Mount Suaru and Crater Mountain contain less interstitial alkali feldspar and more closely resemble calcalkaline basalts. Two basalt specimens from Suaru have more alkaline affinities: one contains modal nepheline, and the other normative (but not modal) nepheline. The scarce basalts of Doma Peaks and Mount Ialibu are poor in olivine and rich in plagioclase phenocrysts. In the Doma basalt, plagioclase phenocrysts are up to 1.5 cm across, and olivine is largely confined to the groundmass. In basalts of Mounts Murray, Duau, and Favenc, plagioclase and olivine phenocrysts are generally less prominent, and orthopyroxene is rarer than in basalts of the more northerly centres. In Mount Bosavi, plagioclase phenocrysts are commonly absent, and nepheline appears in the groundmass of some basalts: they are distinctive rocks, both petrographically and chemically.

Sparsely scattered phenocrysts of largely to completely oxidized (to magnetite + clinopyroxene $\pm$ plagioclase) basaltic or kaersutitic hornblende occur in some basalts of all centres.

close up

~~Basaltic (low-Si) andesites~~

Compared to the basalts, the~~se rocks~~ basaltic andesites contain more orthopyroxene, especially as phenocrysts, and less, or no olivine. Basaltic hornblende is abundant in some rocks, but absent in others, notably those of Doma, Ne and Kerewa and most low-Si andesites of Mounts Giluwe and Ialibu.

close up

~~Andesites and dacites~~

The andesites are typically hornblende-two pyroxene andesites, some of which contain tridymite, cristobalite, or $\beta$-quartz, and some of which contain a trace of biotite. Dacites, containing more than about 10 percent modal free silica, are rare.

ital.

Plagioclase is the dominant phenocryst phase in almost all these rock types. Phenocrysts are large, complexly zoned and twinned, commonly show resorption features, and are commonly clumped together. These features are more pronounced in the larger phenocrysts of the low-Si andesites and andesites. Compositions of the phenocrysts (determined by electron microprobe) range from $An_{45}$ to $An_{84}$ in the core, and from $An_{60}$ to $An_{36}$ at the rim, with most in the range $An_{50}$ to $An_{65}$. Orthoclase content ranges from 0.5 to 4.5 percent.

ital.

Clinopyroxene forms almost ubiquitous euhedral phenocrysts (1-5 mm) and equant groundmass grains. Compositions (determined by electron microprobe) are in the ranges $Ca_{42-54}$, $Mg_{42-54}$, $Fe_{2.2-11}$ (cores) and $Ca_{34-51}$, $Mg_{40-51}$, $Fe_{4.5-18}$ (rims) [illegible], in the diopsidic augite, diopside, and salite fields. Within [illegible], there is an irregular trend towards higher Fe and lower Ca with [illegible] crystallization. Calcium and titanium contents of clinopyroxene [illegible] in the basalts are lowest in rocks from Hagen and Giluwe, higher in [illegible], [illegible] and Favenc, and highest in Bosavi. They follow the variation [illegible] composition. Al in sixfold co-ordination is low and vari[illegible] no indication of any high-pressure influence.

ital. <u>Olivine</u> occurs generally as [illegible] mm) phenocrysts, and commonly also as groundmass grains. [illegible] are up to 7 mm or more across and are commonly irregular [illegible]. Probe analyses show that the smaller phenocrysts range in [illegible] from $Fo_{70}$ to $Fo_{82-85}$ in the core and from $Fo_{60}$ to $Fo_{75-80}$ [illegible]. [illegible] larger phenocrysts in basalts from Mount Hagen range from [illegible] the core to $Fo_{70-84}$ at the rim, and contain chromite incl[illegible] phenocrysts, with reaction rims of opaque oxides $\pm$ amph[illegible]oxene, occur in andesite from one locality. Calculat[illegible] partition between olivine and liquid based on methods of [illegible] (1970) and Cawthorn <u>et al</u>.(1973) show that within experim[illegible] of the high-Mg olivine compositions are compatible [illegible] chemistry. However, the olivine from the andesite (analy[illegible] is clearly shown to be residual and out of equilibrium [illegible].

Large, irregular, commonly [illegible] or crystal clumps of olivine with chromite or chrome-[illegible] occur in basalts from several other centres, [illegible] and Bosavi. Features such as deformation lamell[illegible] (sponge-like texture) were observed in some of [illegible] features, together with the presence of chro[illegible] low CaO contents (zoning from 0.07 - 0.20 in the core to 0.21 - 0.6 at the rim) are suggestive of crystallization under high pressure (non-surface) pressure conditions (Simkin & Smith, 1970; Stormer, 1973).

ital Orthopyroxene compositions (based on 32 analyses of rocks from Hagen, Giluwe, and Murray) are in the range $Wo_{2.4-5.5}\ En_{57-75}\ Fs_{22-39}$, zoning out in one extreme case to $Wo_{7.0}\ En_{66}\ Fs_{27}$. Orthopyroxenes in the low-Si andesites are generally more magnesian than in the andesites. Aluminium contents are low and show no high-pressure influence.

ital Opaque minerals in the basalts are dominated by abundant titano-magnetite (magnetite-ülvospinel solid solution) which contains between 10 and 21 weight percent $TiO_2$. Contents of MgO, $Al_2O_3$, and, most noticeably, $Cr_2O_3$ in the magnetites increase from Hagen and Giluwe to Murray, Duau and Favenc, and then to Bosavi. Other opaque phases are the chromite inclusions in olivine (mentioned earlier), rare pyrite, and even rarer minute blebs of pyrrhotite or chalcopyrite, or both.

ital Chromite in two Hagen basalts is high in $Cr_2O_3$ (46 to 53 percent), and contains 4 to 10 percent MgO and 4 to 11 percent $Al_2O_3$. Chromites in basalts from Duau, Murray, and Bosavi are lower in $Cr_2O_3$ (26 to 42 percent), with MgO 3 to 9.3 percent, $Al_2O_3$ 10 to 17 percent, and 1 to 5 percent $TiO_2$. One analysed (by electron microprobe) grain in a Bosavi basalt is a chromian magnetite-ulvöspinel ($TiO_2$ 20%, $Al_2O_3$ 2.35%, $Cr_2O_3$ 6%, MgO 2.0%).

ital Magnetite (sensu lato) is also dominant in the low-Si andesites and andesites, but there is an increasing amount of exsolution of ilmenite as the host rocks approach dacitic compositions. The magnetites contain between 4 and 14 percent $TiO_2$, and always contain less MgO and more $Al_2O_3$ than the co-existing ilmenite.

## CHEMISTRY

A total of 185 samples have been analysed for major and selected

Table 1

trace elements; representative analyses are presented in Table 1.

Major elements.

Overall, the major element data show a broad diffuse spectrum of compositions. However, on closer examination, each volcanic centre is seen to have its own chemical characteristics and trends. Some of these are illustrated in Figures 2 to 7.

Fig 2

The plot of MgO against total Fe as FeO (Fig. 2) shows the overall lack of iron enrichment, the low total Fe and high Mg contents of the Hagen and Giluwe rocks relative to the other rocks, and the clear separation of the basalts and andesites of Bosavi. These, and the following important features, are shown by the Mg-number - silica plot (Fig. 3)*:

Fig 3

1. The high Mg-numbers of most Hagen and Giluwe rocks, in particular a group of Hagen basaltic rocks, relative to the other rocks.

2. A concentration of analyses of Hagen and Giluwe rocks in the ranges $SiO_2$ 51 to 56 percent and Mg-number 50 to 60.

3. The distinct group of transitional (to alkaline) Bosavi basalts, and the group of Suaru and Karimui rocks with low Mg-numbers.

4. Two Hagen and two Giluwe high-K andesites with very low Mg-numbers: these are high-level intrusive and late-stage cumulodome rocks respectively.

The total alkalis-silica plot (Fig. 4) shows the distinct trends, or grouping of analyses, of rocks from most centres (Doma-Ne, Kerewa, and Favenc produce scattered plots), and, again, the distinct 'cluster'

Fig 4

* Mg-number is calculated as molar MgO divided by molar FeO normalized to 80 percent of total molar Fe as FeO . This eliminates the effect of late-stage oxidation and normalizes $FeO/Fe_2O_3$ ratio to a value proposed (Nicholls & Whitford, this volume) as the average value for the mantle.

of Bosavi basalts. It also shows the relationship between total rock and groundmass compositions for some of the Hagen and Murray rocks. In the case of Mount Hagen, there are two distinct populations, one basaltic (or shoshonitic) the other high-K andesitic, each with a distinct fractionation trend. In the case of Mount Murray, the total-rock/groundmass tie lines are roughly tangential to the total-rock trend, probably reflecting simple crystal fractionation.

The $K_2O$-$Na_2O$ and $K_2O$-$SiO_2$ diagrams (Figs 5 & 6) show the high K/Na ratios and $K_2O$ contents of the Hagen, Giluwe, Suaru, and Karimui rocks relative to those from the other centres. There is also a tendency in the more basic rocks for K/Na and $K_2O$ to decrease with increasing distance from the edge of the Palaeozoic continental plate: values are slightly lower in Murray, Duau and Favenc than in Hagen, Giluwe, Suaru, and Karimui, and lower still in the Bosavi rocks. The diagrams also show the separation of the basalts and the andesites from Hagen and Bosavi, and the diffuse overlapping fields of the more acid rocks of the other centres.

Fig 5
Fig 6
sert on opposite pages

The $TiO_2$ vs $SiO_2$ plot (Fig. 7) shows a broad, roughly linear trend of low $TiO_2$ values decreasing with increasing $SiO_2$. The rocks of Mounts Murray, Yelia, and Giluwe can be distinguished from the remainder on the basis of higher or lower $TiO_2$. The Bosavi basalts form a distinctive group at the basic end of the trend.

Fig 7

Other variations in major element chemistry (not illustrated) are:

1. High variable $Al_2O_3$ contents, reflecting mainly plagioclase phenocryst content. There is a strong negative correlation between $Al_2O_3$ and $SiO_2$ in Hagen and Giluwe rocks in the 51-55 and 50-56 percent silica ranges respectively. In the Hagen rocks, this cuts across a strong overall positive correlation.

2. CaO content shows a strong linear negative correlation with $SiO_2$, and a narrow range of values for any one value of $SiO_2$. There is a cross-trend in the same ranges of Hagen and Giluwe rocks that show the $Al_2O_3$ cross-trend.

3. $Na_2O$ contents increase from 2.5 - 3.5 percent in the basalts to about 4.5 percent in the andesites, in contrast with potassium which increases only slightly or, in some centres, decreases from basalt to andesite.

4. Moderately high $P_2O_5$ contents (0.3-0.7 percent) relative to calcalkaline rocks from other areas (cf. e.g. Taylor, 1969).

Trace elements

Content of the 'incompatible' trace elements such as Rb, Sr, and Ba is generally high and correlates with $K_2O$ content; this appears to be typical of 'shoshonitic' volcanic rocks described by numerous geologists (e.g. Nicholls & Carmichael, 1969). Rb is highest (55 to 90 ppm) in the high-K olivine basalts and low-Si andesites (shoshonites) of Hagen, and in the samples from Giluwe, Karimui, and Doma. The high-K low-Si andesites and high-K andesites from Mt Hagen are generally significantly lower in Rb than the shoshonitic rocks. Sr shows no systematic variation from centre to centre. Rocks from Suaru, Karimui and Aird are high in Sr (over 900 ppm) and those from Yelia low (490-580 ppm) relative to their respective silica contents when compared with rocks from the other centres. Barium is lower (250-400 ppm at 53% $SiO_2$) in Murray, Duau, and Favenc, and in the Bosavi basalts than in the other centres (420-820 ppm at 53% $SiO_2$); it is particularly high in rocks from Suaru and Doma (590-920 ppm).

Table 2

Pb, La, and Ce are also high relative to calcalkaline rocks from other regions (Table 2), but like Rb, Sr, and Ba, are 'normal' for potassic (shoshonitic) volcanic rocks. Y, Zr, Nb, Pb, La, and Ce are higher in rocks from Doma and Kerewa than in those from any of the other centres.

Contents of V, Ni, and Cr, particularly in some of the basaltic rocks of Hagen, Crater, Bosavi, Murray, and Duau, are very high relative to other shoshonitic and calcalkaline volcanic rocks (Table 2). This is at least in part attributable to the high ferromagnesian phenocryst content of these rocks, in particular chromite-bearing olivine.

K/Rb ranges from 225 to 660, with values for most rocks between 225 and 400, and the most common values 255–295 and 325–385. Similar ratios are reported in, for example, high-K andesites from Mt Bagana on Bougainville (Taylor et al., 1969a) and from Peru (Dupuy & Lefèvre, 1974), in the New Zealand andesites, and in the Aeolian trachybasalts and trachyandesites (Table 2). Values from the Peruvian and Fijian 'shoshonites' and the calcalkaline rocks of Japan are higher.

Strontium isotope ratios range from 0.7037 to 0.7059 (Page & Johnson, 1974; R.W. Page, written communication, 1975), within the range for 'modern depleted mantle' (Gill & Compston, 1973). Each volcanic centre has a characteristic narrow range of Sr isotope ratios, with lowest values in Hagen, Giluwe, Suaru, Karimui, and Crater, higher values in Murray and Doma, and highest values in Bosavi.

# DISCUSSION

~~Origin of the parent magmas.~~

The data presented above shows the central Papua New Guinea volcanoes to be a scattered group, each centre with its own petrological 'stamp', and together covering a broad range of compositions. There are, however, regional chemical differences, with shoshonitic and related rocks near the northeastern corner of the Australian Palaeozoic continental plate, and ~~to~~ just-saturated and slightly undersaturated basaltic and related rocks farther to the south and southwest.

Some rocks from the volcanoes have many features in common with island-arc and continental margin volcanic rocks, such as some of those in the Andes (e.g. Lefèvre, 1973), Fiji (Gill, 1970), and the Mediterranean area (e.g. Keller, 1974), which are generally believed to be related to processes in or above Benioff zones. However, the central Papua New Guinea volcanoes, with only one exception are underlain by relatively stable old continental crust, beneath which there is no intermediate to deep-level seismicity attributable to a currently active Benioff zone. The one exception is Mount Yelia, which, as pointed out previously, is unlike any of the other centres either petrologically or in its geological and tectonic setting. It is situated almost directly above a concentrated pocket of intermediate-depth (175-190km) seismicity which is probably connected with some form of underthrusting.

Basaltic rocks from centres nearest the Kubor Range area (Hagen, Giluwe, Suaru, and Karimui) are petrologically similar to the type shoshonites of Wyoming, while the less oversaturated and lower-K basalts from centres farther away, apart from their low $TiO_2$ contents, are more like some continental basalts and alkali basalts. There is now a substantial body of evidence (e.g. Smith & Sbar, 1974) to indicate that volcanism in the Yellowstone Park area of Wyoming is related to a rising mantle plume with accompanying 'hot-spot', uparching and radial tension-fracture system. Continental alkali basalts are generally related to crustal uplift and block-faulting or rifting movements.

Central and northern Papua New Guinea has a history of intermittent volcanism dating back to the Upper Triassic (Mackenzie, 1975). Much of the earlier volcanism involved calcalkaline magmas, and, in the light of abundant observational and experimental data, it is not unreasonable to assume that at least some of the magmas were generated in or above subduction zones. If this was so, then the increasingly intense and widespread magmatic activity until the Middle Miocene reflects a parallel increase in subductive activity as the Indo-Australian plate moved north. Consequently, the asthenosphere beneath the northern edge of this plate would have become increasingly modified northward by addition of water and incompatible elements (e.g. Ringwood, 1974) from subducted oceanic crust. Buckling and uplift of the crust in the area following collision with the Finisterre and Torricelli 'blocks' (e.g. see Johnson, this volume) and the extensive Middle Miocene igneous activity (Page & McDougall, 1970) immediately preceded initiation of volcanism in the late Pliocene-early Quaternary. Geological evidence (Jenkins, 1974; BMR, 1974; Bain et al., 1975) shows that deformation and uplift was most intense in the Kubor Range area (Fig. 1) and decreased to the west, south, and southeast.

The volumetric abundance of basic lavas, commonly with high Mg-numbers, the presence in some of these of possible high-pressure, Mg-rich chromite-bearing olivine, high Cr, Ni, and V contents (in particular — Taylor et al., 1969b), and low Sr isotope ratios are all indicative of an ultimate source of the magmas in the mantle. It is therefore speculatively suggested that the Miocene-Pliocene buckling and uplift triggered off partial melting and diapiric rise in the low velocity layer (cf. Green, 1972, 1973) and that the compositions of the melts derived from these diapirs reflect chemical changes imposed upon the low velocity layer by past episodes of subduction. Further variations in magma composition, such as high-K basalt to high-K low-Si andesite, was probably produced by variations in the amount and depth of partial melting in the diapirs. These variations could in turn have been at least partly controlled by regional variations in the intensity of Miocene-Pliocene lithospheric disturbances. Olivine, pyroxene, probably including, or perhaps exclusively orthopyroxene, and possibly also garnet may have been residual phases. These minerals could account for much of the major element variation, and the enrichment in volatile trace elements (cf. Green, 1972; Ringwood, 1974; Nicholls, 1974), but could not produce the enrichment in K over Na. The compositions of the magmas that finally reached the surface reflect the parent compositions, and the amount and type of subsequent fractionation. Fractionation may also be partly controlled by the degree of deformation of the crust, the less deformed and fractured areas affording easier and quicker access to the surface and hence less opportunity for fractionation.

The type and amount of fractionation varies widely from centre to centre, and it is not proposed to deal with this problem in detail: it will be the subject of part of a later, more comprehensive publication. In Mount Hagen, for example, basalts and andesites cannot be related by simple low-pressure fractionation. The main chemical variations (increased Si, Al, and Na, and decreased K, Mg, Fe, Ca, and Ti) from basalt to andesite can best be explained by fractionation at depth involving a relatively potassium-rich sodium-poor amphibole and olivine, which may be liquidus phases in the basaltic compositions at or near the base of the crust (cf. Nicholls, 1974). There is some evidence, in the form of cross-trends perpendicular to the overall trends of increasing Al and decreasing Ca with increasing Si, of limited plagioclase accumulation in the Hagen rocks.

Mount Giluwe is petrologically similar to Mount Hagen, but the volume of andesite is very small. However, if Giluwe and Ialibu are genetically related, as is considered possible, the Giluwe-Ialibu pair is a close parallel to the basalt/high-K low-Si andesite - andesite association of Hagen. If Ialibu is not related to Giluwe, it is then in the same category as Doma, Ne and Kerewa and also Aird, where the parent basaltic magma does not appear to have reached the surface, and only fractionated magmas were erupted. Crater Mountain, which seems to be out of place among the centres near the Kubor Anticline, and Mounts Murray, Duau, and Favenc all show chemical variation trends compatible with fractionation of a low-K saturated basaltic parent involving residual olivine and clinopyroxene. Mount Bosavi appears to have erupted two separate and distinct batches of magma, one a basic, low-K basalt on the borderline between under and oversaturated, and the other andesitic.

## CONCLUSION

The central Papua New Guinea volcanoes are an example of the value of an integrated study of regional geology and tectonics, as well as volcanic petrology. Their study has shown that although a group of volcanoes is situated in an area in which subduction has occurred and is still occurring, they are not automatically the direct or indirect products of subduction. Not only each group of volcanoes, but each volcano should be treated on its own merits.

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TABLE 1. Major and trace element analyses of representative specimens from some of the central Papua New Guinea volcanoes.

| | MT. HAGEN | | | | | | MT. GILUWE | | | | MT. MURRAY | | | MT. DUAU | | MT. FAVENC | | MT. BOSAVI | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 | 15 | 16 | 17 | 18 | 19 | 20 |
| $SiO_2$ | 50.5 | 52.2 | 54.4 | 56.8 | 58.4 | 60.8 | 51.2 | 54.9 | 56.1 | 56.8 | 49.9 | 52.3 | 58.5 | 50.3 | 59.7 | 52.2 | 62.7 | 47.1 | 52.5 | 60.2 |
| $TiO_2$ | 0.92 | 1.01 | 1.00 | 0.81 | 0.72 | 0.50 | 0.99 | 0.85 | 0.78 | 0.62 | 1.34 | 1.48 | 1.01 | 1.20 | 0.72 | 1.25 | 0.53 | 1.28 | 1.01 | 0.68 |
| $Al_2O_3$ | 13.5 | 15.1 | 15.3 | 17.2 | 18.0 | 18.2 | 17.3 | 16.7 | 15.8 | 18.8 | 14.7 | 16.8 | 17.1 | 15.3 | 17.4 | 17.4 | 17.6 | 14.1 | 16.2 | 17.3 |
| $Fe_2O_3$ | 1.55 | 1.45 | 2.65 | 2.65 | 2.50 | 2.25 | 2.60 | 2.15 | 2.20 | 4.80 | 1.85 | 2.45 | 3.55 | 2.85 | 3.25 | 3.10 | 2.00 | 3.45 | 2.95 | 2.95 |
| FeO | 6.65 | 6.65 | 5.05 | 3.95 | 3.65 | 2.65 | 5.25 | 4.85 | 4.15 | 1.35 | 7.35 | 6.55 | 2.90 | 6.05 | 2.45 | 5.95 | 2.45 | 6.55 | 5.40 | 2.70 |
| MnO | 0.14 | 0.15 | 0.14 | 0.12 | 0.12 | 0.11 | 0.14 | 0.12 | 0.12 | 0.13 | 0.17 | 0.16 | 0.16 | 0.17 | 0.14 | 0.18 | 0.12 | 0.18 | 0.18 | 0.12 |
| MgO | 12.4 | 8.20 | 6.60 | 5.15 | 3.45 | 2.10 | 6.50 | 5.55 | 6.45 | 2.85 | 8.55 | 5.20 | 2.25 | 8.75 | 2.60 | 4.95 | 1.90 | 11.7 | 6.75 | 2.10 |
| CaO | 7.85 | 8.80 | 7.65 | 5.60 | 5.75 | 5.45 | 8.75 | 7.50 | 6.35 | 6.40 | 9.60 | 7.95 | 6.40 | 9.30 | 6.30 | 8.60 | 5.15 | 9.90 | 8.20 | 6.00 |
| $Na_2O$ | 2.65 | 2.85 | 3.30 | 3.85 | 3.85 | 4.05 | 3.15 | 3.50 | 3.40 | 4.15 | 2.70 | 3.25 | 3.65 | 2.50 | 3.65 | 3.20 | 3.90 | 2.90 | 3.55 | 4.05 |
| $K_2O$ | 2.35 | 2.35 | 2.30 | 1.85 | 1.70 | 2.30 | 2.35 | 2.30 | 2.90 | 2.40 | 1.75 | 2.05 | 2.70 | 1.75 | 2.05 | 1.65 | 2.20 | 0.95 | 1.80 | 2.50 |
| $P_2O_5$ | 0.58 | 0.59 | 0.45 | 0.23 | 0.29 | 0.31 | 0.58 | 0.55 | 0.47 | 0.79 | 0.69 | 0.70 | 0.50 | 0.70 | 0.47 | 0.65 | 0.36 | 0.42 | 0.50 | 0.34 |
| $H_2O^+$ | 0.64 | 0.81 | 0.74 | 1.60 | 1.40 | 0.95 | 0.92 | 0.64 | 0.80 | 0.59 | 0.75 | 0.94 | 0.90 | 0.92 | 1.05 | 0.62 | 0.93 | 0.97 | 0.66 | 0.72 |
| $CO_2$ | 0.03 | 0.07 | - | 0.05 | 0.05 | - | 0.07 | - | 0.08 | 0.03 | - - | - | 0.03 | - | 0.10 | 0.05 | 0.01 | 0.02 | - | 0.02 |
| S | 0.03 | 0.03 | 0.02 | 0.02 | 0.02 | 0.02 | 0.02 | 0.01 | 0.02 | 0.03 | 0.04 | 0.03 | 0.04 | 0.04 | 0.06 | 0.02 | 0.03 | 0.04 | 0.02 | 0.04 |
| TOTAL | 99.79 | 100.26 | 99.60 | 99.38 | 99.93 | 99.69 | 100.00 | 99.62 | 99.62 | 99.75 | 100.02 | 99.86 | 99.68 | 99.89 | 99.94 | 99.82 | 99.83 | 99.57 | 99.72 | 99.72 |
| Rb | 74 | 69 | 65 | 41.5 | 48.5 | 64.0 | 65.5 | 82.5 | 108 | 82.0 | 38.0 | 43.0 | 81.0 | 65.0 | 59.5 | 39.2 | 67.9 | 29.0 | 39.5 | 71.0 |
| Sr | 560 | 615 | 980 | 705 | 755 | 935 | 770 | 815 | 710 | 945 | 635 | 765 | 785 | 680 | 925 | 705 | 820 | 675 | 690 | 765 |
| Y | 20.0 | 21.0 | 24.5 | 18.0 | 17.5 | 18.0 | 21.0 | 21.5 | 22.0 | 21.5 | 21.4 | 24.5 | 23.5 | 24 | 23 | 25 | 17 | 20.5 | 22.0 | 21.5 |
| Zr | 83 | 92 | 155 | 155 | 145 | 180 | 84 | 98 | 145 | 125 | 115 | 135 | 285 | - | - | - | - | 146 | 187 | 256 |
| Nb | 3.1 | 2.1 | 3.7 | 5.0 | 6.3 | 4.8 | 2.4 | 2.1 | 5.8 | 4.4 | 6.3 | 9.0 | 12 | - | - | - | - | 8.9 | 13.5 | 14.0 |
| Pb | 6.8 | 9.2 | 17.5 | 12.0 | 14.0 | 17.0 | 8.0 | 11.5 | 14.5 | 13.5 | 8.7 | 10 | 14 | 12 | 16 | 13.5 | 15.5 | 4.8 | 7.3 | 8.6 |
| Ba | 405 | 485 | 935 | 705 | 680 | 815 | 480 | 560 | 650 | 655 | 295 | 330 | 700 | - | - | - | - | 355 | 520 | 760 |
| La | 112 | 13.0 | 35.5 | 23.5 | 25.5 | 31.0 | 16.5 | 19.0 | 25.0 | 21.0 | 14 | 21.5 | 39.0 | - | - | - | - | 26 | 31.5 | 34.0 |
| Ce | 45 | 25 | 51.5 | 44.0 | 36.5 | 39.5 | 33.0 | 22.5 | 46.5 | 32.5 | 21 | 43 | 70.5 | - | - | - | - | 64 | 51.5 | 53.0 |
| Nd | - | - | - | - | - | - | - | - | - | - | - | 27 | - | - | - | - | - | - | - | - |
| Cu | 75 | 105 | 129 | 59 | 53 | 20 | 122 | 85 | 78 | 30 | 135 | 100 | 22 | 95 | 40 | 80 | 20 | 68 | 52 | 42 |
| Co | 78 | 63 | 64 | 32 | 42 | 19 | 44 | 35 | 37 | 20 | 66 | 50 | 21 | 47 | 23 | 36 | 11 | 72 | 46 | 40 |
| Ni | 340 | 115 | 75 | 75 | 25 | 11 | 81 | 70 | 130 | 10 | 95 | 33 | 8.0 | 180 | 23 | 43 | 19 | 222 | 97 | 10 |
| Sc | 25 | 28 | 24 | 18 | 17 | 9 | 25 | 22 | 19 | 10 | 31 | 24 | 12 | 29 | 13 | 22 | 8 | 30 | 22 | 12 |
| V | 210 | 230 | 205 | 150 | 125 | 69 | 260 | 185 | 145 | 69.3 | 285 | 280 | 130 | 280 | 160 | 230 | 56 | 280 | 195 | 120 |
| Cr | 608 | 365 | 244 | 222 | 68.0 | 26.5 | 150 | 163 | 274 | 0.9 | 355 | 75 | 0 | >300 | 60 | 92 | 25 | 780 | 335 | 7.1 |
| Th | - | - | - | - | - | - | - | - | - | - | - | - | - | 1 | 5 | 4 | 7 | - | - | - |
| U | - | - | - | - | - | - | - | - | - | - | - | - | - | 0.3 | 1.4 | 0.2 | 1.3 | - | - | - |
| CIPW NORMS. e.c. | | | | | | | | | | | | | | | | | | | | |
| Q | - | - | 2.00 | 6.97 | 11.2 | 13.4 | - | 2.43 | 3.17 | 7.87 | - | 1.27 | 12.2 | - | 15.0 | 2.49 | 17.6 | - | - | 12.5 |
| C | - | - | - | - | 0.04 | - | - | - | - | - | - | - | - | - | - | - | 0.22 | - | - | - |
| Z | - | - | 0.03 | 0.03 | | 0.04 | 0.02 | 0.02 | 0.03 | 0.02 | 0.02 | 0.03 | 0.06 | - | - | - | | 0.03 | 0.04 | 0.05 |
| or | 13.9 | 13.9 | 13.6 | 11.0 | 10.1 | 13.6 | 13.0 | 13.7 | 17.2 | 14.3 | 10.4 | 12.1 | 16.0 | 10.5 | 12.0 | 9.78 | 13.0 | 5.64 | 10.7 | 14.8 |
| ab | 22.4 | 24.1 | 27.9 | 32.6 | 32.6 | 34.3 | 26.7 | 29.6 | 28.8 | 35.1 | 22.7 | 27.5 | 30.9 | 21.0 | 30.9 | 27.1 | 33.0 | 21.6 | 30.0 | 34.3 |
| an | 18.0 | 21.5 | 20.2 | 24.3 | 26.8 | 24.8 | 26.6 | 23.1 | 19.3 | 25.6 | 22.9 | 25.3 | 22.4 | 25.4 | 25.1 | 28.3 | 23.4 | 22.7 | 23.0 | 21.7 |
| ne | - | - | - | - | - | - | - | - | - | - | - | - | - | - | - | - | - | 1.60 | - | - |
| di | 13.8 | 14.7 | 12.4 | 1.73 | - | 0.65 | 10.4 | 8.85 | 7.24 | 0.99 | 16.6 | 8.12 | 4.98 | 13.2 | 2.18 | 8.20 | - | 19.2 | 11.9 | 4.97 |
| hy | 4.60 | 11.6 | 16.0 | 15.8 | 12.2 | 7.27 | 9.2 | 15.4 | 17.3 | 6.64 | 8.13 | 16.7 | 4.17 | 16.6 | 6.17 | 15.0 | 6.87 | - | 13.7 | 5.36 |
| ol | 20.9 | 8.07 | - | - | - | - | 5.65 | - | - | - | 11.5 | - | - | 4.18 | - | - | - | 19.3 | 2.43 | - |
| mt | 2.33 | 2.22 | 3.95 | 3.95 | 3.68 | 3.28 | 4.17 | 3.21 | 3.25 | 2.94 | 2.76 | 3.69 | 5.23 | 4.16 | 4.71 | 4.49 | 2.93 | 5.18 | 4.38 | - |
| cm | 0.26 | 0.13 | 0.11 | 0.10 | 0.03 | 0.01 | 0.06 | 0.07 | 0.12 | - | 0.15 | 0.03 | - | - | - | - | - | 0.33 | 0.14 | - |
| hm | - | - | - | - | - | - | - | - | - | 2.77 | - | - | - | - | - | - | - | - | - | - |
| il | 1.75 | 1.92 | 1.90 | 1.54 | 1.37 | 0.95 | 1.88 | 1.61 | 1.48 | 1.18 | 2.54 | 2.81 | 1.92 | 2.28 | 1.37 | 2.37 | 1.01 | 2.43 | 1.92 | 1.29 |
| ap | 1.37 | 1.40 | 1.07 | 0.55 | 0.69 | 0.74 | 1.37 | 1.31 | 1.12 | 1.88 | 1.63 | 1.66 | 1.19 | 1.66 | 1.11 | 1.54 | 0.85 | 0.99 | 1.19 | 0.81 |
| cc | 0.07 | 0.16 | - | 0.11 | 0.11 | 0.00 | 0.16 | - | 0.18 | 0.07 | - | - | 0.07 | - | 0.23 | 0.11 | 0.02 | 0.05 | 0.03 | 0.05 |
| others | 0.75 | 0.87 | 0.79 | 1.66 | 1.48 | 0.99 | 0.97 | 0.66 | 0.84 | 0.63 | 0.82 | 1.00 | 0.97 | 0.96 | 1.13 | 0.63 | 0.97 | 1.04 | 0.70 | 0.80 |

Notes: Standard X-ray fluorescence and wet chemical techniques were used. $H_2O^-$ was eliminated from all but an earlier batch of analyses (Mackenzie & Chappell, 1972) by heating powdered samples at 110°C for 12 hours and then storing in a desiccator. Trace elements in 56 samples were measured by direct-reading optical spectrograph. Additional data may be obtained from the author upon request.

Cont.

Table 1 cont.

NOMENCLATURE AND LOCATION OF SPECIMENS, TABLE 1.

1. High-K olivine basalt ("shoshonite"), SE summit area, Mount Hagen.
2. High-K olivine basalt ("shoshonite"), Turuk River, 20km SE of Hagen trig.
3. High-K olivine-hornblende-2 pyroxene low-Si andesite, SE central summit area.
4. Olivine-2 pyroxene-hornblende low-Si andesite, SE summit area.
5. Hornblende-2 pyroxene andesite, central summit area.
6. Tridymite-2 pyroxene-hornblende andesite, central summit area.

7. High-K olivine basalt ("shoshonite"), flow, N summit area, Mount Giluwe.
8. High-K olivine-2 pyroxene low-Si andesite, lava dome, upper NE slopes.
9. High-K (olivine-)2 pyroxene low-Si andesite, lava dome, upper SW slopes.
10. Oxidized hornblende-2 pyroxene andesite, dome, N summit area.

11. Olivine basalt, crater area, Mount Murray.
12. Olivine basalt, northern slopes.
13. Hornblende-2 pyroxene andesite, crater area.

14. Olivine basalt, SE slopes, Mount Duau.
15. Hornblende-2 pyroxene andesite, S slopes.

16. Olivine basalt, S slopes, Mount Favenc.
17. Hornblende-2 pyroxene andesite, SE slopes.

18. Near-saturated alkali-olivine basalt, crater area, Mount Bosavi.
19. Olivine basalt, W slopes.
20. Hornblende-2 pyroxene andesite, Turama River (from crater area?).

TABLE 2. Comparison between trace element data from Papua New Guinea volcanoes and ~~from~~ calc-alkaline and high-K rocks from other parts of the world.

| | Central Papua New Guinea (this paper) | | High-potassium basalts and andesites | | | | Calc-alkaline basalts and andesites | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | Wyoming, U.S.A. (Nicholls & Carmichael 1969) | Peruvian Andes (Dupuy & Lefèvre, 1974) | Fiji (Gill 1970) | Aeolian Islands (Keller 1974)[3] | Andes (Dupuy & Lefèvre, 1974) | | | New Zealand Japan (Taylor & White, 1966) | Bougain-Ville (Taylor et al.,) 1969a) | SW Pacific (Jakeš & White, 1972) |
| | | | | | | | Peru | | Other Areas | | | |
| | Basalts[1] | Andesites[2] | | | | | High-K | Low-K | | | | |
| Rb | 18-108 | 23-108 | 35-180 | 40-100 | 22-75 (187) | 50-100 | 80-200 | 40-100 | 20-21 | 9-44 | 24-88 | 10-45 |
| Sr | 466-1473 | 488-1682 | 600-1500 | 700-1100 | 645-1550 (2010) | 1030-1295 | 300-600 | 290-740 | 480-800 | 235-620 | 550-810 | 330-460 |
| Ba | 300-1048 | 340-1360 | 1360-3900 | 1100-1900 | 455-990 | 700-1000 | 800-1200 | 840-1170 | 400-700 | 140-340 | 180-520 | 115-520 |
| Pb | 4.0-17.5 | 8.0-20 | - | - | 5-13 | - | - | - | - | 3.5-13 | 1.7-7.2 | |
| La | 9.6-78 | 13-65 | 40-80 | - | 14, 8.5 | 27-40 | - | - | - | 9-27 | 7.8-14 | 10-14 |
| Ce | 17-117 | 21-91 | 50-110 | - | 29, 14 | 53.94 | - | - | - | 16-43 | 14-26 | 19-24 |
| Ni | 0-50; 100-340[4] | 3-85(130) | 45-75 (190) | - | 4-25 (92,110) | 11-38 | - | - | - | 16 84 | 3-13 | 5-25 |
| Cr | 10-115; 140-776[4] | 0-114(274) | 90-415 | - | 2-72 (220,360) | 24-150 | - | - | - | 26-220 (480) | 3-33 | 13-4 |
| V | 140-300(370) | 60-200(280) | 100-185 | - | 180-350(670) (177-243) | 200 | - | - | - | 140-170 (215) | 68-175[5] 175-245 | 68-255 |
| K/Rb | 225-400 (660) | | | 340-550 | 431-485 (575) | 300-366 | 200-380 | 250-600 | - | 236-347 (N.Z.) 332-510 (J.) | 300-700 | 340-430 |
| $Sr^{87}/Sr^{86}$ | 0.7037 to 0.7059[6] | | | | 0.7037-0.7045 | 0.703-0.7064 | | | | | | |

Notes: numbers in brackets are extreme values

1. including basaltic/low-Si andesites
2. including dacite
3. "trachybasalts" and "trachyandesites" only
4. Split into 2 groups: low-Si andesites; basalts
5. " " " " " " and andesites (top)
6. Unpublished determinations by R.W. Page, Australian National University, Canberra; also Page and Johnson (1974).

Figure captions

Fig. 1. Locality map, showing distribution of late Cainozoic volcanic centres in central Papua New Guinea, and principal geological features.

Fig. 2. MgO *versus* FeO diagram (wt %).

Fig. 3. Normalized molar Mg-number *versus* wt % $SiO_2$. "All others" includes Bosavi andesites.

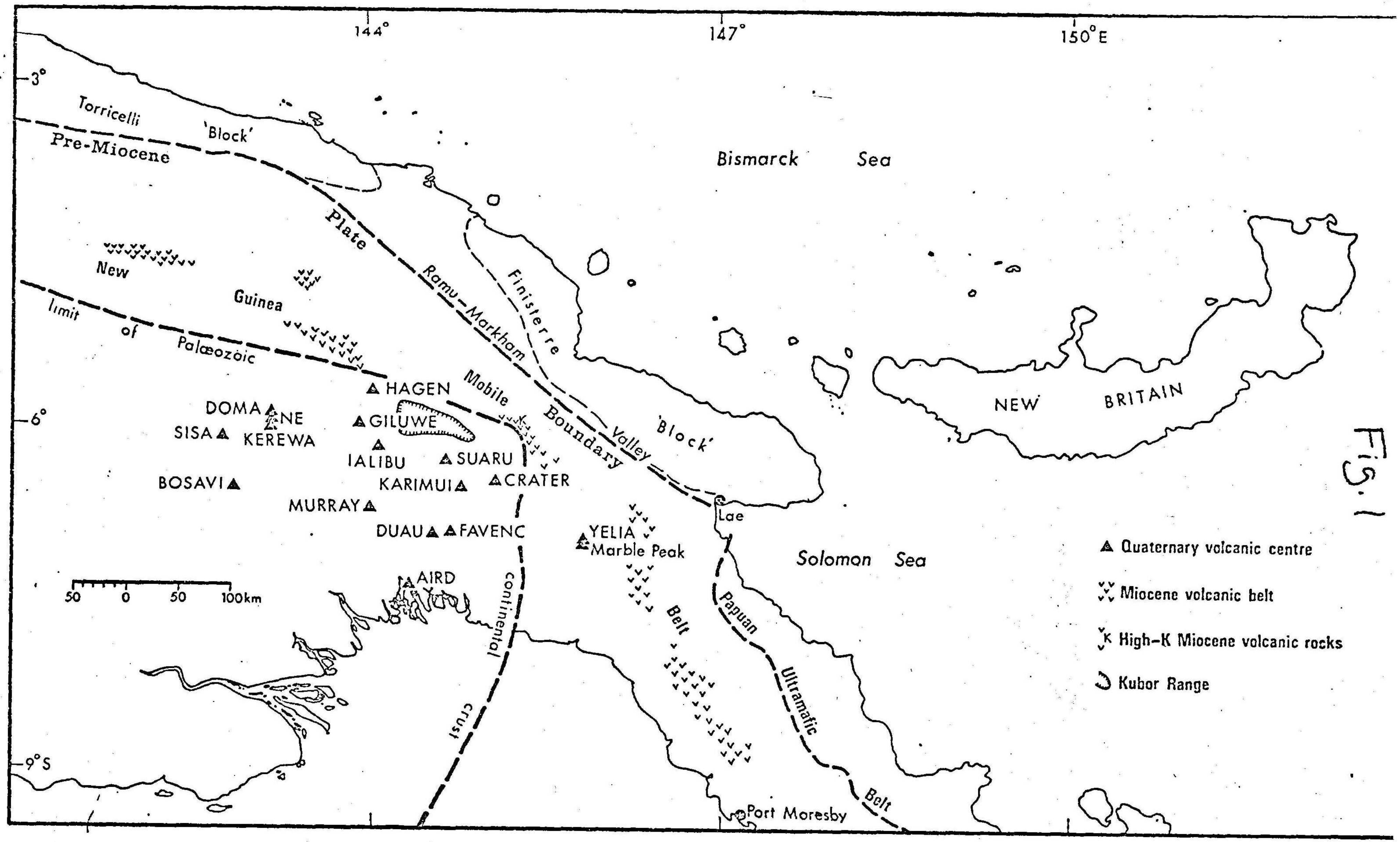
Fig. 4. $Na_2O + K_2O$ *versus* $SiO_2$ diagram (wt %). Groundmass ~~glass~~ δ compositions (open symbols in top diagrams) were obtained using a non-dispersive solid-state detector on a JEOL microprobe and manually scanning a defocussed beam.

Fig. 5. $K_2O$ *versus* $Na_2O$ diagram (wt %).

Fig. 6. $K_2O$ *versus* $SiO_2$ diagram (wt %). Symbols as for Fig. 5.

Fig. 7. $TiO_2$ *versus* $SiO_2$ diagram (wt %). Solid line encloses points representing Giluwe rocks. "All others" includes Bosavi andesites.

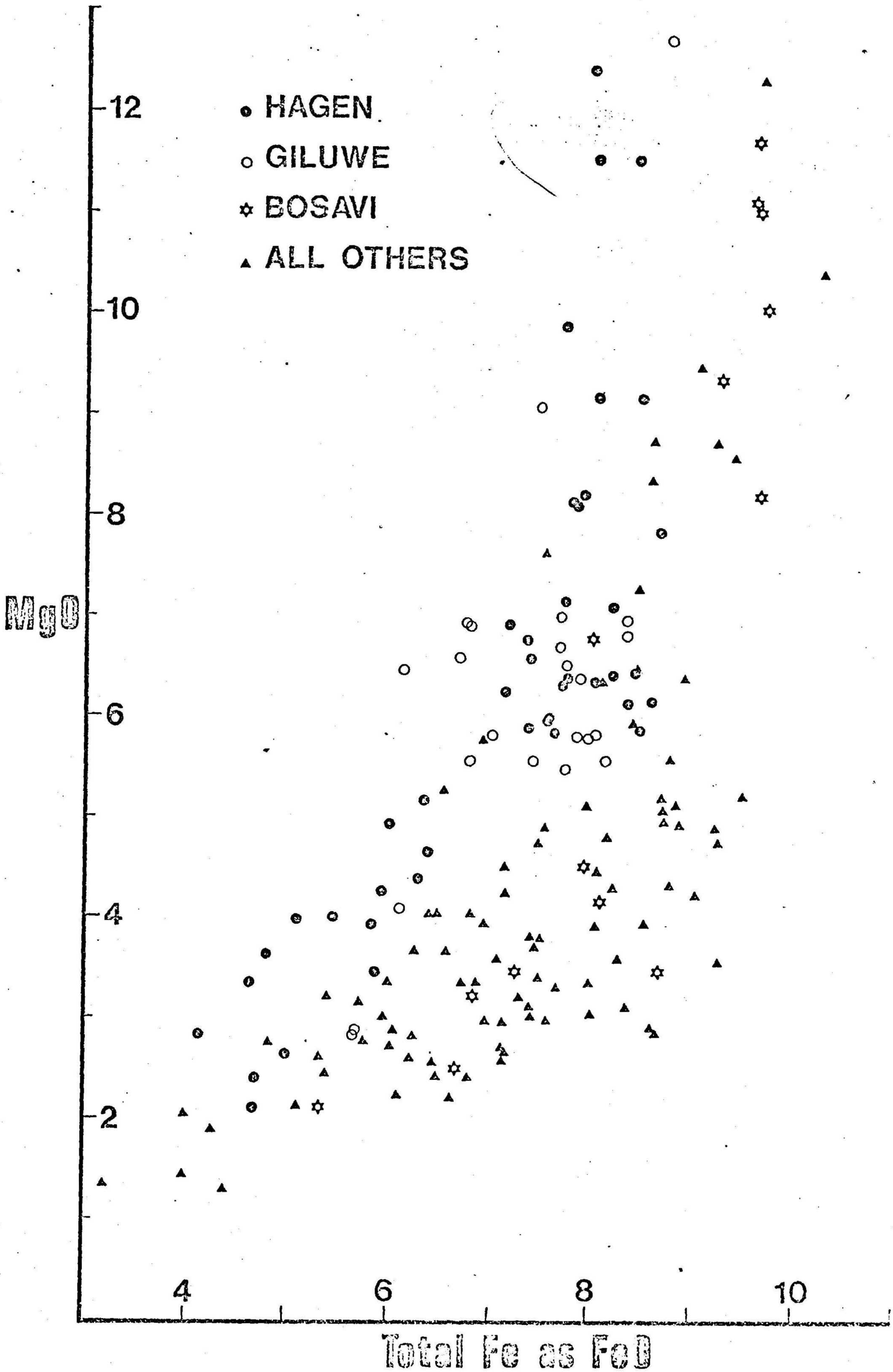
#15



144°
147°
150°E
3°
6°
9°S
Torricelli 'Block'
Pre-Miocene Plate Boundary
Ramu-Markham Valley
Finisterre 'Block'
Mobile Belt
Bismarck Sea
New Guinea
limit of Palæozoic
NEW BRITAIN
HAGEN
DOMA
NE
GILUWE
SISA
KEREWA
IALIBU
SUARU
CRATER
BOSAVI
KARIMUI
MURRAY
DUAU
FAVENC
YELIA
Marble Peak
Lae
Solomon Sea
AIRD
50 0 50 100km
continental crust
Papuan Ultramafic Belt
Port Moresby
Quaternary volcanic centre
Miocene volcanic belt
High-K Miocene volcanic rocks
Kubor Range


Fig. 1

Fig. 2

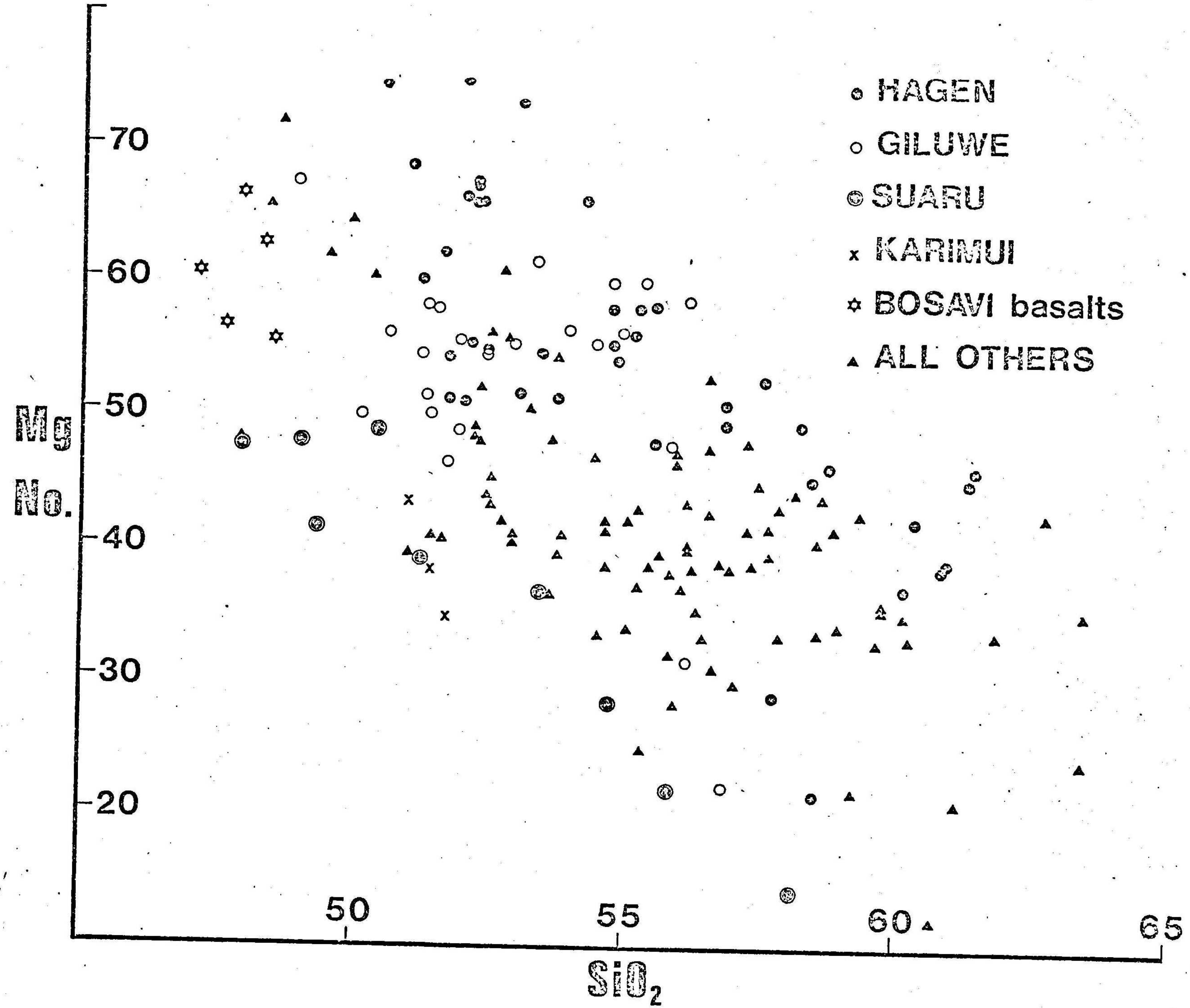



A

C

B

Reduce A-B (27cm) to A-C (14cm)



Mg No.
70
60
50
40
30
20
SiO₂
50
55
60
65
HAGEN
GILUWE
SUARU
KARIMUI
BOSAVI basalts
ALL OTHERS


Fig 3

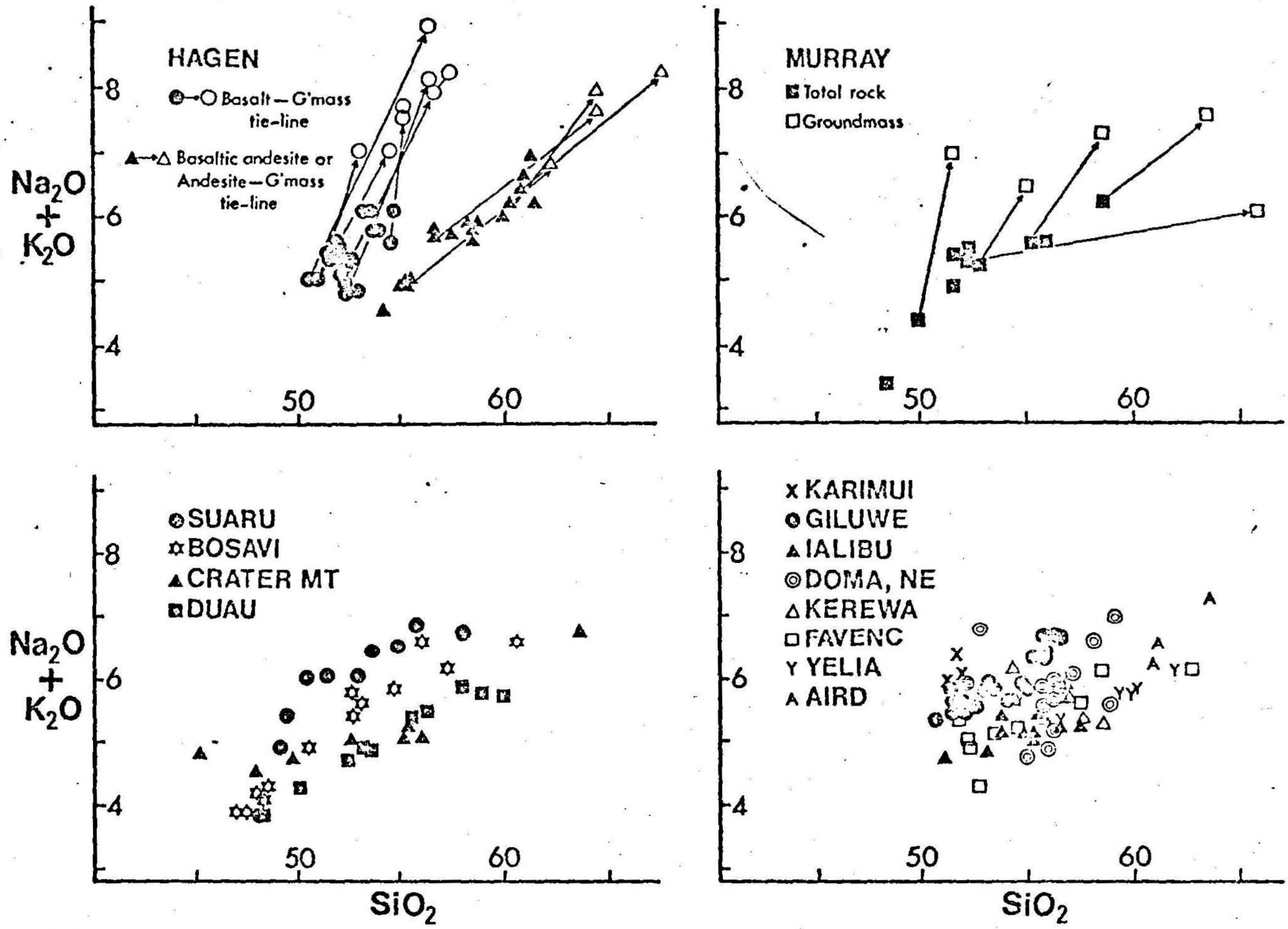



Fig. 4

A

C

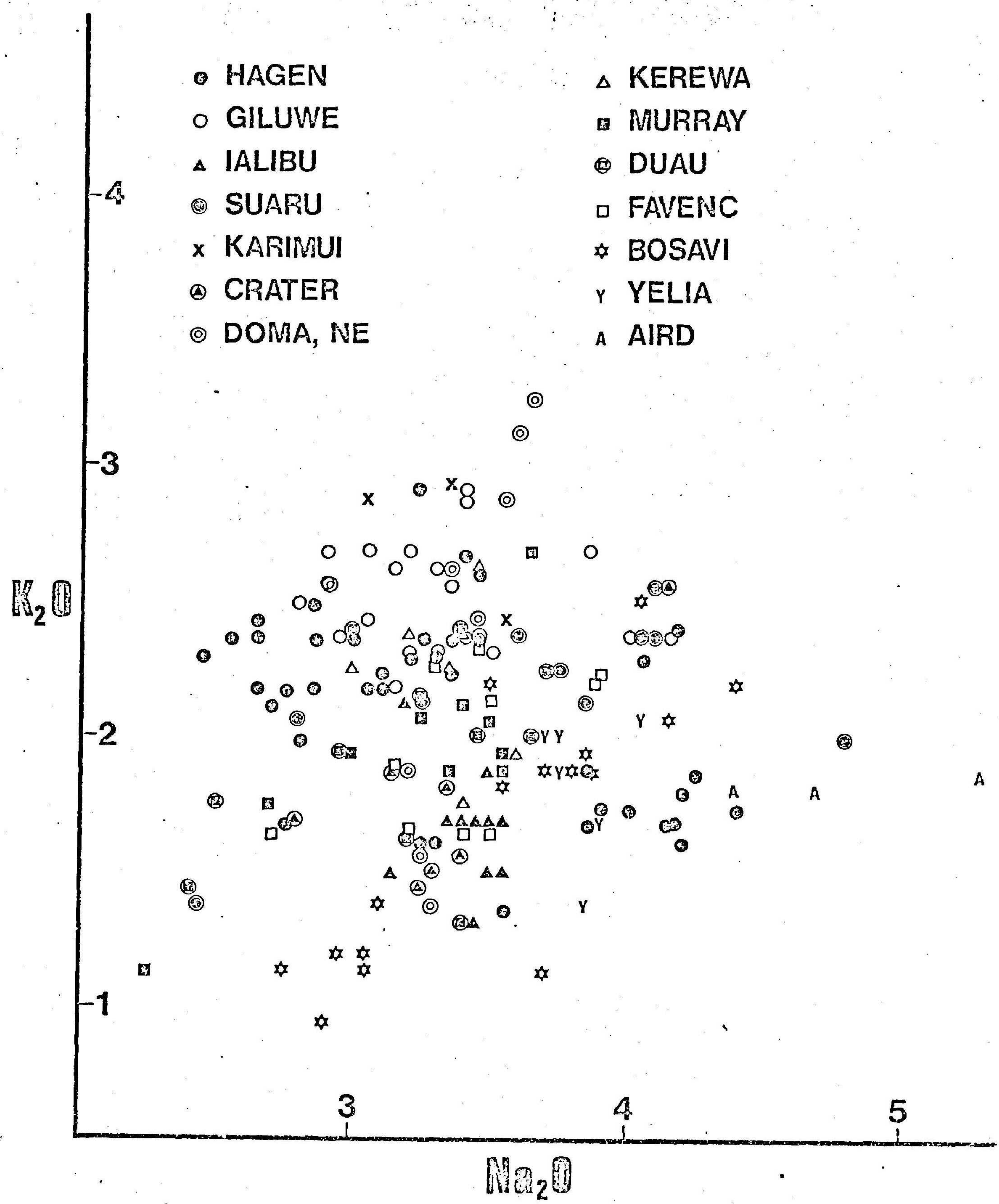
REDUCE A-B to A-C

17.1 cm 12 cm

Fig 5

A C

REDUCE A-B (18.5cm) to A-C (12cm)



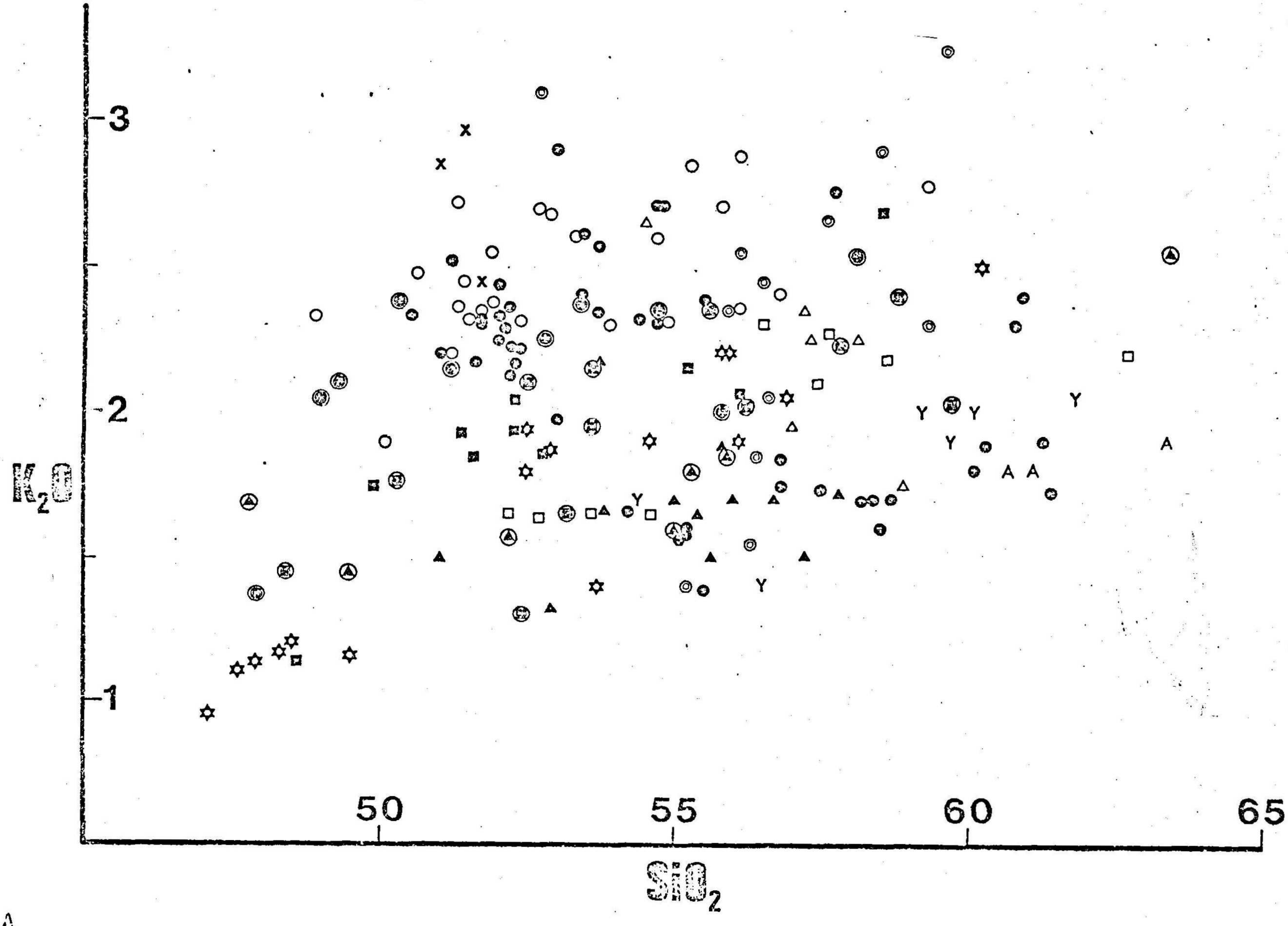
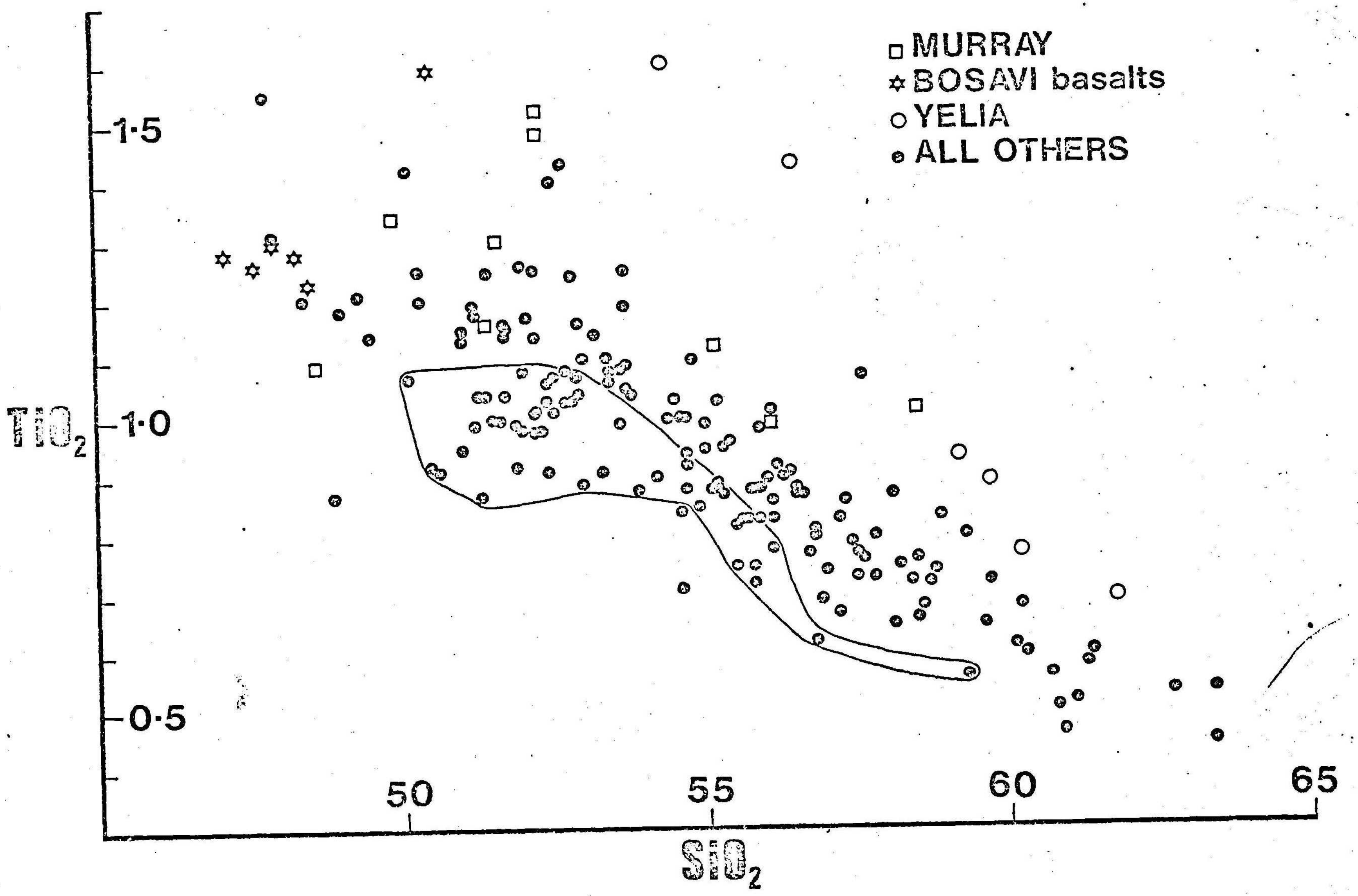



Fig 6.


MURRAY
BOSAVI basalts
YELIA
ALL OTHERS
1·5
1·0
0·5
$TiO_2$
50
55
60
65
$SiO_2$

SUMMARY OF THE 1953-57 ERUPTION OF TULUMAN VOLCANO, PAPUA NEW GUINEA

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Abstract

Tuluman volcano, just south of Lou Island in the Admiralty Group, Papua New Guinea, began erupting in June 1953. Initial activity was submarine, but later the volcano gradually built up to sea-level. Eruptions migrated between several centres which were numbered in order of their initial activity. The vents that reached the surface subsequently developed explosive phases, the most violent of which occurred in February-March, 1955. After a series of submarine effusive, surface explosive, and later subaerial effusive phases, the eruption finished in January 1957.

The eruption produced a small island, Tuluman Island, and a rocky islet to the northwest, composed of lava and pyroclastic deposits. Rock types were silica-rich and included highly vesiculated rhyolite and pumice, commonly with bands of black obsidian.

Although some of the explosive outbursts were strong, the overall pattern was mild when compared with eruptions elsewhere in Papua New Guinea. No warning was given of the initial eruption and no earth tremors were felt. Damage to nearby islands was negligible.

# 20

Introduction

Tuluman volcano is a dominantly submarine volcano, about 1 km south of the southwestern tip of Lou Island, a member of the Admiralty Group of Islands in the northern Bismarck Sea, Papua New Guinea (Fig. 1). Manus Island is the largest and best-known of the Admiralty Group; its main centre and airport is at Lorengau, 45 km northnorthwest of Tuluman. ~~(Fig. 1).~~

Tuluman and Lou Island are on the northwestern side of St Andrew Strait. The southeastern side is made up of a line of islands from Baluan in the south through Mok, Pam, and the St Andrew Islands, to the Fedarb Islands in the northeast. ~~(Fig. 1).~~ Except for the St Andrew Islands, which are composed of coral, all the islands adjoining the Strait are volcanic, and it is likely that the St Andrew Islands also have volcanic rock basements (Johnson & Smith, 1974).

Fig 1

Lou is the largest of the islands in St Andrew Strait and is formed by a chain of overlapping Quaternary volcanic cones. Admiralty Chart Aus. 054 (printed March 1944) shows a shoal in the area to the southwest of Lou near where Tuluman volcano now exists, and this could be the centre of the activity witnessed in 1883 by Miklouho-Maclay (1885). However, no islands or shallow sea bottom were indicated at Tuluman at the time of the first eruption in 1953. Neither did recorded seismicity indicate any local build-up of tectonic stress, or the formation or emplacement of magma in the crust. This seems unusual after a period of quiescence of some 70 years.

Submarine activity began at Tuluman on 27 June, 1953. No earth tremors were felt on nearby islands either before or during the eruption, although the people of Pam Mandian Island ~~(Fig. 1)~~ reported small sea waves which caused minor damage to some canoes after the initial explosion on 27 June. The waves may have been triggered by tremors, or by a small submarine explosion when lava first breached the sea bottom.

The eruption, more accurately described as a series of eruptions of different intensities, continued until 28 January, 1957. Reynolds and Best (1957; in prep.) identified eight principal eruptive centres and referred to them as cones 1 to 6, 7A and 7B, in order of initial activity. When last visited and mapped in 1971, Tuluman volcano was represented by two islands: the largest, Tuluman Island, was about 0.5 km in diameter and up to 23 m above sea-level; the second was a small rocky islet, 400 m to the northwest, which has been called "cone-3 island" (Johnson & Smith, 1974).

The products of the 1953-57 eruption were rhyolitic. Their petrology was discussed by Johnson & Smith (1974), who suggested a derivation from the melting of crust (about 25 km thick below St Andrew Strait). The rubidium/strontium ratios of the Tuluman lavas are exceptionally high compared with those of volcanic rocks from other sources in Papua New Guinea (Page & Johnson, 1974).

The 1953-57 eruption was a complex series of submarine and subaerial, effusive and explosive events. A detailed chronology of these events has been made by Reynolds & Best (in prep.) as a supplement to their early report (1957) which dealt with events only up to August, 1955. This paper presents a summary of the main events and phases of activity; it identifies three main types of activity, and briefly describes the general pattern of the eruption.

Outline of events

27 June - 6 July, 1953: The first activity observed was the voluminous emission of water vapour at the location designated "cone 1" (Fig. 2).

The vapour was generated by the explosion, or fragmentation, of masses of lava which had flaked off submarine lava flows, or which were derived directly from the vent, and floated to the sea surface.

14 November, 1953 - 18 February, 1954: A similar type of submarine activity took place from another location, "cone 2", about 700 m southwest of cone 1. Aerial concussion effects were felt as tremors in buildings on Baluan Island to the south.

Fig 2

9-27 July, 1954: Renewed submarine activity began at cone 2 and continued until a dome of lava built up to sea-level on 11 July. Activity then declined at cone 2, but intensified at a new submarine centre 900 m northwest. This centre built up cone 3 which reached sea-level on 13 July. On 14 July, the first major explosive phase began at cone 3 (Fig. 3) but soon declined, followed by vapour emission. Lava flow resumed at cone 2 on 21 July from a small vent at sea-level and moved outwards in a series of slow-moving concentric waves. Activity finished on 27 July leaving a prominent island at cone 3, consisting of a lava base overlain by a mound of pyroclastic material about 12 m high, and three small rocky shoals or islets up to 3 m high at cone 2.

Fig 3

20 October - 6 November, 1954: Submarine activity took place near cone 1.

10-15 February, 1955: People travelling between Manus and Baluan Islands in January and early February noticed that a slight, but definite, emergence of the shoals at the summit of cone 2 to about 8 m above sea-level had occurred. Strong explosions began at cone 2 on 10 February, but by 11 February it became clear that the major activity was from a centre west of, but adjacent to, cone 2. This new centre was called cone 4. Both cones built up a single island. In addition to the periodic violent explosive outbursts, with ejection of pyroclasts and release of large volumes of gas and vapour, jets of black dust and gas were emitted from the main vents with loud roars from time to time.

11 February - 11 March, 1955: Submarine activity at another centre, cone 5, southwest of cone 4, started on 11 February. It ceased on 13th, but had recommenced by 16th, and another island appeared on 23 February. A series of explosive and effusive phases followed, culminating with violent explosions on 9 March that rocked buildings on Baluan Island to the south, and deposited more than 2 cm of dust on the Pam Islands. The islands of cone 5 and cones 2 and 4 were joined by a low saddle of pyroclastic deposits (Fig. 2A).

12 March - 9 May, 1955: A new submarine source, cone 6, became active about 550 m south of cone 5. Varying rates of lava flow were indicated by the changes in volume of material reaching sea-surface; the flows continued to 9 May. The cone did not reach sea-level.

16 May - 26 June, 1955: Minor explosions, rumbling, and the usual surface manifestations of previous submarine activity were observed in the vicinity of cone 1. A lava dome reached sea-level by the morning of 6 June, and a brief period of explosions followed. Steady vapour emission continued from 20th to 26th when the dome subsided below sea-level. Eruptions of mud were observed at cone 2 (later mud eruptions also took place spasmodically at cone 4 between July and October).

26-28 September, 3-7 October, 1955: Short periods of submarine lava effusion were indicated by the appearance of lava at sea-level east of cone 2. This is attributed to early activity from cone 7.

7 November, 1956 - 28 January, 1957: Following increased vapour emission from north of cone 2, a small cone-shaped island, cone 7A, was built to 3 m above sea-level on 28 November (Fig. 4). Spasmodic explosions and jet-like blasts of vapour and dust characterised this phase of the

Fig. 4

eruption. On 30 November, a short period of lava flow began between the new island and cone 2. This was from another vent - probably belonging to the same cone as 7A, and designated 7B. Mud eruptions were also seen at cone 4 at this time. Activity at 7A continued until about 11 December. As this phase declined, extrusion of lava flows was resumed at 7B and gradually filled the area between the islands of 2 and 7A above sea-level (Fig. 2B). Spasmodic explosions continued into January 1957, but the eruption had ceased by the 28th. The lava pile above sea-level was approximately 430 m by 300 m, and 30 m high at the crater rim; Fig 5 the average height of the lava surface above sea-level was 20 m (Fig. 5).

Subsequent tilting to the northeast, and rapid erosion, resulted in the early removal of the remnants of cone 5, and finally left a large island made of the products of cones 2, 4 and 7, and a small lava island above cone 3 (cf. Fig. 2C).

The main activity outlined above, and some other minor Table 1 events, are listed on an individual cone basis in Table 1.

Types of activity

The eruption consisted of three main types of activity: (1) the extrusion of flows onto the sea floor, building cones, most of which reached the sea surface; (2) ~~explosive events~~ explosions from subaerial vents; and s (3) the extrusion of subaerial lava flows.

(1) Submarine extrusion

This submarine activity produced gas-charged masses of lava which floated to the surface. These lava masses were highly vesiculated, and had red-hot interiors and chilled peripheries; they presumably flaked off from submarine lava flows during cone-building on the seabed.

The arrival of the masses at the surface could usually be expected at intervals of about ten minutes, although there was no regular periodicity.

The reduction in hydrostatic pressure when the masses reached sea-level permitted vesiculation, sometimes explosively. The water vapour generated, and the escaping gases, formed clouds above the remnants of the masses. Most of the lava masses sank within a few minutes of reaching the surface, but some drifted as vesicular lava or pumice fragments for many kilometres. The greater part of the floating masses was submerged.

Submarine activity was probably restrained by the rapid cooling of the lava and the confining hydrostatic pressure of the sea, and possibly by the load effects imposed by the sea-floor sediments and lava accumulation on the flanks of the cone. Submarine activity occurred as one or two individual phases at each of the main vents until the cones reached the surface of the sea. The lava flows appear to have built up to 3 to 4 metres above sea-level by the close of each submarine phase.

(2) Surface explosive activity

A phase of explosive activity took place at each centre after the cone had reached the surface (except for the extrusive phase at cone 7B towards the end of the eruption). Clouds of vapour and ash ascended to heights of 2000 m above sea-level or up to 5000 m in a calm atmosphere; larger debris (mainly red-hot bombs, coarse glassy and pumiceous fragments) fell around the cone, and lapilli rained down over a radius of more than 750 m. Bright yellow, orange and brick-red plumes commonly accompanied the vapour clouds. Gases such as sulphur dioxide and hydrogen sulphide were recognized in small quantities, but

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Although the summit of cone 7 emerged above sea-level at centre 7A a number of factors combined to preclude the formation of a prominent island: the summit was small by comparison with those of other cones, the volume of ejected material from the 7A vent was not as great, and much longer periods elapsed between explosions. The activity from November to December 1956 at cone 7 represented a declining phase of the explosive type of activity.

(3) Subaerial extrusion

The final and main surface effusive phase of the Tuluman eruption began in late December 1956 between the island of cones 2 and 4 and the peak of cone 7. The actual time at which activity changed to effusive is not clear. The explosive phase at cone 7 declined after 11 December, but according to observations made from Baluan Island, vapour emission appeared to increase. By 21 December it was apparent that a separate crater (crater B) of cone 7 had formed to the south of the explosive crater (crater A), and that lava was flowing from this new source. Observations were impaired by the generation of dense clouds of water vapour, and the vent was obviously close to sea-level at this time. By 28 January, 1957, the lava island embracing cone 7 and the island of cones 2 and 4 had formed. Local observations during and after the effusive phase, and aerial inspections indicate that the outpouring of lava was intermittent and that some minor spasmodic explosions occurred.

Later inspections showed that the flows consisted of hard, slightly vesicular obsidian and dark grey vesicular rhyolite. Blocks of lava up to 2 m across were strewn over the surface of the island; these may have been ejectamenta, but some may have spalled off from the flows. Masses of grey pumice with black obsidian bands which show flow lamination and in parts marked convolution were common. The products of the surface effusive activity were apparently very similar to those of earlier submarine phases and to the bombs ejected during the major explosive periods.

The brief period of lava flow at sea-level at cone 2 from 21 to 27 July, 1954, is regarded as the last phase of effusive activity which, up to that time, had been entirely submarine.

Although not observed at the time, a brief period of surface effusive activity at cone 5 took place in early March 1955 during what is regarded as the most violent explosive phase of the eruption. An inspection of the Tuluman Islands shortly afterwards (24 March) showed remnants of an original crater wall of cone 5 composed of pyroclastic debris, and a smaller inner crater the sides of which contained agglomerate interbedded with lava flows.

Pattern of eruption

The most violent phase of activity appears to have been the surface explosive period from cone 4 and 5 in February to early March, 1955. This seems to have been the zenith of the eruption and was preceded by uplift of cone 2 - the only premonitory warning of any of the major phases of activity. This climax took place 21 months after the beginning of the eruption in June 1953. The first 18 months of the eruption produced four major detectable submarine effusive phases, and a further period of submarine activity accompanied the eruption at cone 5; the first subaerial explosive phase took place in mid-1954.

Apart from the submarine lava flows from cone 6, which immediately followed the violent explosive phase at cone 5, other activity in 1955 was of much lower intensity. A quiescent period of 12 months preceded the final stage of the eruption at cone 7 which began on 7 November, 1956. The major subaerial effusive phase occurred in this final period.

The overall pattern of the eruption, as witnessed, therefore appears to have been a build-up period as the volcano approached the surface of the sea (shown mainly by submarine effusive phases for some 20 months), an apparently climactic explosive phase over 2 months, and a period of decreasing intensity for the next 22 months. Although it was prolonged, the eruption was mild compared with some of the other eruptions with known histories elsewhere in Papua New Guinea.

Acknowledgements

The history of the 1953-57 eruption of Tuluman volcano could not have been written without the invaluable data supplied by village people, airline pilots, trawler masters, and particularly the Administration officers (Messrs J. Landman, E.G. Hicks, and W. Murdoch) who were stationed at Baluan Patrol Post during that period.

We are both grateful for the privilege of working with Tony Taylor in the period following the 1951 Lamington eruption, and in the formative years of the Vulcanological Observatory at Rabaul. His enthusiasm was an inspiration, and many of our ideas were guided by his concepts of the mechanics of eruption in island-arcs.

Permission to publish this summary has been given by the Director, Bureau of Mineral Resources, Geology and Geophysics, Canberra, Australia

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P.C.

## Table 1. Phases of activity at Tuluman Volcano, 1953-57

| Cone No. | Period | Weeks | Type of activity |
|---|---|---|---|
| 1 | June 27-July 6, 1953<br>Oct. 20-Nov. 6, 1954<br>May 16-June 6, 1955<br>June 6-20, 1955<br>Sept. 20, 1955 | 1½<br>2½<br>3<br>2<br> | *Submarine**<br>*Submarine*<br>*Submarine to surface*<br>Minor explosive & vapour<br>Vapour emission |
| 2 | Nov. 14, 1953-Feb. 18, 1954<br>July 9-11 & 21-27, 1954<br>Feb. 10-12, 1955 | 14<br>1½<br>½ | *Submarine* (intermittent)<br>*Submarine* to surface<br>Minor explosive |
| 3 | July 11-14, 1954 | ½ | *Surface explosive* |
| 4 | Feb. 11-15, 1955<br>July 21-Aug. 2, 1955<br>Oct. - Nov., 1955<br>Nov. 24-Dec. 4, 1956 | ½<br>2<br>1 to 4<br>1 | *Surface explosive*<br>Minor explosions & vapour<br>" "<br>" " |
| 5 | Feb. 11-13 & 16-23, 1955<br>Feb. 24-Mar. 11, 1955 | 1½<br>2 | *Submarine* to surface<br>*Surface explosive* & effusive |
| 6 | Mar. 12-May 9, 1955 | 9 | *Submarine* (intermittent) |
| 7? | Sept. 26-28, Oct. 3-7, 1955 | 1 | *Submarine* |
| 7A | Nov. 7-26, 1956<br>Nov. 27-Dec. 11, 1956 | 3<br>2 | Mainly vapour emission<br>*Surface explosive* |
| 7B | Nov. 30-Dec. 2, 1956<br>Dec. 11, 1956-Jan. 28, 1957 | ½<br>7 | *Submarine*<br>*Surface effusive* & minor explosions |

* All submarine activity was effusive; stronger phases are underlined.

Figure captions

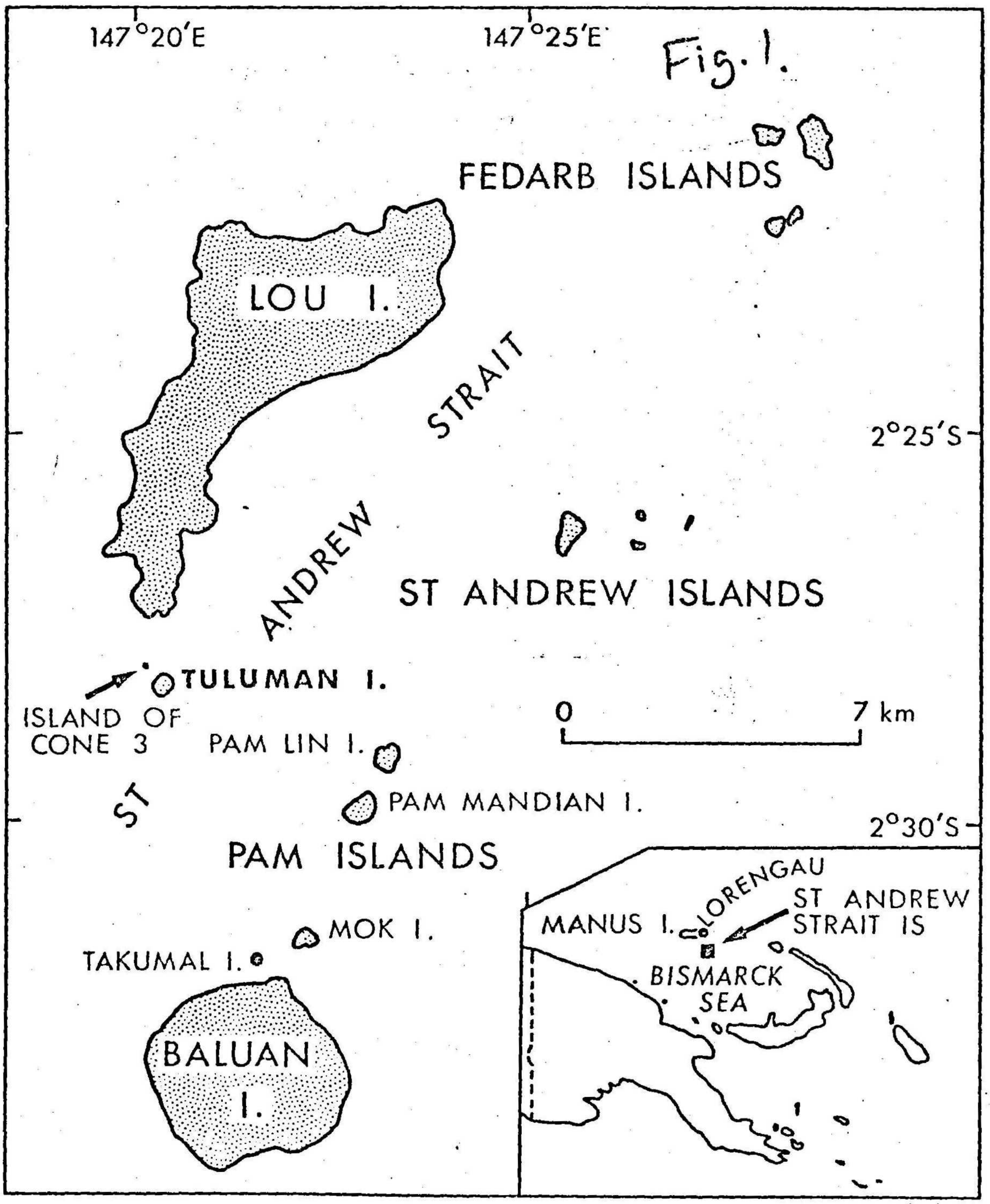
Figure 1. Islands of St Andrew Strait, Papua New Guinea.

Figure 2. Comparison of Tuluman islands and cones, 1955-1971. A: March, 1955 (adapted from Plate 4 of Reynolds and Best, 1957). B: February, 1957 (prepared from photographs taken during aerial inspection; position of cones doubtful). C: From plane table survey by J. Herlihy and F.E. Decker in June, 1962 (broken lines), and modified by R.W. Johnson and R.A. Davies by mapping in July, 1971.

Figure 3. Activity at cone 3, 14 July, 1954, with small horse-shoe island of cone 2 (temporarily dormant) in foreground. Southern tip of Lou Island at right. Aerial view from east.

Figure 4. Activity at cone 7, 28 November, 1956, with island of cones 2 and 4 in foreground and small remnant island of cone 3 on the left. Densely wooded southern tip of Lou Island at back. Aerial view from south.

Figure 5. The Tuluman islands, 12 January, 1957, with lava from cone 7B joining islands of cones 2 and 4; island of cone 3 in foreground. Aerial view from northwest.


147°20'E
147°25'E
Fig. 1.
FEDARB ISLANDS
LOU I.
ST ANDREW STRAIT
2°25'S
ST ANDREW ISLANDS
TULUMAN I.
ISLAND OF CONE 3
PAM LIN I.
0
7 km
PAM MANDIAN I.
2°30'S
PAM ISLANDS
LORENGAU
MANUS I.
ST ANDREW STRAIT IS
BISMARCK SEA
MOK I.
TAKUMAL I.
BALUAN I.

P/A/520

Fig. 3
Neg. no.
G.B/454

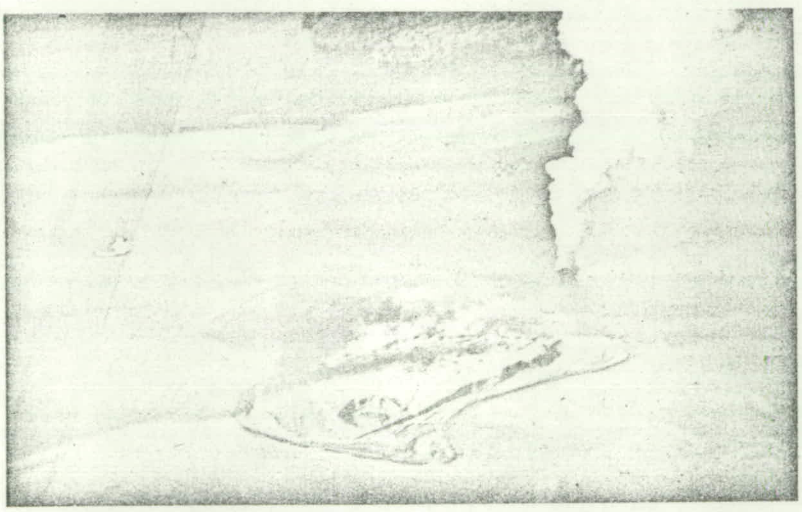


Fig. 4
Neg. no
MRG/712

Fig. 5
Neg. no.
MRG/732.

# FELDSPATHOID-BEARING POTASSIC ROCKS AND ASSOCIATED TYPES FROM VOLCANIC ISLANDS OFF THE COAST OF NEW IRELAND, PAPUA NEW GUINEA: A PRELIMINARY ACCOUNT OF GEOLOGY AND PETROLOGY

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ABSTRACT

The volcanic rocks of the Tabar, Lihir, Tanga, and Feni Islands range from pre-Middle Miocene to Pleistocene, and show an exceptionally wide range of compositions. Strongly silica-undersaturated types are basanites, tephrites, olivine nephelinites, phonolitic tephrites, and tephritic phonolites; these are especially common on the Tanga and Feni Islands. Less undersaturated and silica-saturated types are alkali and "transitional" basalts, trachybasalts, and trachyandesites; these are well-represented on the Tabar and Lihir Islands. Quartz trachytes have been found on all the island groups except Lihir, and undersaturated trachytes and phonolitic trachytes are present in the Tanga group. Tholeiitic basalts, andesites, and dacites all appear to be absent. $K_2O/Na_2O$ values are mainly between 0.5 and 1, and $TiO_2$ contents are notably low (mainly <1 percent). The tectonic conditions which led to the development of these magmas are uncertain. The magmas may be related to a slab of downgoing lithosphere, but the islands are not underlaid by a zone of intermediate or deep focus earthquakes at the present day. An alternative model involving an isostatically controlled fault zone beneath the islands is being explored.

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#21

## INTRODUCTION

Strongly alkaline rocks are rare among the Late Cainozoic volcanoes of Papua New Guinea, and appear to be restricted to the Tabar-to-Feni Islands, which form a narrow, 260 km-long chain off the coast of New Ireland (Fig. 1). The rocks of these islands also show an extremely wide range of compositions (bearing in mind the limited size and number of the islands, Fig. 1), ranging from the strongly undersaturated types (containing a maximum of almost 25 percent normative nepheline) to those near the critical plane of silica undersaturation in the "basalt tetrahedron" (Yoder and Tilley, 1962), to quartz-normative salic compositions.

The islands comprise four groups - Tabar, Lihir, Tanga, and Feni - which are equally spaced about 75 km apart (Fig. 1). The Tabar group in the northwest is closest to New Ireland (about 30 km away), and the Tanga and Feni Islands are furthest away (about 55 km). The Green Islands (Nissan and Pinipel) lie 75 km to the southeast on the extension of the Tabar-to-Feni line (Fig. 1), but they are raised coral atolls and show no volcanic rocks (except for drift pumice on beaches). The line continues southeastwards to Bougainville Island whose Late Cainozoic volcanoes have rocks of calcalkaline compositions but apparently none of alkaline type (Blake and Miezitis, 1967; Taylor et al., 1969; Bultitude, this volume). To the northwest, the Tabar-to-Feni line is roughly co-linear with the St Matthias Group which consists mainly of limestones, although volcanic rocks are known from Mussau Island (White and Warin, 1964).

Fig 1

## GEOLOGICAL INVESTIGATIONS

Glaessner (1915) was the first to draw attention to the alkaline character of rocks from the Tabar-to-Feni Islands. He identified a leucite-like mineral in a "trachydolerite" from Lihir Island, nosean and aegirine-augite in "glass-rich trachydolerites" from Ambitle Island, and aegirine-augite in a "limburgite trachydolerite" of high total-alkali content, also from Ambitle Island. It appears that for fifty years Glaessner's paper was the only record of alkaline rocks in the Melanesian region - until Allen and Deans (1965) described rocks of alnöitic and kimberlitic affinities from Malaita Island, British Solomon Islands.

G.A.M. Taylor was the first to undertake systematic geological work on the islands. He surveyed them in 1969 and 1971, made comprehensive collections of rocks from all the island groups, and obtained many major element whole-rock analyses (see, for example, Johnson et al., 1973). Wallace and I.A. Crick re-surveyed the islands in 1973, and Wallace and Johnson re-visited the islands in 1974.

This paper discusses 44 major element analyses of rocks collected by Taylor, and 40 analyses of rocks from the Wallace-and-Crick surveys. It briefly describes the geology of the islands and their environment, and discusses the mineralogy of 11 Wallace-and-Crick rocks. Johnson is responsible for the processing and plotting of the whole-rock analyses, Wallace is primarily responsible for the geology, and Ellis obtained the mineral analyses. Other reports are in preparation which will provide detailed descriptions of geology and petrology and give more detailed attention to problems of magma genesis.

## PLATE BOUNDARIES AND CRUSTAL STRUCTURE

The tectonic setting of the Tabar-to-Feni Islands is uncertain, despite recent geological mapping and geophysical surveying of the surrounding region (BMR, 1972; Finlayson and Cull, 1973; Murauchi et al., 1973; Furumoto et al., in prep.; Gulf, 1973).

Each of three recent papers on plate tectonics in Papua New Guinea has provided a different interpretation of present-day tectonic activity in the New Ireland region. Johnson and Molnar (1972) proposed that a left-lateral transcurrent plate boundary runs west of, and parallel to, the west coast of New Ireland. Curtis (1973), on the other hand, believed a present-day plate boundary closely follows a narrow zone of deep water that runs west-east north of Manus Island, and down the east side of New Ireland - between the Tabar-to-Feni Islands and the line joining Lyra Reef, the Nuguria Islands, and the Carteret Islands (Fig. 1; Mammerickx et al., 1971). However, present-day seismicity east, and immediately west, of New Ireland is very infrequent (see Denham, 1969, 1973), and it is therefore considered that Krause (1973) was probably justified in choosing not to recognise a present-day plate boundary in this region.

Seismic refraction studies in the region report the following crustal thicknesses: 28-30 km beneath the Feni group (Finlayson and Cull, 1973); 18-19 km halfway between the Green Islands and the southern tip of New Ireland (Murauchi et al., 1973) where the crust is up to 35 km thick (Finlayson and Cull, 1973); 16.2-17.4 km in the channel between New Ireland and the Tanga and Feni Islands (Furumoto et al., in prep.). Further east, refraction profiles across parts of the Ontong-Java Plateau (which begins about 190 km east of the Feni group) give depths to the Moho from 35 km to well over 40 km (Furumoto et al., 1973, in prep.). These data are insufficient to characterise the thickness and structure of the crust beneath the entire Tabar-to-Feni chain, although they suggest the crust there is thinner than that beneath New Ireland, and thicker than that of true oceanic crust found in the Pacific basin.

A southwest-northeast seismic reflection profile run between the southern end of New Ireland and the Ontong-Java Plateau by Furumoto et al. (in prep.) shows a series of horst- and graben-like structures for much of its length. It suggests that the Green Islands and the narrow zone of deep water east of them occupy a zone controlled by normal faulting (see also Gulf, 1973). The seismic reflection profiles by Gulf (1973) indicate substantial accumulations of sediment in the channel between New Ireland and the Tabar-to-Feni Islands, and faulting and a dome-like "basement" structure in the vicinity of the Tabar group.

SUMMARY OF GEOLOGY

The volcanic rocks of the Tabar-to-Feni Islands are Tertiary and Pleistocene. Holocene rocks have yet to be identified, and no eruptions have been reported from any of the islands, although thermal areas are present in all the island groups (especially on Lihir and Ambitle Islands; see Fisher, 1957). The islands of both the Tabar and Lihir groups form north-south lines, whereas those of the Tanga and Feni groups form roughly arcuate patterns. Many of the Tabar-to-Feni Islands have coral terraces which indicate periods of uplift and tilting throughout the Cainozoic.

The volcanic rocks of Simberi Island, the northernmost of the Tabar group, are overlaid by Middle Miocene reef limestone (micropalaeontological determinations by D.J. Belford, pers. comm., 1975). Rocks of Tertiary age are probably also present on Tatau Island, and at the north-western end of Tabar Island, but the southern part of Tabar Island is a dissected volcano which is probably Pleistocene. The K-Ar age of one rock from the northwestern part of Tabar Island is 0.986 $\pm$ 0.08 m.y. (determination on a plagioclase concentrate)*. Faulting is common in the Tabar group; one north-south fault which is especially prominent on Tabar Island may continue northwards onto Tatau Island, and even to Simberi Island. Alluvial gold was mined on a small scale in the 1920's from a stream on the northeast side of Tatau Island.

The two northernmost islands of the Lihir group - Mahur and Masahet - are raised coral reefs (with volcanic cores), but Lihir Island itself is a subaerial volcanic complex consisting of at least three major centres, all of which are probably Pleistocene. Thermal activity is widespread on the floor of a breached caldera on the east side of the island.

Except for Boang Island, which consists of raised coral reef, the islands of the Tanga group appear to be parts of the same volcano. The walls of a submerged caldera are preserved on Malendok, Tefa, and Lif Islands, and Bitlik and Bitbok Islands appear to be post-caldera extrusions. K-Ar ages

* All K-Ar dates supplied by the Australian Mineral Development Laboratories, Adelaide.

of two rocks from the Tanga group are 1.14 ± 0.08 m.y. (determined on biotite - a second determination gave 1.08 ± 0.08 m.y.) and 0.187 ± 0.02 m.y. (whole-rock age).

Ambitle (or Feni, or Anir) and Babase Islands comprise the Feni (or Anir) group. The two islands define a rough semicircle, suggesting their volcanism may have been controlled by a ring fracture. Oligocene limestone is present on Ambitle Island, but the volcanic rocks appear to be mainly Pleistocene. Two K-Ar determinations on biotite from one sample gave ages of 0.68 ± 0.10 and 0.49 ± 0.10 m.y. Ambitle Island shows an eroded central crater, open to the southwest, which contains a quartz-normative salic extrusion. Thermal areas are found throughout the central area and in the southwest of Ambitle Island. Babase Island consists of a low cone and crater (in the east), and limestone (in the west) which overlies an extrusion of similar type to the one inside the central crater of Ambitle Island. The K-Ar age of one Babase rock is 1.53 ± 0.15 (hornblende concentrate).

~~Cognate~~ Xenoliths, consisting of clinopyroxene and smaller amounts of opaque minerals, amphibole, and zoned plagioclase, are found in the lava flows of several islands. These pyroxenite inclusions range from a few millimetres to over 20 cm in diameter, and appear to be crystal cumulates.

One striking feature of the geology of some islands is the abrupt change in composition between older and younger rocks. On Tabar Island, in the Tanga group, and on Ambitle Island, earlier rocks are mainly silica-undersaturated (strongly undersaturated in Tanga and Ambitle), but the later rocks are salic and strongly quartz-normative. In these three areas, at least, the process of magma generation must have changed drastically to produce such divergent rock types.

## CHEMICAL ANALYSES

Variations in chemical compositions of the 84 analysed rocks are illustrated in Figures 2-to-7.

Some of the analysed rocks are chemically altered - especially those from the older Tertiary centres of the Tabar group - and a measure of this alteration is indicated on the variation diagrams. Sample points without a vertical dash on top of the appropriate symbol represent "fresh" rocks which contain less than 1 percent $H_2O$-, and less than 1 percent $CO_2$. The vertical dash indicates that the rock contains more than 1 percent $H_2O$-, or more than 1 percent $CO_2$, and therefore is possibly altered. These "altered" rocks contain less than 2 percent $H_2O$-, or less than 2 percent $CO_2$-, except for three samples from the Tabar group which have $CO_2$ values of 2.40, 2.45, and 2.65 percent. This division into "fresh" and "altered" rocks is largely arbitrary, and may obscure the fact that some of the strongly alkaline lavas could have been volatile rich, containing relatively high proportions of "primary" $CO_2$ and $H_2O$.

Seventy-three of the 84 chemical analyses have $Fe_2O_3$ values equal to, or greater than, those of FeO. This appears to correspond with the exceptionally low ulvöspinel contents and, consequently, the high $Fe_2O_3$/FeO values of the homogeneous titanomagnetite grains in these rocks (see Appendix). In order that the chemistry of these rocks can be compared on the same basis with the compositions of others in which $Fe_2O_3$/FeO values are much lower, the CIPW norms of the 84 samples have been calculated using the oxidation adjustment recommended by Irvine and Baragar (1971) - i.e., if $Fe_2O_3 > TiO_2 + 1.5$ wt. percent, sufficient $Fe_2O_3$ is converted to FeO so that $Fe_2O_3 = TiO_2 + 1.5$. All the variation diagrams in Figures 2-to-7 show the values of the whole-rock analyses recalculated to 100 percent on a volatile-free basis ($H_2O$+, $H_2O$-, $CO_2$, S, $SO_3$, and Cl, all excluded) after use of the Irvine and Baragar transform.

## PETROCHEMISTRY

### Alkalinity

Most of the analysed rocks from the Tabar-to-Feni Islands are nepheline-normative, as shown in Figure 2. In Figure 3, the broad spread of

Fig 2

points in the Hawaiian alkaline field shows a positive slope, but at the silica-poor end the points define a nearly vertical array that cuts upward through the field boundary from the tholeiitic field.

There is an absence of rocks in the silica range 59-64 percent, but all those rocks showing silica contents greater than 64 percent fall in the Hawaiian tholeiitic field (Fig. 3). As shown in Figure 2, these rocks have high Differentiation Indices (DI), and more than 9 percent normative "quartz" (see Fig. 2 for definition of "quartz"). The extremely wide range in compositions of the Tabar-to-Feni rocks is also illustrated in Figure 2: the rocks contain between 23 percent normative nepheline and 17 percent normative "quartz", and show a range of DI values between 4 and 95.

Fig 3

Potash:soda ratios

Most analysed rocks from the Tabar-to-Feni Islands have $K_2O/Na_2O$ values greater than 0.5, and a few have values greater than 1 (Fig. 4). These values are typical of rocks from a wide range of tectonic settings, including rocks of the "shoshonite association" (Joplin, 1968) which are believed to be characteristic of island-arcs and recently stabilized orogenic areas (cf. Jakeš and White, 1972; Mackenzie and Chappell, 1972; Johnson et al., 1973).

Fig 4

Potash:silica relationships

The ratios of $K_2O$ to $SiO_2$ in the Tabar-to-Feni rocks show a wide range (Fig. 5), and the pattern of variation is similar to that for $Na_2O+K_2O$ versus $SiO_2$ (cf. Fig. 3). Silica-poor rocks, especially, show a wide range of $K_2O$ values, and form a steep overall trend whose slope decreases as silica increases to about 59 percent. Most of the quartz-normative rocks (> 64 percent $SiO_2$) contain only moderate amounts of $K_2O$.

$K_2O:SiO_2$ relationships are used widely for purposes of rock classification, and in Figure 5 the Tabar-to-Feni samples are compared with two taxonomic schemes proposed recently for island-arc rocks. A comparison of these schemes illustrates the confusion that exists in current nomenclature for potassic rocks, particularly in island-arc regions. Some volcanic rocks from oceanic islands and continental areas also plot in the same parts of the $K_2O$:$SiO_2$ diagram as do many Tabar-to-Feni rocks, and this adds further confusion, as many of them have been given names different to those normally used in island-arc areas (e.g. the basalt - trachybasalt - tristanite - trachyte series of Gough, Tristan da Cunha, and Nightingale Islands; Tilley and Muir 1964). While recognising the importance of $K_2O$ contents as possible genetic indicators, the taxonomic scheme adopted in this paper uses $K_2O$ values only as qualifiers to a system of rock names based primarily on normative mineralogy (see below).

Fig 5

Magnesium:iron ratios

In Figures 2-to-5 the Tabar-to-Feni rocks show broad fields of chemical variation, but in the MgO versus FeO + $Fe_2O_3$ diagram in Figure 6 they show a relatively narrow trend. None of the rocks of intermediate MgO content have the high total-iron values of tholeiitic rocks, and the general trend of slight iron-enrichment is that shown by alkaline associations and by many hypersthene-normative island-arc associations (Fig. 5).

Fig 6

Titania and alumina contents

The histogram in Figure 7A shows that most rocks from the Tabar-to-Feni Islands contain less than 1 percent $TiO_2$, and that only one sample contains more than 1.5 percent $TiO_2$. It also shows that most of the Tabar and Lihir rocks contain more than 0.8 percent $TiO_2$, whereas most of those from Tanga and Feni have less than 0.8 percent $TiO_2$.

Fig 7

Rocks containing less than 1.5 percent $TiO_2$ are characteristic of basaltic rocks (DI$<$50, $SiO_2$$<$54 wt. %) from circumoceanic regions (island-arcs, continental margins, etc), whereas those from oceanic islands generally contain more than 2 percent $TiO_2$ (Chayes, 1965). The consistently low $TiO_2$

contents of the Tabar-to-Feni rocks are therefore a persuasive reason for regarding these rocks as "circumoceanic" types.

The $Al_2O_3$ contents of the Tabar-to-Feni rocks have a wide range - from less than 7 percent in pyroxenite inclusions to more than 21 percent in three rocks from the Tanga group (Fig. 7B). In general, however, $Al_2O_3$ values are high - mainly between 16 and 20 percent - and again typical of circumoceanic volcanic rocks (cf. Chayes, 1965).

Geochemistry

Page and Johnson (1974) presented $Sr^{87}/Sr^{86}$, strontium, and rubidium values for 10 Tabar-to-Feni rocks. The $Sr^{87}/Sr^{86}$ values are uniform (0.7040 - 0.7044) and only slightly higher than those of other Late Cainozoic rocks in the Bismarck Archipelago. They are typical of values for many other island-arc rocks, and are too low to suggest that the compositions of the magmas have been changed significantly by contamination from old radiogenic crust (Page and Johnson, 1974). Strontium contents are notably high (1223 - 1667 ppm). Rubidium values range between 17 and 124.

## TAXONOMY AND DISTRIBUTION OF ROCK TYPES

The rock names used in this study are based almost entirely upon whole-rock chemical compositions. The rock names are derived initially from the grid shown in Figure 8, and are assigned the prefixes "potassic" and "sodic" when $K_2O/Na_2O$ values are greater than, or less than, 0.5, respectively. The amount of modal olivine distinguishes basanites (olivine $>1\%$ by volume) from tephrites (olivine $<1\%$). The term "olivine nephelinite" is used here in a purely chemical sense for basaltic rocks rich in normative nepheline and without normative albite (cf. Green and Ringwood, 1968), and its use does not necessarily imply the presence of modal nepheline (cf. samples 1 and 4, Tables 1 and 2). The "transitional" basalts (Coombs, 1963) contain low amounts of normative hypersthene, but show the mineralogy of alkali

Fig 8

basalts (e.g. sample 3, Tables 1 and 2); none of the analysed basaltic rocks are true tholeiites.

Although this taxonomic system provides an internally consistent series of well-known rock names, it is essential to point out that some of these names have synonyms; for example, the "potassic ne-trachyandesites" are roughly equivalent to the "tristanites" of Tilley and Muir (1964). In particular, it should be noted that the "potassic transitional basalts", "potassic alkali basalts", and "potassic trachybasalts", are the same as many rocks grouped by some petrologists in the "shoshonite association" (i.e. the absarokites, shoshonites, banakites, etc.; Joplin, 1968). However, unlike the rocks of the shoshonite association, many Tabar-to-Feni rocks are strongly silica-undersaturated, and they cannot be assigned familiar names which relate them to the shoshonite association. The decision not to use rock names of the shoshonite association has also been influenced by recent accounts which ignore, side-step, or recommend abandoning these names (Streckeison, 1967; Coombs and Wilkinson, 1969; Nicholls and Carmichael, 1969; Irvine and Baragar, 1971).

The following generalisations on rock type distribution can be made from a comparison of Figures 2, 4, and 8.

1. "Potassic" rocks appear to be more common than "sodic" ones in all the island groups (Fig. 4).
2. The most common analysed rocks from the Tabar and Lihir Islands are low in normative "quartz", or low in normative nepheline. Potassic trachybasalts appear to be common on both island groups, and potassic alkali basalts seem to be common on Lihir Island.
3. Rocks showing high normative nepheline values appear to be the most common types on the Tanga and Feni Islands. Potassic phonolitic tephrites seem to be especially abundant.

4. Undersaturated salic rocks (DI > 75) have been found only on Malendok Island, in the Tanga group (ne trachytes and one phonolitic trachyte).

5. Q-trachytes are present on Tabar Island, on Bitlik and Bitbok Islands (Tanga group), and on Ambitle and Babase Islands (Feni group).

PETROGRAPHY AND MINERALOGY

Table 1

The Tabar-to-Feni rocks show an exceptionally wide range of petrographic types. In this short report it is not possible to give a comprehensive account of this mineralogical diversity, but by way of illustration the petrography and mineralogy of 11 rocks are summarised in Table 1. As shown in Table 2, these 11 rocks cover a wide range of the chemical spectrum. There is, however, a bias towards the rocks of Ambitle Island where strongly undersaturated types rich in feldspathoid and analcite phenocrysts are especially well developed. Additional mineralogical notes are provided in the Appendix.

Table 2

MODELS FOR PETROGENESIS

There appears to be no simple explanation for the wide range of rock compositions represented on the Tabar-to-Feni Islands. The broad compositional spectrum of the basic rocks suggests that a wide range of primary magmas may have been produced by partial melting in the upper mantle, and the even wider range of intermediate and salic rocks suggests that, if these are differentiates of the basic rocks, many different liquid lines of descent were followed. The Q-trachytes appear to form a distinct chemical group; they could have had an origin not directly connected with the development of the more basic rocks, and of all the rock types seem to be the most likely candidates for possible derivation by crustal anatexis.

It is difficult to establish a feasible model to explain the petrogenesis of primary magmas beneath the Tabar-to-Feni Islands. This difficulty results largely from the uncertain nature of the tectonic setting that gave rise to the volcanism, and to the absence of relevant experimental data on rocks of these compositions. There are, however, no compelling geological or petrological reasons to suggest that the magmas are associated with a sea-floor spreading regime, or related to relative movements of lithospheric slabs over asthenospheric "plumes" (Morgan, 1972). At the present time, therefore, it is considered that the most feasible models are those which recognise the Tabar-to-Feni magmas as broadly "circumoceanic" in type - i.e., they have developed in regions where plate convergence and subduction have taken place, and where partial melting in the upper mantle may have taken place under hydrous conditions.

Curtis (1973) suggested that in late Eocene to Oligocene times a submarine trench (a subduction zone) existed on the west side of a New Ireland island-arc (see Hohnen, in press), and that left-lateral transcurrent movement along the trench in the late- and post-Miocene translated New Ireland to its present position northeast of New Britain. This mechanism would destroy the trench, and disconnect the downgoing part of the lithosphere beneath New Ireland. It is possible, therefore, that this disconnected slab could act as a potential source of water that led to the generation of primary magmas in the overlying mantle during the descent of the slab into the asthenosphere.

The narrowness of the Tabar-to-Feni chain, and the $K_2O:SiO_2$ relationships shown in Figure 5, could be explained by proposing a near-vertical subducted slab of lithosphere beneath the volcanic chain. The $K_2O$ contents of island-arc rocks are believed to be a function of depth to the underlying Benioff zone (e.g. Dickinson, 1970; Nielson and Stoiber, 1973; Whitford and Nicholls, this volume), and the wide range of $K_2O$ values shown in Figure 5 might be due to primary magmas of different compositions (1) originating from a wide range of depths immediately above a near-vertical slab, (2) rising more or less vertically through common channels, (3) fractionating, and (4) erupting from the same volcanoes in a narrow zone above the slab (note how in Fig. 5 the trend of points transects the "h" lines of Ninkovich and Hays, 1972).

If the zone of deep water north and east of the Tabar-to-Feni Islands is the remnant of a submarine trench, a possible alternative subduction model is that a plate was consumed from the northeast, and that the downgoing slab was not disconnected by faulting. However, this model becomes unnecessarily complicated when it is developed to explain the Tertiary evolution of the New Ireland island-arc at a site further from the zone of deep water than the volcanic chain. In addition, the seismic reflection profile presented by Furumoto et al. (in prep.) does not provide evidence that the zone of deep water is necessarily a submarine trench.

There is, in any case, no evidence for the existence of a downgoing slab - disconnected or otherwise - beneath the Tabar-to-Feni Islands. Intermediate and deep focus earthquakes have not been reported from beneath the volcanic chain; the raised coral reefs of the islands indicate uplift, rather than sinking of sea-floor over a downgoing slab of lithosphere; and no tholeiitic basalts, andesites, and dacites (the rocks most characteristic of circum-Pacific island-arc volcanoes) appear to be present on the islands.

The possibility of another model is being explored - that the volcanic chain overlies a zone of faulting which marks the boundary of a region that has undergone isostatic readjustments. Talwani and Eldholm (1973) identified major faults at the boundaries of continental and oceanic crust in various parts of the World, attributing them to isostasy. Analagous faults may exist beneath the Tabar-to-Feni Islands between the Pacific oceanic crust in the northeast and the thicker crust of the Bismarck Sea/New Ireland region in the southwest. Evidence of these faults is provided by the seismic reflection profile given by Furumoto et al. (in prep.), and evidence for isostatic instability of the islands is provided by the raised reefs, and by Bouguer gravity anomalies (+160 and +180 mgal) which, according to Finlayson and Cull (1973), are probably due "to the crust's supporting accumulated volcanic piles without isostatic compensation."

In this isostatic model it is proposed that, during the early Tertiary, northeastward subduction beneath the New Ireland island-arc introduced water and partial melts of the downgoing slab to the mantle peridotite above the slab, producing hydrous minerals such as amphibole (e.g. Green, 1972), and probably an uneven spatial distribution of incompatible elements such as potassium. During the late Tertiary and Quaternary, an isostatically controlled, Tabar-to-Feni faulted zone extended to depths where partial melting could take place under hydrous conditions by releases of total-pressure caused by the faulting. This hydrous melting produced a wide variety of basic primary magmas which had the low titania contents of other circumoceanic basalts, and which fractionated to give rocks containing hydrous minerals (zeolites, amphibole, and biotite).

This model is unorthodox, but is perhaps justified in that the compositions of the Tabar-to-Feni rocks are unusual in the context of Late Cainozoic volcanism in the southwest Pacific region.

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## Appendix. Mineralogical notes

### Haüyne

Phenocrysts of haüyne in samples 7, 8, 9, and 10 (Table 1) have been partially analysed with the electron microprobe (S and Cl are undetermined). As shown in Figure A, there are pronounced chemical variations between, and within, the phenocrysts, and the most common type of substitution appears to be a replacement of K by Na and Ca. In the trend of Ca substitution, electrostatic neutrality is maintained by a decrease in the ideal 1:1 Al-Si substitution, or else the Al-Si substitution is unaffected, and neutrality is apparently restored by $SO_4$-Cl(or S) paired substitutions. (Many of the haüyne crystals contain zones of minute inclusions of pyrite which are, perhaps, evidence for S-Cl zoning).

Analyses of the cores and rims of four haüyne phenocrysts in three different samples indicate two different patterns of zoning. One crystal in sample 8 and another in sample 10 show an increase in K and a decrease in Na and Ca from core to rim. A second crystal in sample 10, however, and one in sample 7, show a different pattern: from core to rim, K decreases, Na increases, and Ca shows only a slight decrease (i.e., more or less a simple K-Na substitution). These preliminary results indicate complexities in the types of possible substitutions in haüyne, and the need for more detailed studies with the microprobe.

Analcite

Twelve phenocrysts and groundmass grains of analcite in five rocks have been analysed with the microprobe. They show a restricted and erratic compositional variation, even though they are present in a wide variety of rock types (e.g. samples 2, 5, 6, and 10). The analcite phenocrysts in any one rock do not necessarily show identical compositions (e.g. sample 6).

The most obvious chemical variation is a slight Al-Si substitution ($SiO_2$ = 50-57%, $Al_2O_3$ = 27-23%). Wilkinson (1965) noted that the principal replacement of Si by Na and Al in analcite from the Square Top intrusion (New South Wales) decreases as temperature of crystallisation decreases. In contrast, the compositions of the analcite in the Tabar-to-Feni rocks do not appear to vary significantly from one rock to another, and therefore appear to be independent of temperature.

Iron-titanium oxide

Optically homogeneous titanomagnetite is the only iron-titanium oxide found in the eleven rocks examined with the microprobe. The ulvöspinel contents (calculated by the method of Carmichael, 1967) of these titanomagnetite crystals are much lower than in those of other basic igneous rocks: in some specimens (e.g. no. 4) there is insufficient Ti to even satisfy the Mn, Mg, and Ca components of the ulvöspinel molecule ($2R''O.TiO_2$), so that a small amount of Mg-Al spinel is also present in solid solution.

The analysed Tabar-to-Feni rocks have high $Fe_2O_3$, and low $TiO_2$, contents. As these are also the characteristics of the analysed, optically homogeneous titanomagnetite crystals, it seems likely that the apparently high oxidation state of the rocks is not due to secondary oxidation processes, but rather is a reflection of a high oxidation state in the magmas before their eruption.

Clinopyroxene

The analysed clinopyroxene phenocrysts of all the rocks are pale yellow-green salite. Many of them show weak pleochroism, and oscillatory zoning is common. The margins are more Fe-rich than the cores, but crystals in sample 4 have margins which are lower in Fe, Ti, and Al, and higher in Si and Mg, compared to the cores, because of the presence of titanomagnetite grains in the margins.

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Table 1. Summary of petrography and mineralogy of eleven rocks from Tabar-to-Feni Islands

| Sample [a] | Rock type [b] | Phenocrysts [c] | Groundmass [c] |
|---|---|---|---|
| 1. | Potassic olivine nephelinite | 31% Cpx, < 1% Ol ($Fo_{86}$), < 1% Tm (extremely rare) | Hypocrystalline. Rounded patches of alkali-bearing chabazite (6% of total rock, up to ca. 0.4 mm across, apparently of primary origin, identification confirmed by X-ray diffraction), in matrix of Cpx, Tm, & brown glass. |
| 2. | Potassic basanite | 24% Cpx, 8% Ol ($Fo_{90}$), 2% Tm, < 1% Pl, < 1% pseudomorphs, probably after Amph. | Hypocrystalline. Pl ($An_{58-69}$), Cpx, Tm, analcite (patches), Bl, Ap, & turbid brown glass. |
| 3. | Potassic transitional basalt | 26% Cpx, 19% Pl (zoned, $An_{66-81}$ in one grain), 3% Ol ($Fo_{86}$, partly pseudomorphed by "iddingsite"), 2% Tm. | Holocrystalline. Pl ($An_{62}$, $An_{65}$), Cpx, Tm, Ap, some alteration to secondary minerals. |
| 4. | Potassic olivine nephelinite | 31% Cpx, 1% Ol ($Fo_{84}$), 1% Tm. | Holocrystalline. < 1% leucite microphenocrysts in matrix of Cpx, Tm, Pl ($An_{41}$, $An_{64}$, interstitial), & leucite. |
| 5. | Sodic tephrite | 33% Cpx, 4% Tm, 3% Pl ($An_{76}$, $An_{82}$), 2% analcite (n = 1.492), < 1% Ol. | Holocrystalline. Pl, Cpx, Tm, & analcite. |
| 6. | Potassic phonolitic tephrite | 17% Cpx, 3% Pl ($An_{82}$, $An_{87}$), 3% Tm, (these 3 phenocryst minerals also in aggregates), 2% analcite (n = 1.489), < 1% Amph (basaltic hornblende), < 1% Ap, < 1% Bl. | Holocrystalline. Cpx, Pl ($An_{85}$), Bl, Tm, Hb, analcite, minor calcite, and zeolites (possible chabazite and mesolite). |
| 7. | Potassic phonolitic tephrite | 30% Pl (strong oscillatory zoning, cores of $An_{75}$ to margins of $An_{42}$) rimmed by Kf ($Or_{51}$ $Ab_{46}$ $An_3$), 14% Cpx, 4% Tm, 1% haüyne (brownish or bluish due to alteration of pyrite inclusions, some with colourless margins), < 1% Ap, < 1% pseudomorphs, probably after Amph. | Holocrystalline. Pl ($An_{31}$), A'clase ($Or_{20}$, $Ab_{75}$, $An_5$), Tm, Cpx, rare sodalite, chalcopyrite, & secondary calcite. |
| 8. | Potassic tephritic phonolite | 11% Pl ($An_{36-40}$), 6% Cpx, 4% haüyne (similar in appearance to those in sample 7), 2% Tm, 2% pseudomorphs, probably after Amph, < 1% Ap, < 1% sodalite. | Holocrystalline. A'clase microphenocrysts in matrix of A'clase ($Or_{30}$ $Ab_{66}$ $An_4$, $Or_{23}$ $Ab_{71}$ $An_6$), Tm, Cpx, & possible analcite. |
| 9. | Sodic ne-trachyandesite | 19% Pl ($Or_4$ $Ab_{54}$ $An_{42}$ to $Or_7$ $Ab_{69}$ $An_{24}$), 3% Cpx, 2% haüyne (n = 1.5045, colourless where fresh - especially at margins, greyish or brownish where altered pyrite inclusions present), 1% Tm, 1% pseudomorphs, probably after Amph, < 1% Ap (dusty brown-grey, containing fine-grained inclusions). | Holocrystalline. Ab, A'clase ($Or_{23}$ $Ab_{73}$ $An_4$ to $Or_{12}$ $Ab_{79}$ $An_9$), Cpx, Tm, Ap, & Bl. |
| 10. | Potassic ne-trachyte | 5% Pl ($An_{58}$), 3% haüyne (mainly colourless cores, bluish margins, & turbid brownish rims; some mainly brown, where altered pyrite inclusions present; rare crystals have rims of Kf - $Or_{85}$ $Ab_{14}$ $An_1$), 2% Cpx, < 1% Tm, < 1% Bl. | Holocrystalline. Pl microphenocrysts in matrix of Pl ($An_{36}$), Tm, Cpx, Kf & A'clase ($Or_{55}$ $Ab_{44}$ $An_1$ to $Or_{29}$ $Ab_{70}$ $An_1$), analcite. |
| 11. | Potassic Q-trachyte | 57% alkali feldspar (one analysed grain = $Or_7$ $Ab_{89}$ $An_4$; another = $Or_{39}$ $Ab_{61}$), < 1% Amph (tremolite), < 1% Bl (largely pseudomorphed by Tm), < 1% Tm. | Holocrystalline. Quartz, alkali feldspar, Amph (pale blue riebeckite), Tm, & sericite. |

a. Locality descriptions (numbers in parentheses are BMR registered sample numbers)

1. *In situ* outcrop, 1.5 km west of Nanum River outlet, southwest coast of Ambitle Island, Feni group (73680007).
2. Boulder in stream, 2 km southeast of Napekur village, Simberi Island, Tabar group (73680045).
3. *In situ* lava flow, stream 1.5 km south of Datawa village, Tabar Island, Tabar group (73680034).
4. *In situ* outcrop, Waramung Bay, west coast of Ambitle Island, Feni group (74400047).
5. Boulder near head of stream draining southwest part of caldera at Luise Harbour, Lihir Island, Lihir group (74400042).
6. *In situ* outcrop, 1.5 km up Niffin River, southwest side of Ambitle Island, Feni group (73680010).
7. Boulder on track south of Balum Plantation, Babase Island, Feni group (73680005).
8. *In situ* outcrop, 0.8 km south of Nanum River outlet, south coast of Ambitle Island, Feni group (74400048).
9. Boulder by roadside, 1.2 km south of Nanum River outlet, south coast of Ambitle Island, Feni group (73680011).
10. *In situ* outcrop, summit of peak above Put Plantation, southwest coast of Malendok Island, Tanga group (73680013).
11. *In situ* outcrop, near head of stream 1 km southeast of Tiripats Plantation, Tabar Island, Tabar group (73680037).

b. Samples listed in order of increasing Differentiation Indices.

c. Abbreviations for common rock-forming minerals: Ol = olivine, Fo = fosterite, Cpx = clinopyroxene, Amph = amphibole, Bl = biotite, Tm = titanomagnetite, Pl = plagioclase, An = anorthite, Ab = albite, Or = orthoclase, A'clase = anorthoclase, Kf = potash feldspar, Ap = apatite.

Table 2. Chemical analyses[a] and CIPW norms of eleven rocks listed in Table 1.

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 |
|---|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 45.8 | 46.6 | 49.0 | 45.9 | 46.6 | 46.0 | 50.3 | 54.5 | 55.8 | 54.9 | 68.5 |
| $TiO_2$ | 0.67 | 0.75 | 1.18 | 0.80 | 0.95 | 0.70 | 0.75 | 0.80 | 0.68 | 0.39 | 0.30 |
| $Al_2O_3$ | 10.2 | 12.9 | 13.8 | 12.1 | 15.4 | 17.4 | 18.7 | 17.90 | 17.7 | 21.1 | 16.3 |
| $Fe_2O_3$ | 5.60 | 3.95 | 4.50 | 5.80 | 6.30 | 5.65 | 4.95 | 4.35 | 3.90 | 2.40 | 1.46 |
| FeO | 4.90 | 5.10 | 6.15 | 5.00 | 5.40 | 4.15 | 3.20 | 2.50 | 2.20 | 1.71 | 0.46 |
| MnO | 0.20 | 0.18 | 0.20 | 0.20 | 0.22 | 0.24 | 0.26 | 0.13 | 0.12 | 0.18 | 0.04 |
| MgO | 9.05 | 11.0 | 7.50 | 8.30 | 5.80 | 4.40 | 2.85 | 3.05 | 2.70 | 1.01 | 0.42 |
| CaO | 14.7 | 12.3 | 11.80 | 14.4 | 11.80 | 11.1 | 9.55 | 5.75 | 5.30 | 3.15 | 0.55 |
| $Na_2O$ | 2.85 | 2.70 | 2.10 | 3.65 | 3.60 | 3.50 | 4.60 | 6.55 | 6.85 | 4.85 | 5.95 |
| $K_2O$ | 2.15 | 1.60 | 1.70 | 2.15 | 1.75 | 3.35 | 2.75 | 3.40 | 3.35 | 6.65 | 4.85 |
| $P_2O_5$ | 0.58 | 0.38 | 0.29 | 0.53 | 0.60 | 0.64 | 0.48 | 0.43 | 0.40 | 0.19 | 0.01 |
| $H_2O+$ | 1.54 | 1.79 | 1.08 | 1.29 | 1.30 | 2.25 | 0.67 | 0.25 | 0.15 | 1.50 | 0.29 |
| $H_2O-$ | 1.70 | 0.35 | 0.80 | 0.03 | 0.36 | 0.59 | 0.67 | 0.09 | 0.13 | 1.44 | 0.33 |
| $CO_2$ | 0.05 | 0.25 | 0.10 | 0.05 | 0.05 | 0.15 | 0.20 | < 0.05 | 0.05 | 0.05 | 0.10 |
| Cl | 0.06 | < 0.01 | < 0.01 | n.d. | n.d. | 0.03 | 0.18 | n.d. | 0.36 | 0.21 | 0.01 |
| $SO_3$ | 0.01[b] | 0.005 | 0.04[b] | 0.03[b] | 0.04[b] | 0.005 | 0.04[b] | 0.07[b] | 0.11 | 0.04[b] | 0.04[b] |
| S | n.d.[c] | 0.023 | n.d. | n.d. | n.d. | 0.028 | n.d. | n.d. | - | n.d. | n.d. |
| sub-total | 100.06 | 99.88 | 100.24 | 100.23 | 100.17 | 100.18 | 100.15 | 99.77 | 99.80 | 99.77 | 99.61 |
| less O = Cl | 0.01 | - | - | - | - | 0.01 | 0.04 | - | 0.08 | 0.05 | - |
| Total | 100.05 | 99.88 | 100.24 | 100.23 | 100.17 | 100.17 | 100.11 | 99.77 | 99.72 | 99.72 | 99.61 |
| D.I. | 25.92 | 26.52 | 28.36 | 28.39 | 33.19 | 38.90 | 48.49 | 66.71 | 70.25 | 76.83 | 93.64 |

CIPW norms[d]

| | | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Q | - | - | - | - | - | - | - | - | - | - | 13.74 |
| | C | - | - | - | - | - | - | - | - | - | 0.68 | 0.29 |
| | or | 9.43 | 9.72 | 10.24 | 6.01 | 10.55 | 20.45 | 16.56 | 20.26 | 20.03 | 40.72 | 28.99 |
| | ab | - | 8.91 | 18.12 | - | 12.70 | 4.11 | 22.80 | 35.31 | 40.30 | 28.53 | 50.91 |
| | an | 9.02 | 18.87 | 23.67 | 10.45 | 21.11 | 22.60 | 22.69 | 9.49 | 7.76 | 14.91 | 2.69 |
| | lc | 2.94 | - | - | 5.40 | - | - | - | - | - | - | - |
| | ne | 13.55 | 7.89 | - | 16.98 | 9.95 | 14.34 | 9.13 | 11.14 | 9.93 | 7.58 | - |
| di | wo | 26.19 | 17.24 | 14.23 | 24.45 | 14.44 | 12.51 | 9.35 | 6.87 | 6.76 | - | - |
| | en | 16.05 | 11.81 | 8.61 | 14.61 | 7.38 | 6.18 | 4.42 | 3.95 | 3.97 | - | - |
| | fs | 8.65 | 4.06 | 4.85 | 8.57 | 6.70 | 6.09 | 4.81 | 2.60 | 2.47 | - | - |
| hy | en | - | - | 3.48 | - | - | - | - | - | - | - | 1.06 |
| | fs | - | - | 1.96 | - | - | - | - | - | - | - | - |
| ol | fo | 5.14 | 11.45 | 4.87 | 4.46 | 5.15 | 3.60 | 1.97 | 2.60 | 1.99 | 1.83 | - |
| | fa | 3.05 | 4.34 | 3.02 | 2.89 | 5.15 | 3.91 | 2.37 | 1.89 | 1.36 | 1.69 | - |
| | mt | 3.27 | 3.35 | 3.96 | 3.39 | 3.62 | 3.30 | 3.32 | 3.36 | 3.20 | 2.84 | 0.75 |
| | hm | - | - | - | - | - | - | - | - | - | - | 0.96 |
| | il | 1.32 | 1.46 | 2.29 | 1.54 | 1.84 | 1.37 | 1.45 | 1.53 | 1.31 | 0.77 | 0.58 |
| | ap | 1.43 | 0.93 | 0.70 | 1.28 | 1.45 | 1.57 | 1.16 | 1.03 | 0.96 | 0.47 | 0.02 |

a. Supplied by Australian Mineral Development Laboratories, Adelaide.

b. $SO_3$ value is a total sulphur ($SO_3$ + S as $SO_3$) determination.

c. n.d. = not determined.

## FIGURE CAPTIONS

Figure 1. Locality maps.

Figure 2. Differentiation Indices (Thornton and Tuttle, 1960; DI = $\Sigma$ Q + ab + or + ne + lc) plotted against wt. percent normative nepheline (Ne), or normative "quartz" ("Q"), where "Q" is normative quartz plus the silica of normative hypersthene. The values of Ne and "Q" are those obtained after using the Irvine and Baragar (1971) transform (see text). When the $Fe_2O_3/FeO$ values of the original chemical analyses are used in the norm calculation, the points are displaced towards the left, and some nepheline normative rocks become "quartz" normative. Vertical dash on symbol indicates "altered" rocks (see text).

Figure 3. $Na_2O + K_2O$ versus $SiO_2$ diagram. Straight line separates Hawaiian "alkaline" rocks (upper field) from "tholeiitic" ones (MacDonald, 1968), and has been extrapolated to high silica values. The line is not a perfect field boundary for the Tabar-to-Feni rocks, as some nepheline-normative (alkaline) rocks fall below the line, and some hypersthene-normative (tholeiitic) ones fall above it. Vertical dash on symbol indicates "altered" rocks (see text).

Figure 4. $K_2O$ versus $Na_2O$ diagram. Dashed lines indicates $K_2O/Na_2O = 1$. Solid line indicates $K_2O/Na_2O = 0.5$. Vertical dash on symbol indicates "altered" rocks (see text). Note that more than half of the rocks in which $K_2O/Na_2O > 1$ are "altered" (see text).

Figure 5. $K_2O$ versus $SiO_2$ diagram. Two dashed lines are the field boundaries used by Mackenzie and Chappell (1972) to define a lower field of "calcalkaline" rocks, a middle one of "high-K calcalkaline" rocks, and an upper one of "shoshonitic" rocks; rocks containing more than about 3 percent $K_2O$ fall outside the scope of this classification. The three solid lines are schematic boundaries which correspond to depths (h) to Benioff zones beneath volcanoes in Alaska, the Izu-Marianas arc, Chile, and Indonesia, and which were proposed by Ninkovich and Hays (1972) to define a "calcalkaline suite" (below h = 150 km), an "Atlantic alkaline suite" (between h = 150 km and h $\gg$ 300 km), and a "Mediterranean alkaline suite" (above h $\gg$ 300 km). Vertical dash on symbol indicates "altered" rocks (see text).

Figure 6. MgO versus $FeO+Fe_2O_3$ diagram, where FeO and $Fe_2O_3$ values are those obtained after use of the Irvine and Baragar (1971) transform (see text). Solid lines for Gough and Hawaii are drawn through points representing the "average" rock compositions given by LeMaitre (1962) and MacDonald (1968), respectively. The "Talasea" line is defined by 11 rocks analysed by Lowder and Carmichael (1970) from the northern part of Willaumez Peninsula, New Britain,

and illustrates a trend of "mild" iron-enrichment for a hypersthene-normative island-arc association (basalts, andesites, dacites, and rhyolites). Note that except for the two rocks from the Feni group marked with a star, the Tabar-to-Feni rocks constitute a relatively narrow band of chemical variation. Vertical dash on symbol indicates "altered" rocks (see text).

Figure 7. Histograms for (A) $TiO_2$ and (B) $Al_2O_3$ contents of Tabar-to-Feni rocks.

Figure 8. Classification of Tabar-to-Feni rocks based on CIPW norm calculations and the Differentiation Index of Thornton and Tuttle (1960). Solid lines define field boundaries between rocks represented on the Tabar-to-Feni Islands. Dashed lines are approximate outer boundaries of compositional fields. The horizontal lines correspond to the Differentiation Indices used by Baker et al. (1964) for the rocks of Atlantic oceanic islands. X-Y is the arbitrary line used by Coombs and Wilkinson (1969) to separate mildly undersaturated rock associations from strongly undersaturated ones. Q-, hy-, and ne-, indicate that quartz (sensu stricto), hypersthene, and nepheline, respectively, appear in the CIPW norm.

Appendix

Figure A. Weight percent $K_2O$-$Na_2O$-$CaO$ variations in haüyne phenocrysts. Tie lines connect core-rim pairs (arrow heads point to rims). 7, 8, 9, and 10 are rock sample numbers (cf. Table 1), and A, B, and C are different phenocrysts in the same rock.

Figure 1. Locality maps

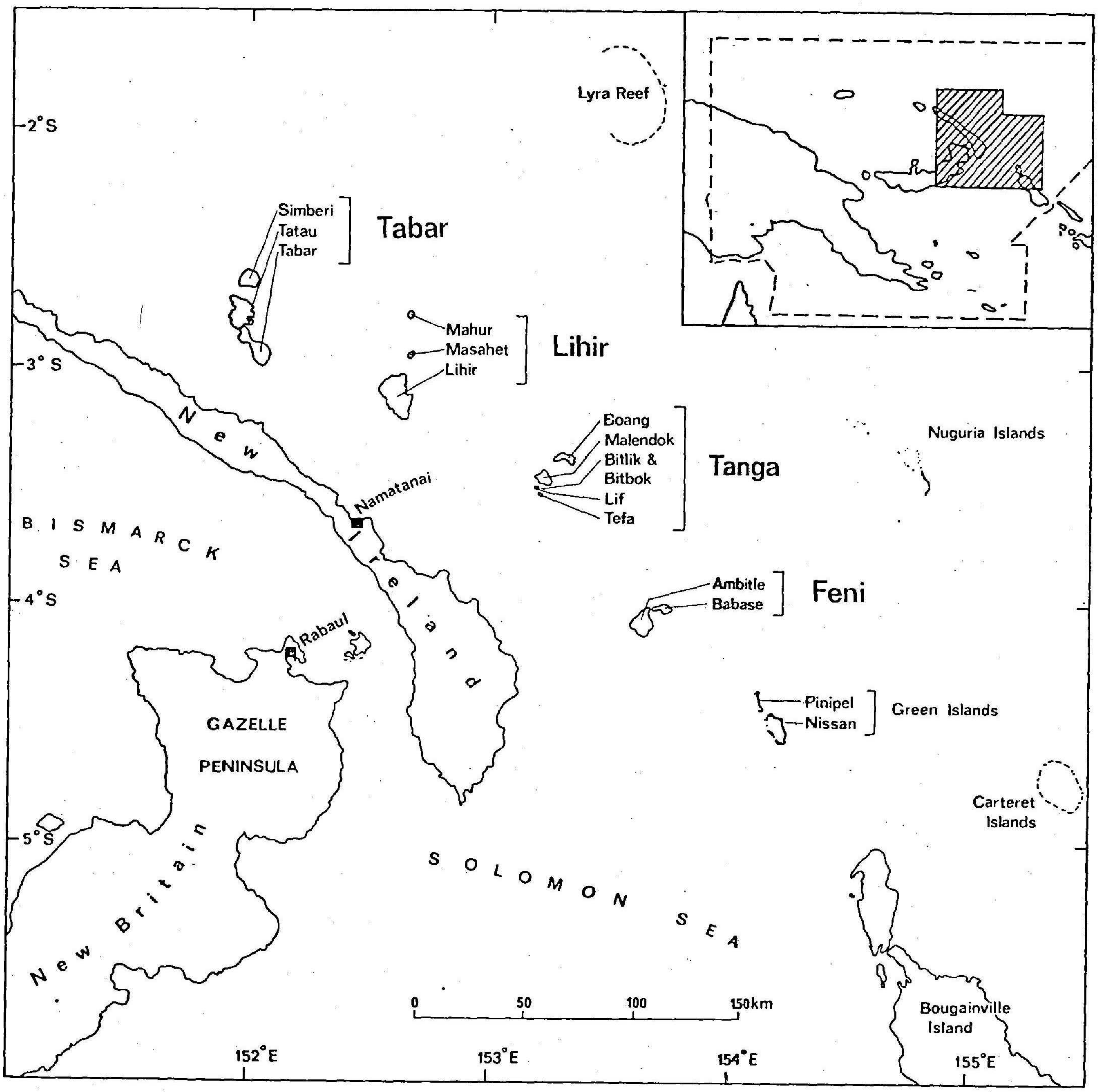


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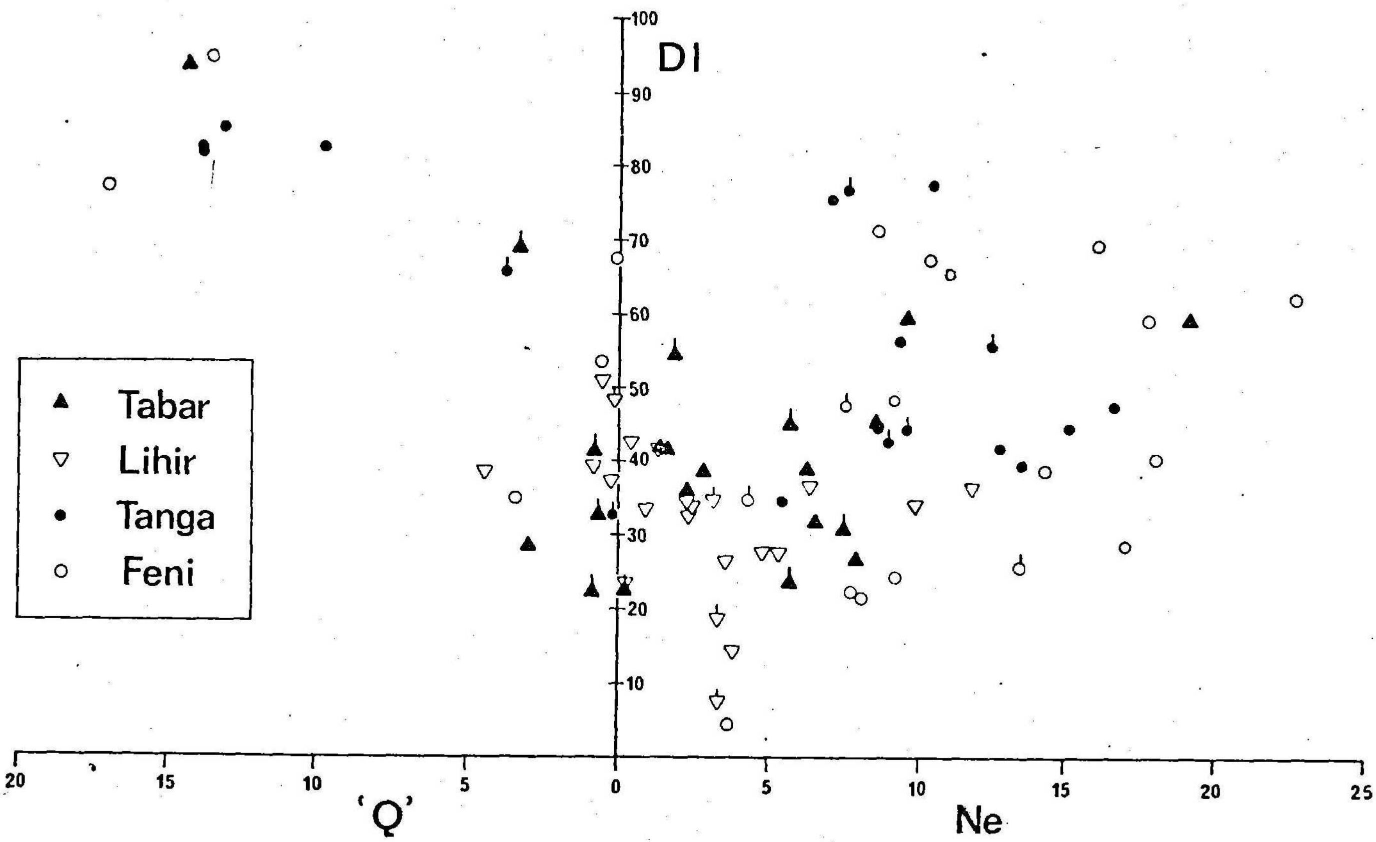
Figure 2. Differentiation Indices (Thornton and Tuttle, 1960; DI = $\Sigma$ Q + ab + or + ne + lc) plotted against wt. percent normative nepheline (Ne), or normative "quartz" ("Q"), where "Q" is normative quartz plus the silica of normative hypersthene. The values of Ne and "Q" are those obtained after using the Irvine and Baragar (1971) transform (see text). When the $Fe_2O_3/FeO$ values of the original chemical analyses are used in the norm calculation, the points are displaced towards the left, and some nepheline normative rocks become "quartz" normative. Vertical dash on symbol indicates "altered" rocks (see text).



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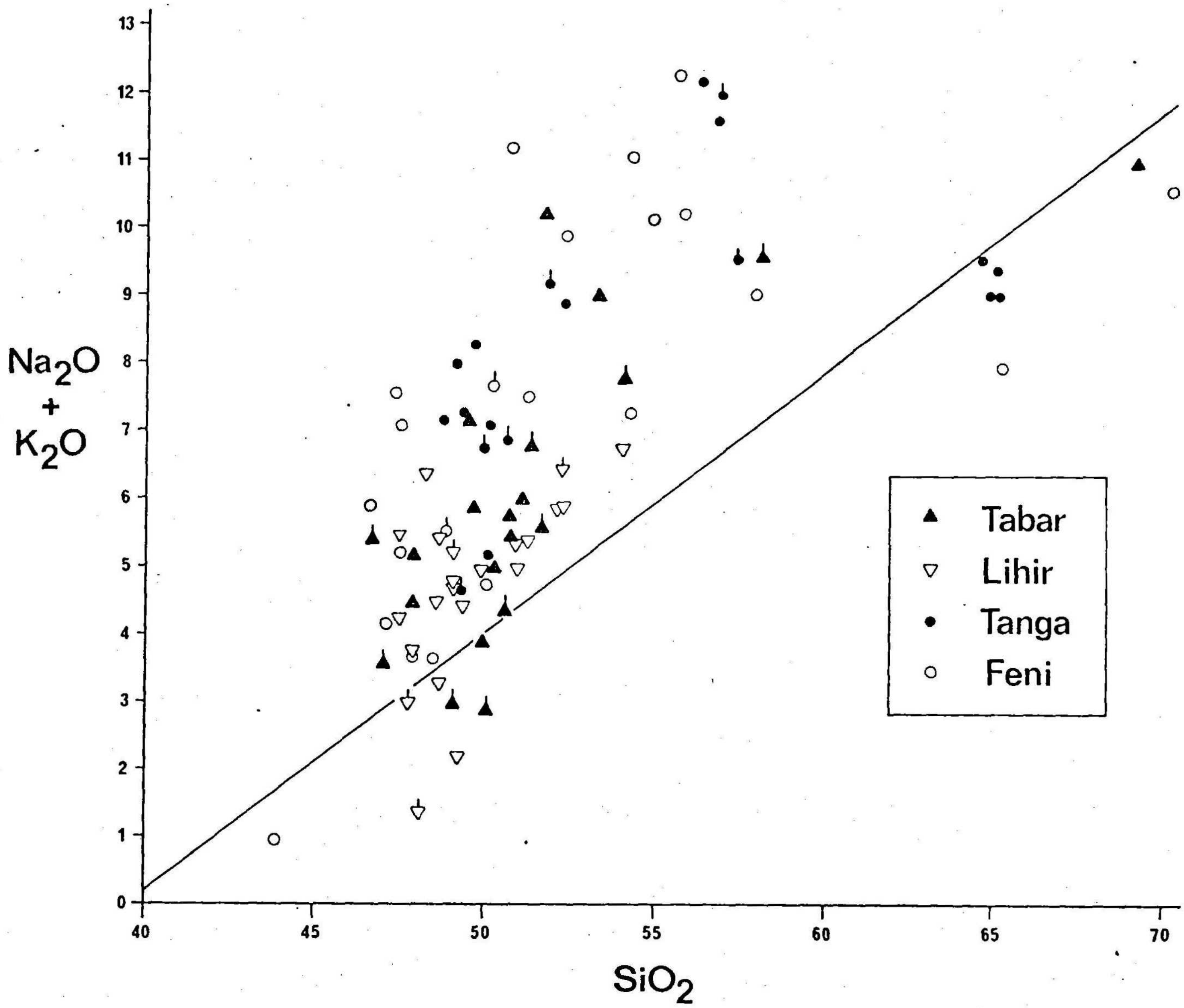

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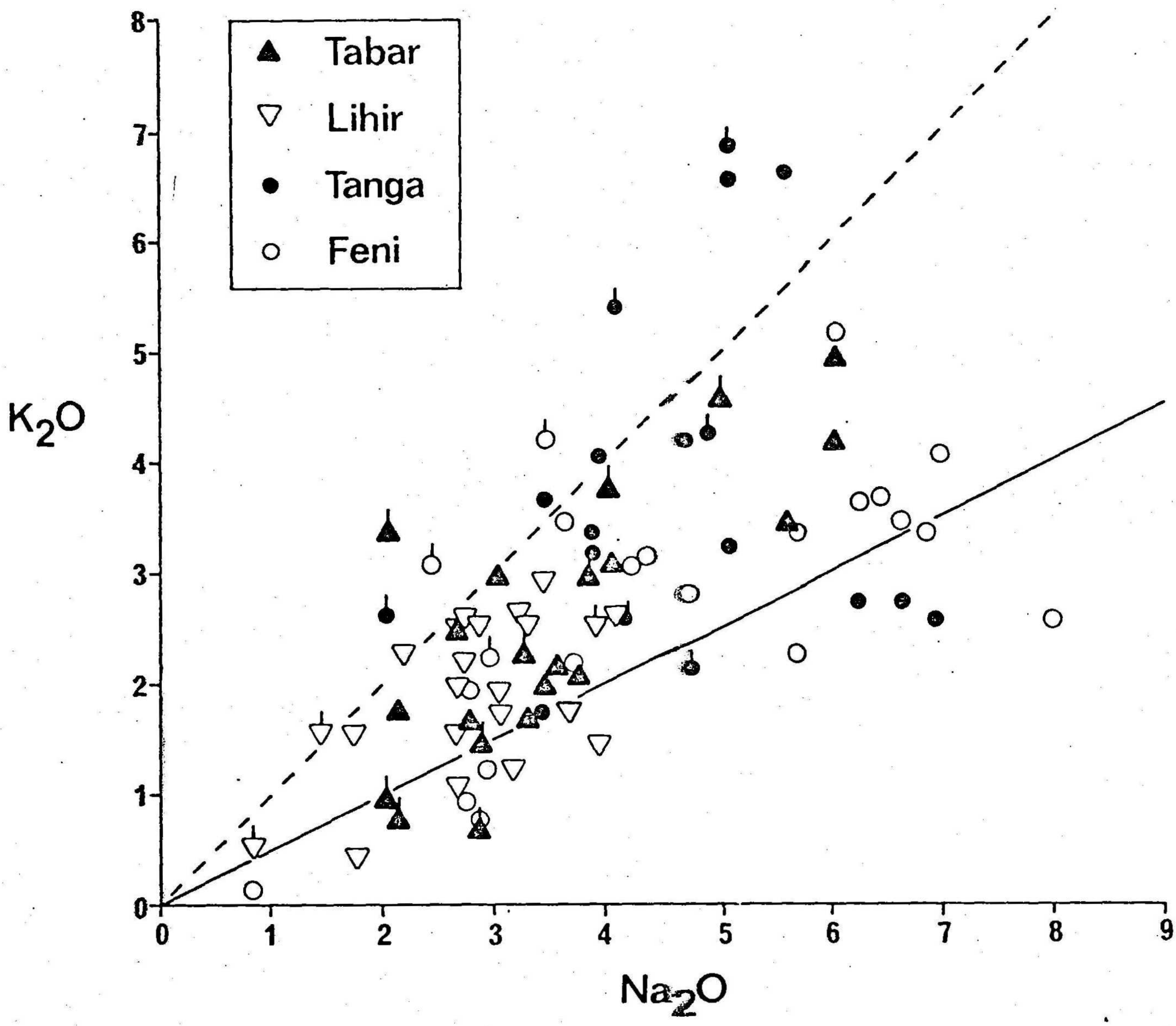


Figure 5. $K_2O$ _versus_ $SiO_2$ diagram. Two dashed lines are the field boundaries used by Mackenzie and Chappell (1972) to define a lower field of "calcalkaline" rocks, a middle one of "high-K calcalkaline" rocks, and an upper one of "shoshonitic" rocks; rocks containing more than about 3 percent $K_2O$ fall outside the scope of this classification. The three solid lines are schematic boundaries which correspond to depths (h) to Benioff zones beneath volcanoes in Alaska, the Izu-Marianas arc, Chile, and Indonesia, and which were proposed by Ninkovich and Hays (1972) to define a "calcalkaline suite" (below h = 150 km), an "Atlantic alkaline suite" (between h = 150 km and h ≥ 300 km), and a "Mediterranean alkaline suite" (above h ≥ 300 km). Vertical dash on symbol indicates "altered" rocks (see text).

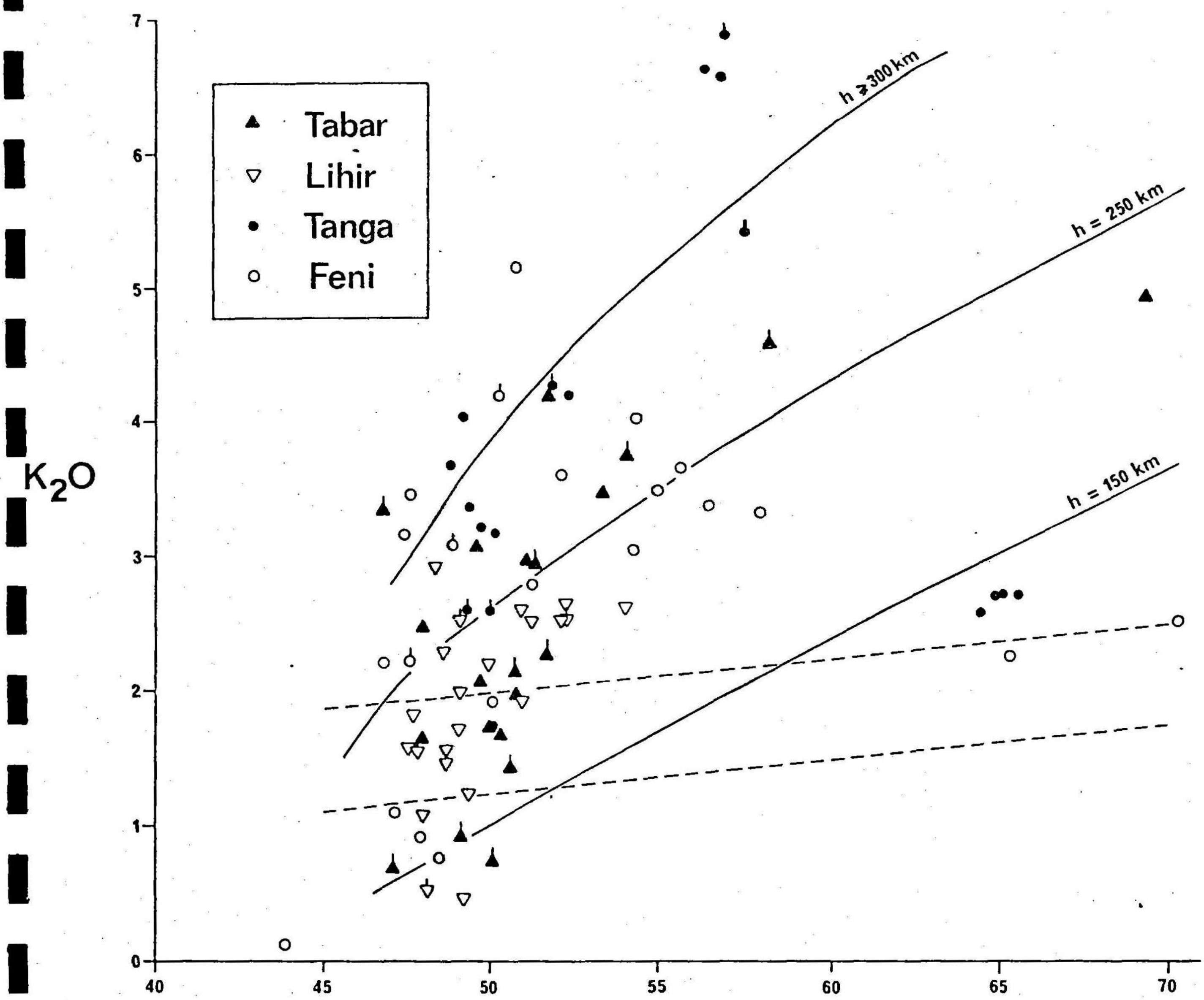


Figure 6. MgO versus $FeO+Fe_2O_3$ diagram, where FeO and $Fe_2O_3$ values are those obtained after use of the Irvine and Baragar (1971) transform (see text). Solid lines for Gough and Hawaii are drawn through points representing the "average" rock compositions given by LeMaitre (1962) and MacDonald (1968), respectively. The "Talasea" line is defined by eleven rocks analysed by Lowder and Carmichael (1970) from the northern part of Willaumez Peninsula, New Britain, and illustrates a trend of "mild" iron-enrichment for a hypersthene-normative island-arc association (basalts, andesites, dacites, and rhyolites). Note that except for the two rocks from the Feni group marked with a star, the Tabar-to-Feni rocks constitute a relatively narrow band of chemical variation. Vertical dash on symbol indicates "altered" rocks (see text).

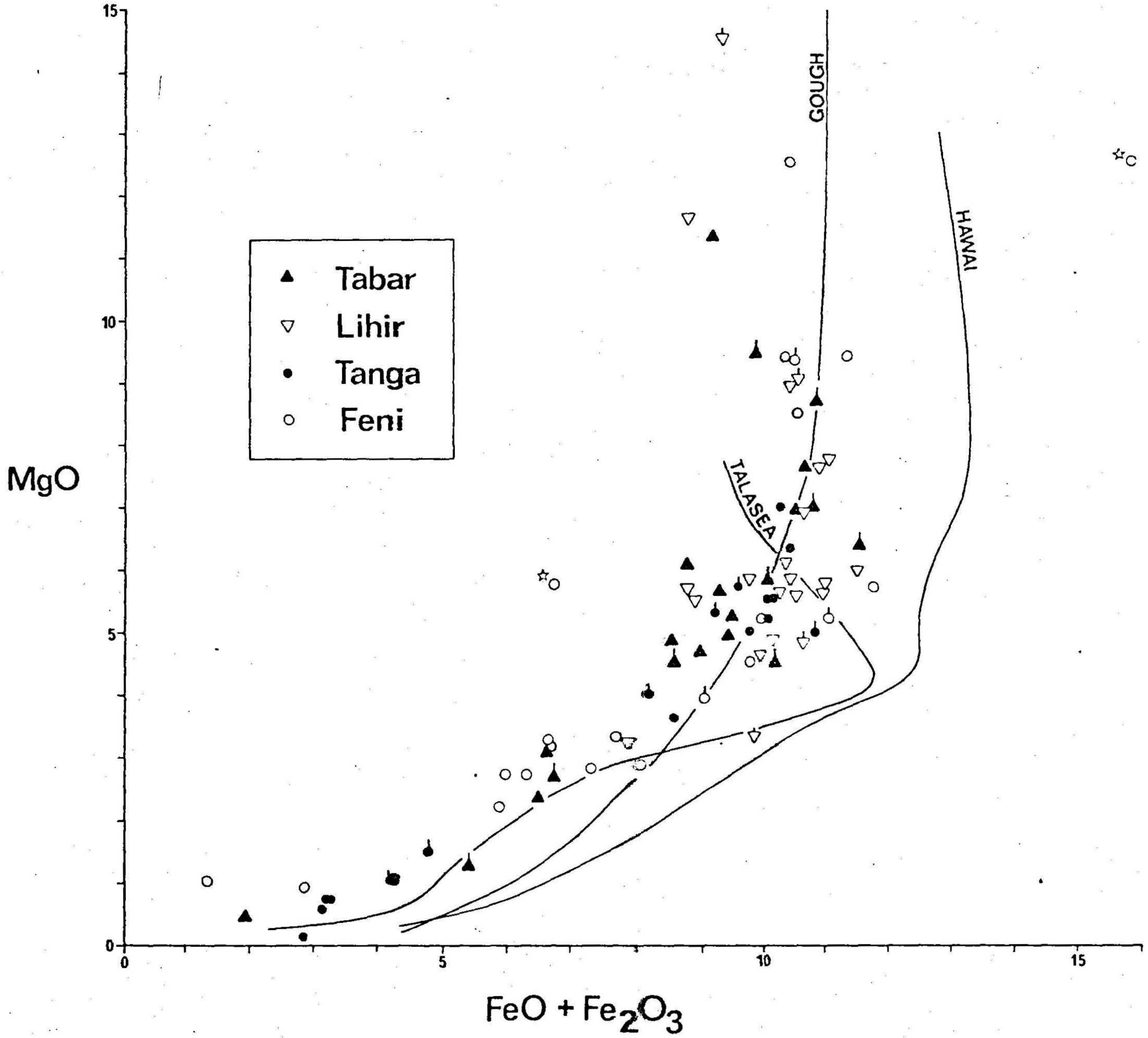
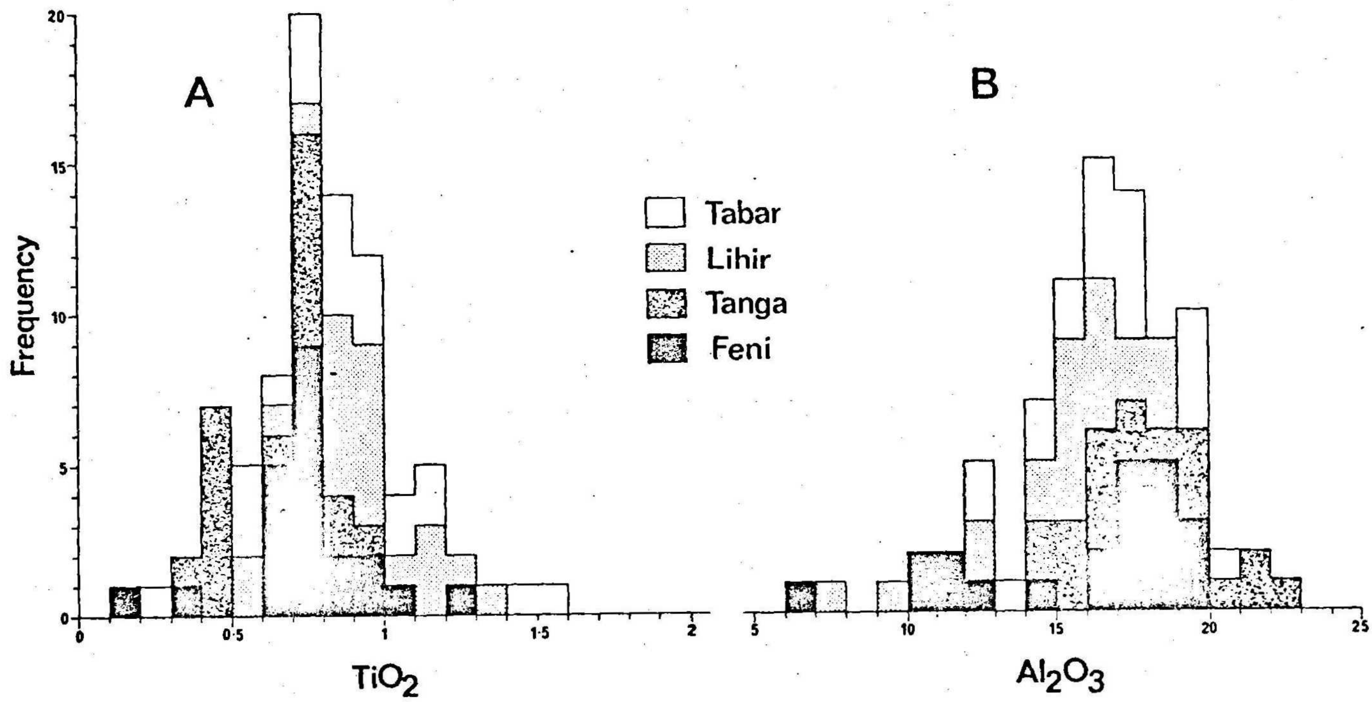
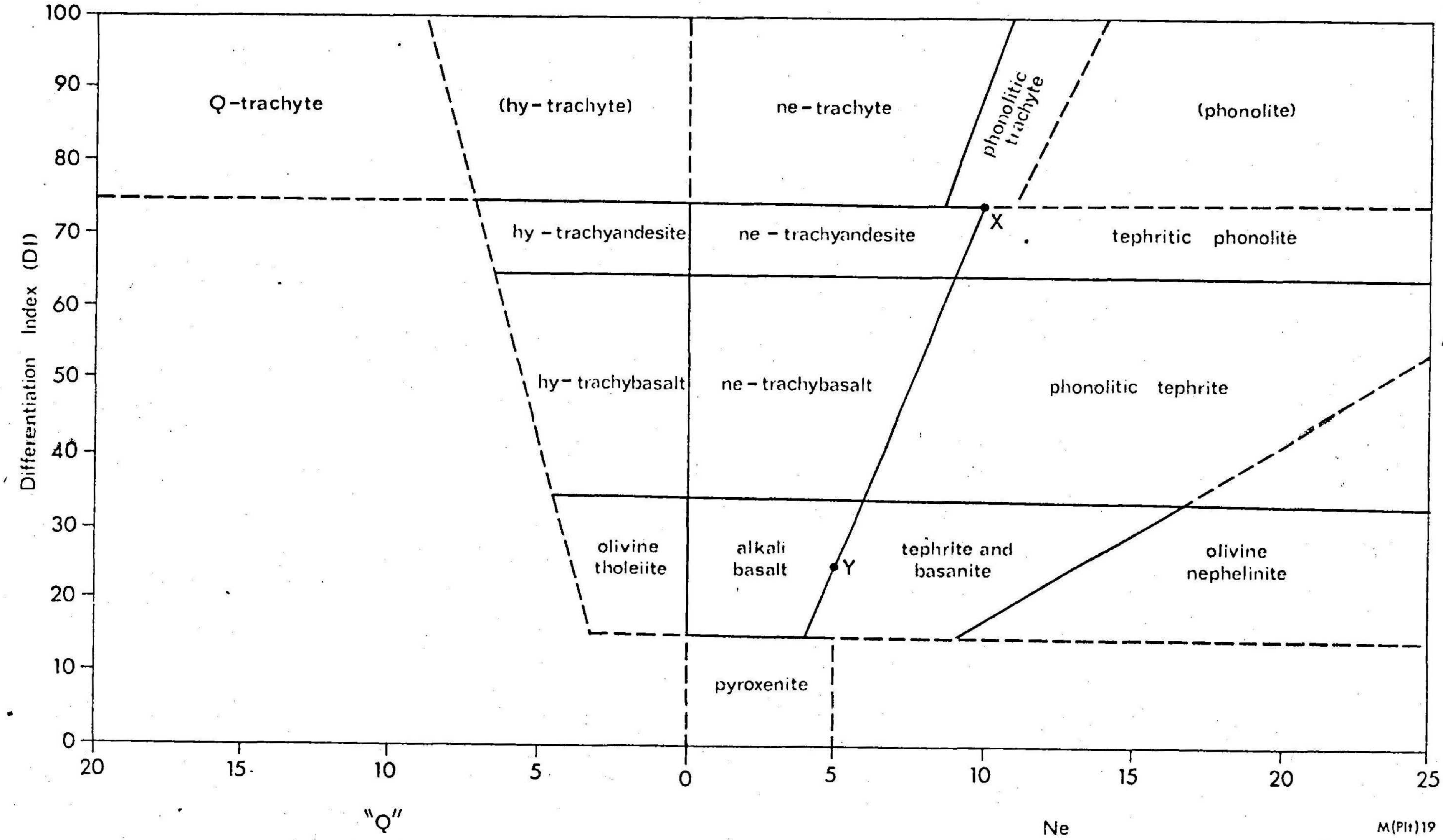


Figure 7. Histograms for (A) $TiO_2$ and (B) $Al_2O_3$ contents of Tabar-to-Feni rocks.



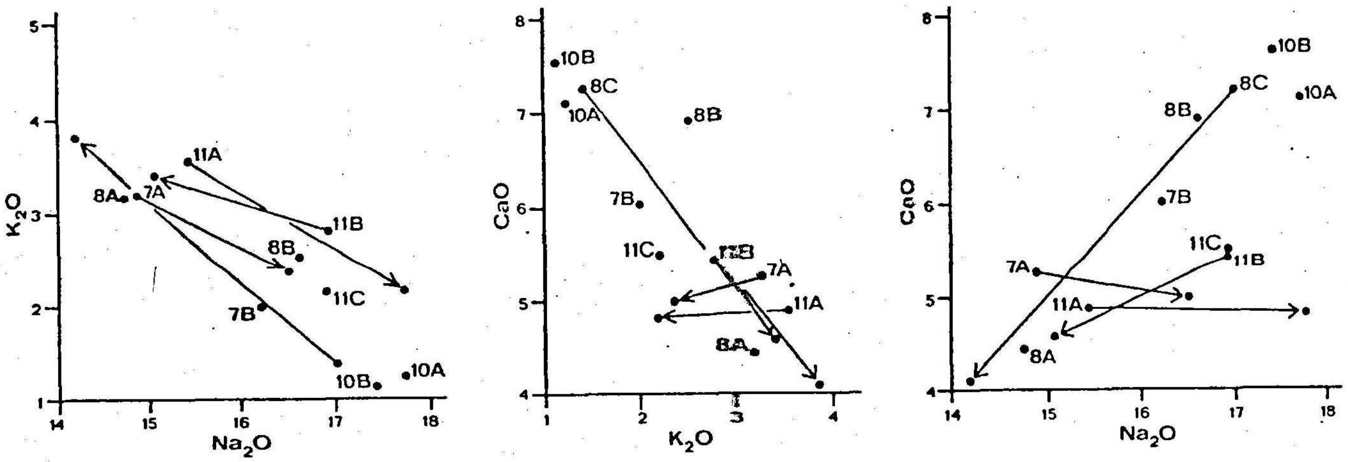


Q-trachyte
(hy-trachyte)
ne-trachyte
phonolitic trachyte
(phonolite)
hy-trachyandesite
ne-trachyandesite
X
tephritic phonolite
hy-trachybasalt
ne-trachybasalt
phonolitic tephrite
olivine tholeiite
alkali basalt
Y
tephrite and basanite
olivine nephelinite
pyroxenite
Differentiation Index (DI)
100
90
80
70
60
50
40
30
20
10
0
20
15
10
5
0
5
10
15
20
25
"Q"
Ne
M(Plt)19


Fig. 8

Figure 8. Classification of Tabar-to-Feni rocks based on CIPW norm calculations and the Differentiation Index of Thornton and Tuttle (1960). Solid lines define field boundaries between rocks represented on the Tabar-to-Feni Islands. Dashed lines are approximate outer boundaries of compositional fields. The horizontal lines correspond to the Differentiation Indices used by Baker et al. (1964) for the rocks of Atlantic oceanic Islands. X-Y is the arbitrary line used by Coombs and Wilkinson (1969) to separate mildly undersaturated rock associations from strongly undersaturated ones. Although Johannsen (1937) and Tilley and Muir (1964) objected to the term "trachyandesite", Coombs and Wilkinson (1969) argued that the term should be retained because of its wide usage. Q-, hy-, and ne-, indicate that quartz (sensu stricto), hypersthene, and nepheline, respectively, appear in the CIPW norm.

Figure A. Weight percent $K_2O$-$Na_2O$-CaO variations in hauyne phenocrysts. Tie lines connect core-rim pairs (arrow heads point to rims). 7, 8, 10, and 11 are rock sample numbers (cf. Table 1), and A, B, and C are different phenocrysts in the same rock.



ERUPTIVE HISTORY OF BAGANA VOLCANO, PAPUA NEW GUINEA, BETWEEN 1882 AND 1975

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ABSTRACT

Bagana is a highly active volcano on Bougainville Island and, unlike the other active volcanoes in Papua New Guinea, appears to be built mainly of thick steep-sided blocky andesitic lava flows with rounded steep-sided fronts up to about 150 m high; pyroclastic deposits are relatively minor. The eruptive history of Bagana before about 1950 is poorly documented and largely unknown. Since 1947 the volcano has been almost continuously active, the longest interval with no reports of any abnormal activity being about 6 years (from late 1954 to late 1959). Eruptions are marked mainly by the effusion of thick, slow-moving, blocky lava flows from the summit area and violently explosive activity which produces nuées ardentes and voluminous ash clouds. Major eruptions occurred in 1950, 1952 and 1966. Lava extrusion may continue for months, even years, and is not always accompanied by explosive activity. Lava was extruded more or less continuously from 1972 to early 1975, but no reports of any explosive activity were received at the Central Volcanological Observatory, Rabaul. The characteristic activity between eruptions is the almost continuous emission of voluminous white vapours from the summit area.

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INTRODUCTION

Bagana volcano ($06^{\circ}$ 09'S; $155^{\circ}$11'E) rises about 1750 m above sea level (Branch, 1967), 19 km northeast of Torokina, in central Bougainville (Figs 1, 2). With the possible exception of Manam, Bagana has been the most consistently active volcano in Papua New Guinea (Fisher, 1954, 1957). However, very little information has been published on the observed eruptive histories of either volcano. This paper is the first attempt to comprehensively collate observations of eruptions from Bagana and to a large extent it parallels a similar paper on Manam volcano (Palfreyman and Cooke, this volume). Sixteen new analyses of rocks from Bagana are also presented; many of the specimens come from lava flows erupted since 1950.

Fig1

GEOLOGICAL SETTING

Bougainville Island is about 203 km long, up to 62 km wide and up to about 2590 m high and is the largest island of the Solomon Group, a northwesterly-aligned island chain on the southwestern margin of the Pacific Ocean. Bougainville is made up of mainly Cainozoic volcanic rocks, sedimentary rocks derived from the volcanics, and subordinate Lower Miocene and Pleistocene limestones; it is dominated by an axial chain of high mountains formed mainly on post-Miocene volcanoes and derived pyroclastic and sedimentary deposits (Blake and Miezitis, 1967; Speight, 1967; Blake, 1968). There are also several dioritic to granodioritic intrusions, some of which may be the cores of deeply eroded volcanoes. A large porphyry - copper type deposit, associated with the intrusion of quartz diorite and granodiorite into andesite, is being mined at Panguna, in southern Bougainville (Blake and Miezitis, 1967; Macnamara, 1968). Pliocene to Recent volcanics cover more than half the island and are the products of several named and unnamed volcanoes (Fig. 2), of which the best known are the active Bagana volcano and the dormant Balbi and Loloru volcanoes (Blake and Miezitis, 1967; Blake, 1968). The other volcanoes on the island are considered to be extinct.

Fig2

Bougainville occupies an island-arc environment (see Coleman, 1970). A deep submarine trench, thought to be marking the boundary between converging plates lies to the south of New Britain and to the west of Bougainville (Fig. 1; Denham, 1969, 1973; Johnson and Molnar, 1972; Curtis, 1973a)), where it reaches a depth of about 9100 m (Denham, 1969). Earthquake activity is high in the vicinity of Bougainville Island and many intermediate-focus earthquakes have been reported (Denham, 1969, 1973; Curtis, 1973a). A well-defined Benioff zone dips northeastwards at between 80° and 90° beneath Bougainville (Denham, 1969). The orientation of the Benioff zone and the focal mechanism solutions for this boundary suggest that the floor of the Solomon Sea underthrusts Bougainville in a northeasterly direction (Denham, 1969, 1973; Ripper, 1970; Johnson and Molnar, 1972; Curtis, 1973a, b; Krause, 1973). Estimates of the rate of subduction of the downthrusted slab range from about 7.8 to 11 cm per year (Curtis, 1973b; Krause, 1973). Cross-sections presented by Denham (1969) and Curtis (1973a) show that very few earthquakes with epicentres at depths between about 200 and 400 km have been recorded in the Bougainville region. The apparent gap in activity may indicate that the lower part of the downthrusted slab beneath Bougainville has become detached from the major part of the plate and has sunk under its own weight (see Oliver et al., 1973; also see Halunen and von Herzen, 1973, for an alternative explanation) and that the deep-focus earthquakes are caused by the disjoined section of lithosphere.

The post-Miocene volcanoes of Bougainville are built up of calc-alkaline lavas and pyroclastic deposits, mostly of andesitic composition, although some dacitic rocks are also present (Blake, 1968). The chemistry and mineralogy of the rocks have not been studied in great detail, but Taylor et al. (1969) described low - Si and high - K andesites from the island and postulated that they originated by processes involving sea-floor spreading and transportation of oceanic crust down the steeply dipping seismic plane beneath Bougainville where it was remelted to form andesites.

BAGANA VOLCANO

Fig. 3

The volcano forms a roughly symmetrical cone with slopes of about 30°, rising above deeply dissected mountains to the south, north and east (Fig. 3). The volcano is situated at the head of the drainage basins of the Torokina and Saua Rivers and is flanked to the west and southwest by a series of low-lying coalescing, volcano-alluvial fans made up mainly of volcaniclastic debris from Bagana. The superficial parts of the cone are built predominantly of thick, blocky andesitic lava flows, but volcaniclastic debris is interspersed between the flows. The surfaces of the flows consist of a jumble of loose, angular, smooth-sided blocks of massive to moderately vesicular andesite up to several metres across and are commonly characterized by hummocks and hollows and prominent transverse, arcuate pressure ridges. Older, deeply eroded lava flows have crudely jointed, massive interiors up to about 30 m thick. The more extensive flows have reached the base of the cone and some abut against and partly overlie the foothills of the adjacent mountains.

The flows are generally bounded for most of their length by prominent marginal levées which project up to about 60 m above the central axes of the flows. The depths of the troughs below the enclosing levées decrease downslope and near the base of the cone the levées lose their identity and the lavas spread out to form broad, steep-sided flows with rounded flow-fronts up to about 150 m high and several hundred metres wide; most flows are less than 50 m wide near the summit. The inner walls of the levées are vertical or dip steeply inwards towards the central axes of the flows and are generally worn smooth as a result of abrasion by lava debris carried down by the hotter, more mobile, central parts of the flows.

Violent explosive activity has also been a feature of many eruptions. The northern flanks of the cone are littered with breadcrust bombs (up to about 1 m long) and large blocks of lava (up to about 1 m across), commonly with impact craters up to about 1 m deep and 2 m in diameter.

SUMMARY OF ERUPTIVE HISTORY BETWEEN 1882 and 1975

The following account is a summary of published data, of unpublished reports of the Bureau of Mineral Resources (Records), and of data in the files of the Central Volcanological Observatory, Rabaul.*

The data on eruptions from Bagana are not comprehensive; many of the records are incomplete; some are conflicting. Bagana is isolated geographically and the area around the volcano only sparsely populated by Europeans. Furthermore, the upper part of the cone is shrouded in clouds for a large part of most days and much minor activity has probably not been reported. Most of the reports of activity came from Europeans living on Bougainville and from the crews of boats and aircraft in the area. Several photographs of the volcano taken between 1943 and 1974 have also been examined by the writer, and form the basis of the sequence of lava flows shown in Figure 4. Because of inadequate records, difficulty was found in correlating some lava flows with reported effusive activity, particularly in the 1950's and 1960's when the volcano was not frequently inspected. Much of the apparent conflict in the reports probably stems from the fact that, on occasions, slow extrusion of lava continued intermittently for years over much the same path.

Fig 4

Activity prior to 1948

Guppy (1887a; p.VIII) reported that in 1882 Bagana was the only volcano in "active eruption" in the Solomon Islands. He also reported being told by a native chieftain of a "great explosion" (p. 22) from Bagana in December 1883 or January 1884 which killed "several natives". According to local informants, the volcano had been in more or less continuous eruption during at least the previous 15 to 20 years (Guppy, 1887b).

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* The writer acknowledges the use of unpublished reports by G.A.M. Taylor, J.G. Best, M.A. Reynolds, G.W. D'Addario, R.F. Heming, I.E. Smith, M.G. Mancini, W.D. Palfreyman, R.A. Davies, and R.J.S. Cooke of the Volcanological Observatory.

In September 1937, loud explosions from Bagana were heard at Kieta, about 50 km away (Fig. 1). Ash ejected during a severe eruption on 15 May 1938 fell on Kieta (Fisher, 1939), and the upper part of the cone became mantled by loose, hot lava blocks; a large area to the southwest of the mountain was covered by ash. Water vapour issued from the flanks of the volcano in numerous places, and boiling springs and hot quicksands were reported from around the base of the cone.

Aerial photographs taken in 1943 show abundant white vapour issuing from the summit area of Bagana, and those taken on 7 April 1943 show a newly erupted lava flow extending down the north-northeastern flank of the volcano (Fig. 5). A later wartime aerial photograph shows the lava flow extending to the base of the cone and bounded by prominent marginal levées for most of its length.

Fig 5

In 1945-46 "light" explosions accompanied by ejection of "scoria" were reported (Fisher, 1957; p. 73). Aerial photographs taken on 20 June 1947 show voluminous white vapour being emitted from the summit area, and another new lava flow with particularly prominent marginal levées extending down the north-northeastern side of the cone. This flow was extruded over the top of the 1943 flow on the upper part of the cone, but bifurcated about halfway down the north-northeastern flank (Fig. 4).

Activity between 1948 and 1953

Towards the end of 1948, spasmodic explosive activity commenced and continued for about 12 months (Fisher, 1957), after which the frequency and intensity of eruptions increased. Eruptions during 1950 produced ash clouds, nuées ardentes (the first reported from Bagana) and lava flows, and a lava dome was extruded in the summit crater. The most violent eruptions were from June to early October, and the height of the activity took place at the end of this period when there were up to 5 eruptions a day, each accompanied by a loud roaring noise. Ash was ejected up to about 10 000 m, and new lava flows were reported. The nuées ardentes completely devastated the upper Saua

River valley (Fig. 4). Subsequent mudflows filled the channel of the river, and much of the low land towards the coast was flooded.

In December 1950, activity was confined to summit vapour emission, extremely slow movement of a lava flow down the south-southwestern flank, and spasmodic rumblings (Taylor, 1956). Forest had been flattened by the passage of recent nuées ardentes up to about 3 km from the base of the mountain on the southern and western sides. The fallen tree trunks were orientated radially away from the crater. Bark on standing stumps was bruised and pitted on the sides facing the volcano but was not charred.

Bagana was "active throughout most of 1951" (Fisher, 1957; p. 73), but activity intensified on 29 February 1952, and continued until October, producing audible rumblings, earth tremors, explosions, nuées ardentes and lava flows. A new crater formed on the northern side of the summit area, and by 10 March 1952 a new lava flow extended down the southwestern flank of the volcano to the base of the cone (Fig. 4). J.G. Best observed 15 nuées ardentes and 45 explosive eruptions between 18 March and 31 March, 1952; most of the nuées ardentes failed to reach the base of the volcano, but new vegetation in part of the area devastated by nuées ardentes in 1950 was stripped and flattened up to about 1 km from the base of the volcano on the western side. Ash from one of the nuées was still hot (90°C) when collected about an hour after deposition. Three of the explosive eruptions observed by Best were extremely violent; ash-laden clouds were ejected to about 10 000 m above sea level. Incandescent blocks of lava were commonly observed at night tumbling down the western flank of the cone from the summit area.

Ash clouds, explosive eruptions, possible nuées ardentes, and rumblings were produced between June and September, 1953. A new crater, reported to have formed on the northern side of the summit during June, produced a series of powerful explosions. Particularly violent eruptions took place on 10 and 25 July and during August 1953, but there were no further reports of activity until 1959.

Activity between 1959 and 1961

In January 1959, G.A.M. Taylor reported that the viscous lava flows produced during the 1948-53 activity had descended the northwestern and southwestern flanks to the foot of the cone. The summit crater was reported to be completely filled by an irregular, hummocky lava dome. In September frequent earth tremors were felt in the vicinity of the volcano, and continuous emission of vapour clouds, frequent audible explosions and a glow in the summit area at night were reported up to early May 1960 when Bagana again erupted with the ejection of ash clouds to about 3 000 m.

During an aerial inspection on 5 May 1960 G.A.M. Taylor observed a lava flow which appeared to be hot and moving down the southwestern flank of the volcano. Photographs taken during the inspection show what appears to be the new flow extending down the side of the cone towards the head of the Saua River and another recent flow extending down the southwestern flank of the volcano to the head of the Torokina River (Fig. 4).

The only reported activity in 1961 was on 26 July when brown, ash-laden clouds were seen rising above the summit area.

Activity between 1962 and 1965

Bagana next erupted on 15 February 1962. Major activity - marked by earth tremors, emission of ash clouds and effusion of lava - continued until April. Part of the eastern rim of the crater was reported to have been destroyed. Many of the streams issuing from the base of the cone were hot and dead fish were common in some.

In July 1962 lava was seen descending the south-southwestern flank of the cone. A divided vapour column suggested that the lava was issuing from two separate vents within the crater. The southeastern part of the crater was reported to be growing and blocks of lava were seen tumbling from this part of the summit. Lava was extruded from the southwestern part of the crater for some time and branched into two separate flows on the lower slopes of the volcano (Figs 4, 6). By 24 June 1963 the two flows had extended to the foot of the mountain - one on the southern flank, the other on the south-southwestern flank (Figs 4, 6).

Fig 6

Comparison of aerial photographs taken in 1960, 1962 and 1963 with those taken in 1947 confirm that in the intervening years several lava flows had been extruded. Their distribution is shown in Figure 4.

Extensive minor damage was caused in southern Bougainville by earth tremors between 17 and 23 April 1964, and on 24 April loud explosions and vast "mushroom-shaped" eruption clouds emanated from Bagana. Between 6 and 9 October, the most prominent activity was the continuous emission of voluminous white vapour clouds from the summit area with "mild steam" explosions from a small crater in the western side of the summit area about every 10 minutes (Branch, 1967; p.11). A deep red glow was seen in the crater at night. A new lava flow, most probably initiated in late April 1964, had descended the south-southwestern flank of the volcano (Fig. 4) and was slowly advancing at the rate of a few centimetres per hour (Branch, 1967). There was no evidence of any recent nuées ardentes.

Activity in 1966 and 1967

Build up of activity early in 1966 culminated in a large eruption at the end of May. Increased vapour emissions from fumaroles on the western flank of Bagana and fluctuating glows at night in the summit area and from several places on the flanks of the volcano were reported on 20 March. Incandescent blocks of lava were observed tumbling down the southwestern slopes of the volcano between 20 and 25 March, resulting in several reports that a new vent had opened up on the western flank of the volcano.

On 1 April an aerial inspection by mining company geologists showed that the summit crater was full (presumably due to growth of a lava dome), but no lavas were reported flowing down the outer flanks. No abnormal activity was detected by a volcanologist from the Observatory during an aerial inspection of the volcano on 4 April.

Observed activity throughout May included: lavas flowing down the southern, southwestern, and eastern flanks of Bagana; severe explosions on 5 and 7 May; red glows in the summit area on nights between 16 and 26 May; and lava flowing on 26 May on the southern and northwestern flanks (with minor extrusion of lava on the eastern flank). Voluminous dark grey ash-laden clouds were erupted on 27 and 29 May and a new vent was reported at the southwestern end of the summit.

On 30 May a large eruption sent a nuée ardente down the southern slopes of the volcano. A northeast-trending ridge at the foot of the cone diverted most of the dense basal part into the Saua River valley where the nuée reached a point about 9.5 km from the crater. At least part of the ash cloud rising above the basal part of the pyroclastic flow jumped the ridge, and continued southwards, destroying a fan-shaped area of forest extending up to about 6 km from the base of the volcano. Up to 4 m of pyroclastic and lava debris were deposited by the nuée on the floor of the upper Saua River valley; the sides of the valley were almost completely stripped of vegetation and most trees were flattened. Movement over irregular terrain produced vortices in the nuée and, in places, trees were felled in the direction from which the nuée descended. The larger of the two newly erupted lava flows extending down the southern side of the cone had reached the lower slopes of the volcano by the time of the ground inspection between 10 and 13 October 1966. The lava flow was reported to have been extruded through a prominent breach in the southern wall of the crater and was still slowly moving. By 12 April 1967 the front of the larger of the 1966 lava flows on the southern flanks of the volcano had reached the head of the Saua River valley (Fig. 4).

Bagana was inspected from the air by R.F. Heming on 23 September 1967, following a report of emission of a dark brown ash cloud. The summit area was described as containing a clearly defined crater, about 200 m in diameter, with a small lava dome which spilled through a break in the south-southeastern rim of the crater. Dense white vapour clouds were being continuously emitted from the surface of the dome and from numerous small vents around the crater rim. A new lava flow, about 160 m long, was observed to the east of the larger 1966 flow on the southern flank. Heming postulated that lava had been extruded more or less continuously since May 1966 and that more mobile lava had recently broken through the solidified crust of the 1966 flow in the summit area and started a new lava flow. The front of the 1966 flow at the head of the Saua River valley had not moved since the previous April but the steep upper parts of the flow had been mobile in the interim. Withdrawal of more mobile lava from the axial parts had caused the centre of the flow to subside, producing prominent marginal levees. Lava had piled up on the lower slopes of the cone and formed a large protruberance near where the lava flow makes a fairly sharp bend and flattens out (Fig. 7).

Emission of black ash-laden clouds were reported on 23 October, 22, 23 and 30 November, 1967.

Activity between 1969 and 1974

Activity during an aerial inspection on 4 May 1970 consisted of emissions of dense white vapour from many parts of the summit crater, including the surface of a lava dome. This dome, which began growing after the May 1966 eruption, appeared to fill the old crater and almost to have reached the level of the crater rim. It was reported that there was very little change in the state of activity since the last inspection on 3 December, 1969.

A fresh overflow of blocky lava, less than 300 m long, was observed blocking the southeastern breach in the crater rim during an aerial inspection on 6 June 1970, and some fresh lava was also noted on the northwestern side of the cone. The lava dome was reported to be at a lower level than when inspected in December, 1969. Continuous emission of white vapour from the crater rim and lava dome and intermittent ash-laden vapour ejections, some accompanied by rumbling explosions, took place during the latter half of 1970 and the first half of 1971. On 11 October 1970 a blocky lava flow of recent origin was reported to extend down the western side of the cone to about 1000 m above sea-level. Incandescent blocks of lava were observed cascading down the front of the flow and white vapour was being discharged from its upper surfaces. This flow had not been seen during the last inspection of Bagana in June 1970, and was not mentioned in later Observatory reports.

Explosive volcanic activity was renewed in March 1971, reaching a climax late in the month. A new blocky lava flow was observed on the southern flank of the cone on 23 March and 2 April, and was reported to have been extruded through a breach in the southeastern section of the crater wall; it followed much the same path as, and extensively covered, the larger of the flows extruded in 1966 down the southern flanks (Fig. 4). High marginal levées and transverse pressure ridges were well developed. On 23 March the crater appeared to be completely filled with blocky lava and a vent had formed near the centre of the crater, but by 2 April some of the lava had spilled through the breach in the crater rim.

Activity during late March and early April 1971 consisted of single explosive events which produced voluminous "cauliflower-shaped" ash-laden clouds up to about 5 500 m above sea level. The frequency and magnitude of these explosive events slowly diminished during April, and activity had ceased altogether by 24 April.

A new blocky lava flow extending to the base of the cone was observed during the aerial inspection on 24 February 1973. A single report of a red glow in the summit one night, a vague story of a possible explosion heard from a remote area and second hand reports (undated) of red glows seen by villagers are the only events reported in 1972 and early 1973. The volcano had been previously inspected on 9 December 1971 but no new lava flows were reported. However, photographs taken during the inspection show a recent lava flow at the foot of the cone on the eastern side of the main 1971 lava flow (Fig. 4). It is highly likely that this flow was a forerunner to, and forms part of, the voluminous lava flow observed in February 1973 (and labelled 1972-73 in Fig. 4). A huge cloud of brownish "smoke" had been reported billowing out of the side of the new lava flow, about 150 m below the summit, on 20 February 1973. In the light of the writer's observations between 15 and 27 November 1973 it seems certain that the cloud was produced by lava debris avalanching down the side of the lava flow. Most of the vapour emissions observed on 24 February 1973 originated from a lava dome which did not appear to overlap the crater rim. It was reported that the newest flow and the other recent flows did not appear to have flowed through the prominent breach in the southern rim of the crater, but to have overflowed relatively high parts of the southeastern rim.

Lava was still being slowly extruded from the summit when the writer visited the area in November 1973. Blocks in the upper central part of the new flow reported on 24 February 1973 had fresh black surfaces whereas earlier erupted lava blocks along the margins of the flow and on the surface near the front of the flow had dull grey weathered surfaces (Fig. 8). Lava in the central channel was at about the same height as the stationary margins. Many of the blocks at and near the front of the flow were warm; a few were too hot to handle. White vapour issued under low pressure from numerous points along the length of the flow. Smells of hydrogen sulphide were particularly strong down-wind from the summit crater. On several nights a dull red glow was observed in the summit area. Glimpses of a lava dome were obtained from a vantage point about 100 m below the summit on the southeastern flank of the volcano and extensive deposits of sulphur (cf. Branch, 1967) were observed on the upper part of the cone from high on the northern flank of the volcano.

Avalanches of loose lava blocks were common from the summit area, down the steep sides of the new flow, and off the face of the flow, particularly where it took a sharp downward bend about halfway up the cone. The cascading blocks generated large billowing clouds of pale brown dust (finely comminuted lava debris) formed by attrition between blocks in the slowly moving lava. Blocks of incandescent lava tumbling down the flow face gave rise to red streaks in the night and generated voluminous dust clouds that obscured the sky.

Quiet effusive activity continued during 1974 and early 1975 resulting in the building of a very thick broad lava flow on the lower slopes of the cone. However, the lava flow appeared inactive when inspected from the air by R.J.S. Cooke on 26 June 1975. Lava in the central channel of the flow had drained downhill leaving high marginal levées. A small flow of recent origin was reported on the southeastern side of the cone, adjacent to the 1972-75 flow, and lava was observed flowing down the north-northwestern flank (Fig. 4). The prominent lava dome ~~(Fig. 10)~~ observed by Cooke on 24 February 1973 had been destroyed and a new dome was reported to be building up in the summit area.

PETROLOGY

Sixty thin sections of volcanic rocks from Bagana have been
Table 1 examined. Chemical analyses of 16 specimens are presented in Table 1. In common with most volcanic rocks from island arcs, those of Bagana are highly porphyritic, especially in plagioclase (Table 1). Clinopyroxene (probably mainly augite) is the next most abundant ubiquitous phenocryst, and all the rocks examined contain minor amounts of iron - titanium oxide phenocrysts. Some rocks contain rare, small olivine phenocrysts (<1 percent by volume) which, in a few samples, are mantled by small grains of calcium-rich clinopyroxene; coronas of calcium-poor pyroxene are absent. The majority of the lavas contain no olivine, minor orthopyroxene as small, faintly pleochroic

Some lavas contain scattered inclusions, the most common type bearing abundant large white plagioclase phenocrysts. These inclusions superficially resemble a porphyritic micro - ?diorite intrusive exposed beneath the lavas near the head of the Saua River. Rare inclusions of andesite and layered gabbro were also found.

Fig 9

The rocks display only a very limited range in major element abundances (Table 1; Figs 9a, b) and are similar in composition to those previously described from Bagana by Blake and Miezitis (1967) and Taylor et al. (1969). They are quartz-normative, predominantly low-silica (53 per cent $< SiO_2 <$ 57 per cent; Taylor et al., 1969) andesites with major element abundances which appear to be typical of calc-alkaline andesites from island arcs. In recent years, the cause of variation in potassium content of volcanic rocks in island arc regions has received much attention. Rocks with similar $K_2O$ contents (for comparable $SiO_2$ contents) have been reported from several parts of Papua New Guinea (Jakeš and White, 1969; Jakeš and Smith, 1970; Mackenzie and Chappell, 1972; Johnson et al., 1973), but the significance of the chemistry of the andesites from Bagana cannot be fully evaluated until the chemistry of volcanic rocks from other post-Miocene volcanoes on Bougainville has been studied in detail. Andesites with $K_2O$ contents greater than 2.5 per cent have been reported from Balbi volcano and Malabita Hill on Bougainville (Blake and Miezitis, 1967; Taylor et al., 1969).

Cainozoic volcanic rocks so far analysed from Bougainville fall in a straight band extending from near the centre of the AFM diagram towards the alkalies apex (Fig. 9a), although they are more enriched in iron than calc-alkaline volcanics from the eastern Papua New Guinea mainland (Jakeš and White, 1969; Johnson et al., 1973). They plot mainly in the field of the "high-alumina basalt series" of Kuno (1968; p. 627) on a total alkalies versus silica diagram (Fig. 9b) and show only a gradual overall increase in total alkalies with increasing $SiO_2$ content. Analyses of rocks from Bagana listed by Blake and Miezitis (1967) and Taylor et al. (1969) tend to have slightly higher total alkalies than those listed in Table 1 (also see Fig. 9b). The reason for this is not known, but it could possibly be due to systematic analytical errors. Strontium isotope analyses of 4 of their specimens (from Bagana, Balbi and Takuan volcanoes, and Malabita Hill) yielded $Sr^{87}/Sr^{86}$ ratios ranging from 0.7038 to 0.7040 (Page and Johnson, 1974). Page and Johnson interpreted these uniform and relatively low ratios as indicating that the lavas were derived from relatively homogeneous upper mantle sources with relatively little (if any) contamination by old radiogenic crustal material.

DISCUSSION

Bagana vies with Manam for being the most active volcano in Papua New Guinea but because of its isolated position the eruptive history is poorly documented, especially prior to about 1950. Since the Second World War the volcano has been almost continuously active, the longest interval with no reports of any abnormal activity being about 6 years (from late 1954 to late 1959). Major eruptions, characterized by violent explosive activity occurred in 1950, 1952 and 1966, during which voluminous ash clouds, nuées ardentes and lava flows were produced; fifteen nuées ardentes were observed in a 14-day period in March 1952. However, effusion of lava is not always accompanied by explosive activity. Lava was extruded more or less continuously from 1972 to early 1975 but there was virtually no reports of any associated explosive activity.

Bagana differs from the other active volcanoes in Papua New Guinea in that the principal eruptive activity has been the extrusion of viscous, sluggish, blocky lava flows of predominantly low-silica andesite composition; pyroclastic deposits are relatively minor. Slow extrusion of lava commonly continues for months, in some cases (for example, the 1972-75 lava flow) for years, and relatively thick (up to about 150 m) steep-sided flows are produced. Since 1943 at least 18 lava flows with fronts at the base of the cone or on the lower and intermediate slopes of the volcano have been extruded. It has not been possible to determine any cycle or periodicity in the pattern of eruptive events, possibly because of the general lack of comprehensive observations; nor is there any apparent link between eruptions from Bagana and eruptions from volcanoes along the southern margin of the Bismarck Sea, six of which erupted between 1972 and 1975 (Cooke et al., this volume). Of the lava specimens analysed, none (with the possible exception of 74400065 whose age is not precisely known) of those erupted since 1950 have $SiO_2$ contents greater than 55.5 per cent. It is the relatively old lava flows that are characterized by relatively low total iron, CaO, MgO and $TiO_2$, and high $SiO_2$, contents.

Since the first recorded observations in 1882 the main activity from Bagana between eruptive episodes has been the almost continuous emission of dense white vapours from the summit area. Marked fluctuations (generally of short duration) in the volume of vapours emitted and wisps of pale blue "smoke" were observed by the writer and have been noted by other observers.

Bagana is far removed from any major centres of population, but there are more than 30 villages within a radius of about 20 km of the volcano. Except for falls of ash, "normal" eruptions from Bagana are unlikely to seriously affect many areas beyond the base of the volcano. The lava flows are slow-moving and none have reached far beyond the foot of the cone. However, nuées ardentes and mudflows pose a potential threat to areas far beyond the base of the volcano. The area most likely to be affected is the low open country drained by the Torokina and Saua Rivers on the western and southwestern sides of the volcano. In May 1966 a nuée ardente swept down the valley of the Saua River for about 9.5 km from the crater. In 1950 nuées completely devastated the upper part of the Saua River valley. The data therefore indicate that centres close to, or in, the valleys of the Saua and Torokina Rivers and within 15-20 km of the volcano are likely to suffer adversely in a very major or catastrophic eruption.

ACKNOWLEDGEMENTS

The writer acknowledges the assistance provided by the Geological Survey of Papua New Guinea in making the study possible. He also thanks R.J.S. Cooke, C.O. McKee and R. Bill of the Central Volcanological Observatory, Rabaul, for help received during the field work and preparation of the paper, and his wife, Joyce, for drawing the diagrams. This paper is published with the permission of the Director, Bureau of Mineral Resources, Canberra.

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## Captions of figures 1 to 9

Fig. 1. Papua New Guinea and the northwestern part of the British Solomon Islands.

Fig. 2. Sketch map of Bougainville Island.

Fig. 3. Bagana volcano from the north on 5 May 1960, showing some of the lava flows erupted during the previous seventeen years (see Fig. 4), and voluminous white vapours issuing from the summit area. Photograph taken by G.A.M. Taylor.

Fig. 4. Recent lava flows from Bagana volcano. The small dome shown partly encircled by lava flows at the foot of the cone on the northwestern side of the volcano is densely vegetated and poorly exposed; it may have formed relatively early in the development of Bagana volcano or it may be due to unrelated volcanic activity. The sample numbers shown are all preceded by the prefix 744000-.

Fig. 5. Bagana from the north on 7 April 1943, showing emission of white vapours from the summit area and a newly erupted lava flow extending more than half way down the northern side of the cone. Note that lava in the axial parts of the flow is darker than lava along the margins and forming the surface of the adjacent older flows. In the foreground is the crater of the extinct Billy Mitchell volcano. The deeply dissected extinct Reini volcano lies between, and to the left of, Billy Mitchell and Bagana volcanoes. Behind Bagana are the channels of the Saua (left) and Torokina (right) Rivers. Photograph taken by U.S. Air Force.

Fig. 6. Vertical aerial photograph of Bagana volcano taken on 24 June 1963, showing recent lava flows on the southern and south-western flanks of the volcano (see Fig. 4). Photograph taken by Royal Air Force.

Fig. 7. Close-up of hummocky protruberance in the more extensive of the 1966 lava flows on the southern flank of Bagana volcano. The protruberance formed as a result of the withdrawal of lava from the central part of the flow on the higher, steeper part of the cone and its accumulation on the lower more gently sloping part of the cone. Photograph taken by R.F. Heming on 23 September 1967.

Fig. 8. Upper part of Bagana on 18 November 1973 from the eastern side of the 1972-75 lava flow, showing: (1) recently erupted dark lava in the central parts of the flow and dull grey, older, more weathered lava on the margins of the flow; (2) voluminous white vapours emanating from the southeastern part of the summit area and from an apron of recently erupted lava on the upper southeastern part of the cone; (3) the distinct notch in the rim of the summit area to the left of the 1972-75 flow which appears to have been extruded over a high part of the crater rim; and (4) the grey streak (arrowed) down the side of the 1972-75 lava flow marking the site of rock falls from the top of the flow.

Fig. 9. Chemical variation diagrams for Cainozoic volcanic rocks from Bougainville Island.
a. AFM (total iron as FeO) diagram. The limits of Kuno's (1968; p. 649) "hypersthenic rock series" (calc-alkaline series) are shown by the broken lines.
b. Total alkalies versus silica diagram. The boundary between Kuno's (1968; p. 627) "pigeonitic rock series" (tholeiitic series - below) and his "high-alumina basalt series" (above) is shown by the broken line.

~~Fig. 10. Lava dome in summit crater of Bagana volcano, 24 February 1975. Photograph taken by R.J.S. Cooke.~~

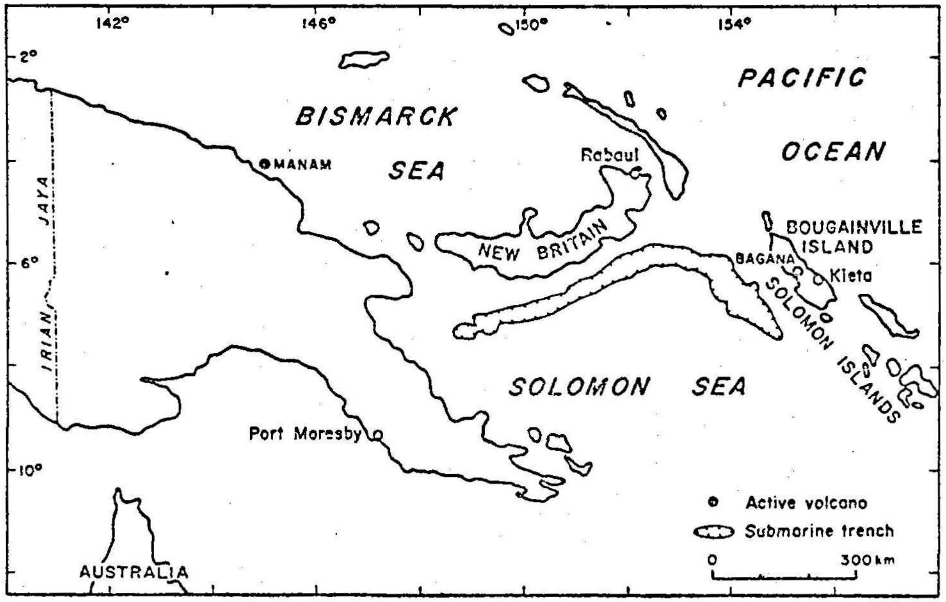

142°
146°
150°
154°
2°
6°
10°
PACIFIC
OCEAN
BISMARCK
SEA
MANAM
Rabaul
NEW BRITAIN
BOUGAINVILLE
ISLAND
BAGANA
Kieta
IRIAN JAYA
SOLOMON ISLANDS
SOLOMON SEA
Port Moresby
AUSTRALIA
Active volcano
Submarine trench
0
300 km

Fig. 1

#22

Fig. 2.

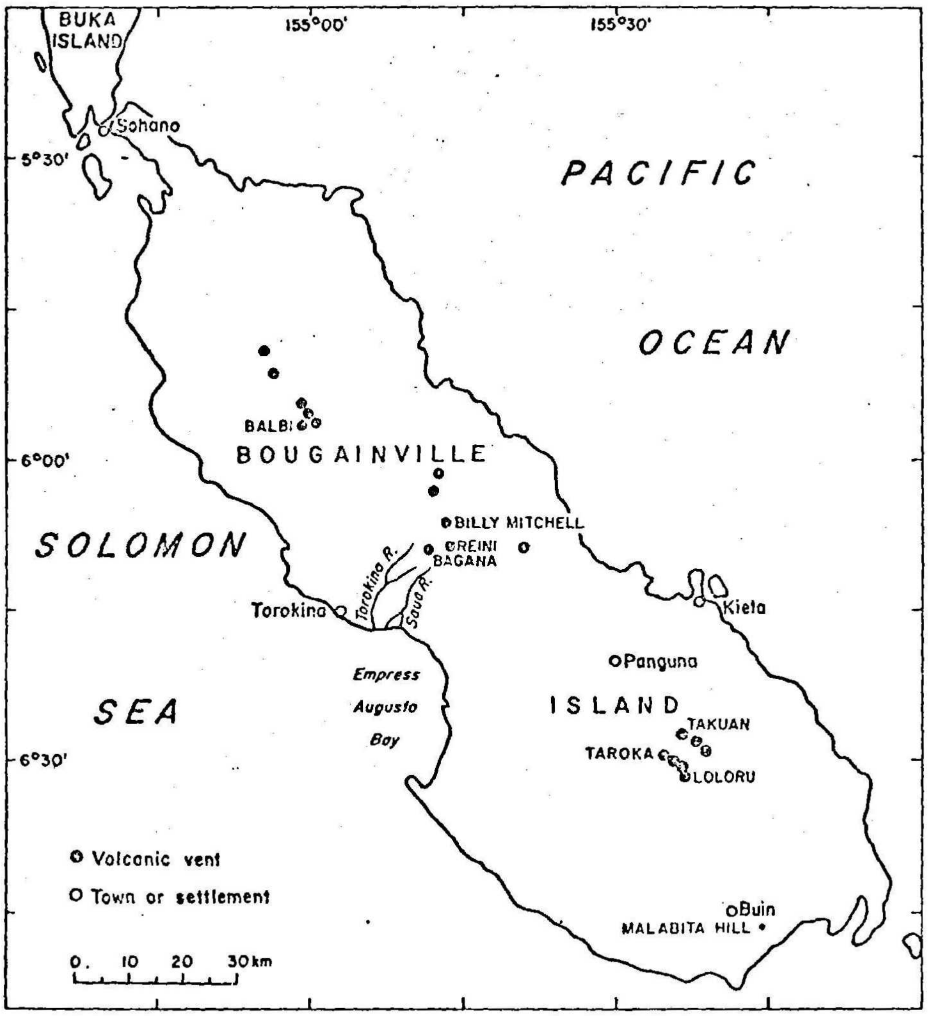
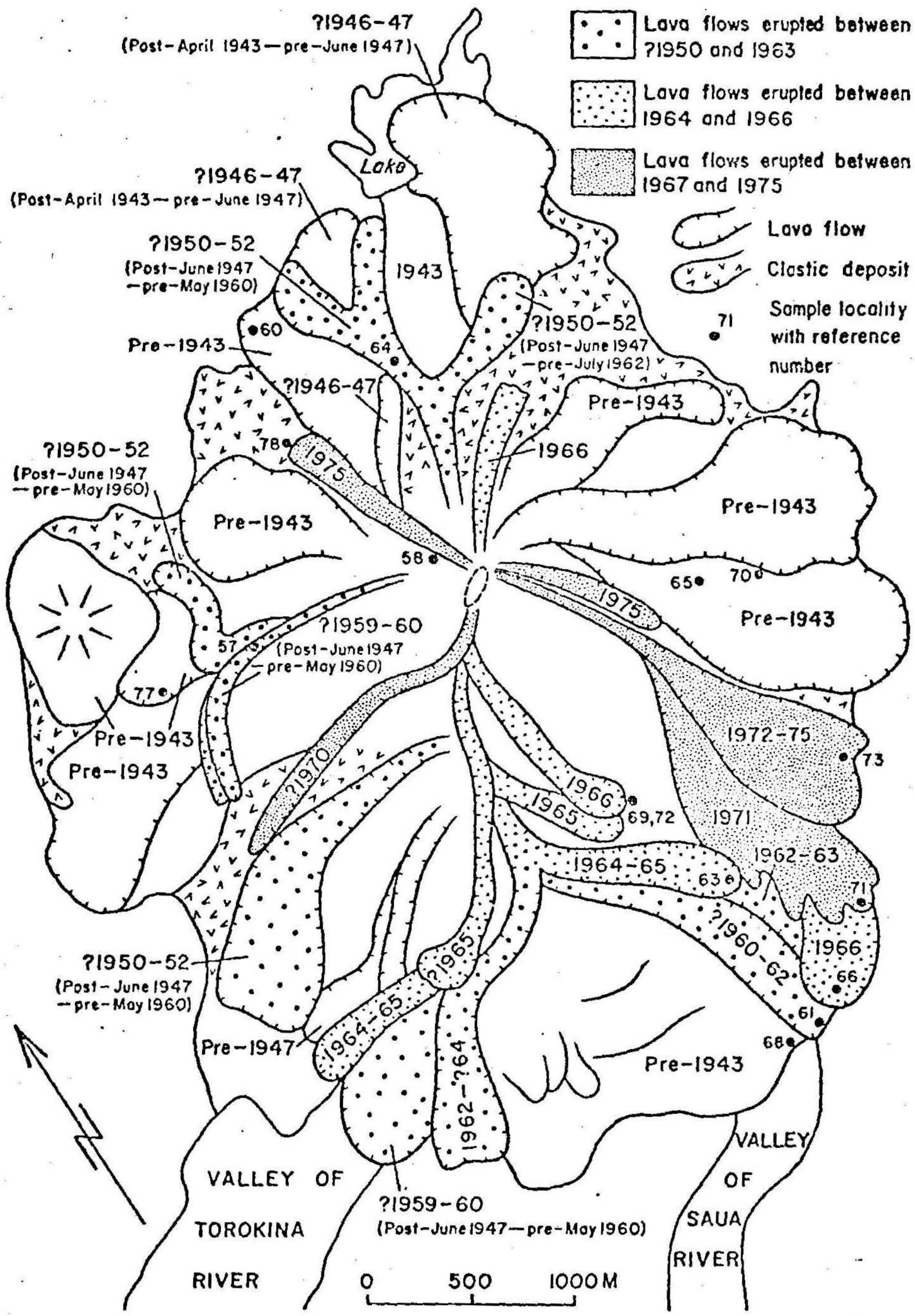

BUKA ISLAND
155°00'
155°30'
Sohano
5°30'
PACIFIC
OCEAN
BALBI
BOUGAINVILLE
6°00'
BILLY MITCHELL
SOLOMON
REINI
BAGANA
Torokina R.
Saua R.
Torokina
Kieta
Panguna
Empress
Augusta
Bay
SEA
ISLAND
TAKUAN
TAROKA
LOLORU
6°30'
Volcanic vent
Town or settlement
Buin
MALABITA HILL
0 10 20 30 km

Fig 4



?1946-47
(Post-April 1943—pre-June 1947)
Lava flows erupted between ?1950 and 1963
Lava flows erupted between 1964 and 1966
Lava flows erupted between 1967 and 1975
Lava flow
Clastic deposit
Sample locality with reference number
71
Lake
?1946-47
(Post-April 1943—pre-June 1947)
?1950-52
(Post-June 1947 —pre-May 1960)
1943
?1950-52
(Post-June 1947 —pre-July 1962)
60
Pre-1943
64
?1946-47
Pre-1943
78
1975
1966
?1950-52
(Post-June 1947 —pre-May 1960)
Pre-1943
Pre-1943
58
65
70
1975
Pre-1943
?1959-60
57
(Post-June 1947 —pre-May 1960)
77
Pre-1943
Pre-1943
1972-75
73
?1970
1966
1965
69,72
1971
1962-63
1964-65
63
71
?1960-62
1966
66
?1950-52
(Post-June 1947 —pre-May 1960)
?1965
61
68
Pre-1947
1964-65
?1962-64
Pre-1943
VALLEY OF TOROKINA RIVER
VALLEY OF SAUA RIVER
?1959-60
(Post-June 1947—pre-May 1960)
0 500 1000M

Fig. 3
Neg. No
5 35/3

Fig 5
Neg. No
G AB/64

Fig. 6

Neg. No

GAB/62

Fig 7

Neg. No

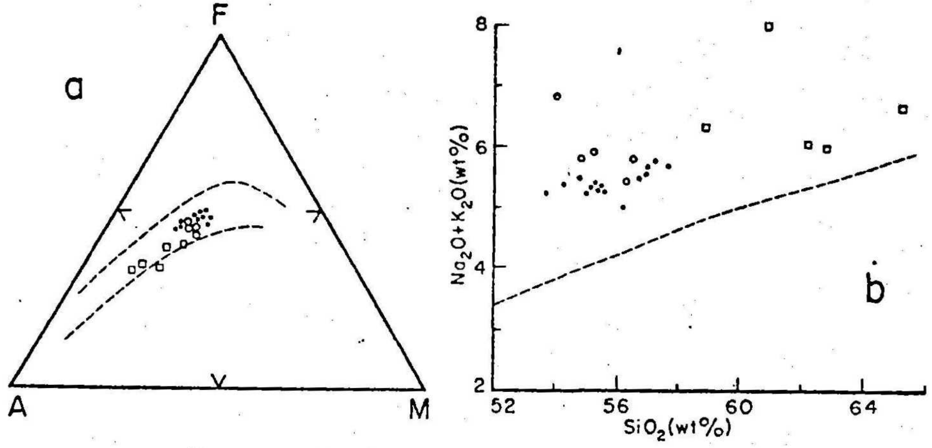
533/7

Fig. 8

Neg. No.

M/1545

Fig. 9




• Bagana volcanics — this study
∘ Bagana volcanics — Taylor et al.,1969
▫ Other Bougainville volcanics—Taylor et al.,1969

Table 1. CHEMICAL* AND MODAL ANALYSES OF 16 ROCKS FROM BAGANA VOLCANO

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 | 15 | 16 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 53.7 | 53.7 | 54.3 | 54.8 | 55.0 | 55.2 | 55.3 | 55.4 | 55.5 | 55.5 | 56.2 | 56.7 | 56.9 | 56.9 | 57.2 | 57.6 |
| $TiO_2$ | 0.96 | 0.88 | 0.85 | 0.83 | 0.85 | 0.81 | 0.82 | 0.85 | 0.80 | 0.81 | 0.73 | 0.74 | 0.73 | 0.67 | 0.67 | 0.68 |
| $Al_2O_3$ | 17.8 | 17.1 | 17.6 | 17.8 | 17.8 | 17.9 | 18.2 | 17.9 | 18.0 | 17.9 | 17.3 | 18.0 | 18.1 | 17.9 | 18.1 | 18.1 |
| $Fe_2O_3$ | 4.10 | 4.15 | 3.90 | 4.35 | 4.25 | 3.45 | 4.45 | 4.15 | 3.95 | 3.95 | 5.05 | 4.35 | 3.65 | 5.65 | 3.30 | 3.95 |
| FeO | 5.10 | 4.95 | 4.80 | 4.10 | 4.25 | 4.85 | 3.70 | 4.30 | 4.30 | 4.25 | 2.80 | 3.15 | 3.50 | 1.90 | 3.80 | 2.90 |
| MnO | 0.18 | 0.17 | 0.17 | 0.17 | 0.17 | 0.18 | 0.18 | 0.17 | 0.17 | 0.17 | 0.16 | 0.16 | 0.16 | 0.16 | 0.16 | 0.16 |
| MgO | 3.85 | 4.25 | 3.85 | 3.65 | 3.45 | 3.35 | 3.40 | 3.35 | 3.25 | 3.30 | 3.80 | 2.70 | 2.70 | 2.75 | 2.65 | 2.55 |
| CaO | 8.30 | 8.55 | 8.50 | 8.05 | 8.20 | 8.10 | 7.95 | 8.10 | 7.90 | 8.10 | 8.10 | 7.40 | 7.45 | 7.55 | 7.30 | 7.30 |
| $Na_2O$ | 3.60 | 3.60 | 3.75 | 3.80 | 3.70 | 3.75 | 3.75 | 3.75 | 3.70 | 3.75 | 3.60 | 3.85 | 4.00 | 3.95 | 4.05 | 3.95 |
| $K_2O$ | 1.60 | 1.61 | 1.61 | 1.65 | 1.50 | 1.57 | 1.61 | 1.51 | 1.61 | 1.51 | 1.40 | 1.66 | 1.66 | 1.61 | 1.70 | 1.70 |
| $P_2O_5$ | 0.35 | 0.36 | 0.36 | 0.35 | 0.34 | 0.36 | 0.35 | 0.35 | 0.37 | 0.35 | 0.31 | 0.35 | 0.32 | 0.35 | 0.34 | 0.34 |
| $H_2O^+$ | 0.02 | 0.17 | 0.02 | <0.01 | 0.06 | 0.04 | <0.01 | 0.01 | 0.07 | 0.04 | 0.04 | 0.50 | 0.19 | 0.09 | 0.15 | 0.21 |
| $H_2O^-$ | 0.08 | 0.07 | 0.06 | 0.06 | 0.06 | 0.08 | 0.05 | 0.05 | 0.03 | 0.06 | 0.06 | 0.10 | 0.13 | 0.17 | 0.15 | 0.11 |
| $CO_2$ | <0.05 | 0.05 | 0.05 | 0.05 | 0.05 | 0.05 | <0.05 | 0.05 | 0.05 | <0.05 | 0.05 | <0.05 | 0.05 | 0.05 | 0.05 | 0.05 |
| Total S | 0.01 | 0.02 | 0.015 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.02 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.015 |
| Total | 99.65 | 99.63 | 99.84 | 99.67 | 99.69 | 99.70 | 99.77 | 99.95 | 99.72 | 99.70 | 99.61 | 99.67 | 99.55 | 99.71 | 99.63 | 99.62 |
| CIPW norms | | | | | | | | | | | | | | | | |
| Q | 4.41 | 4.11 | 4.33 | 5.86 | 7.00 | 6.12 | 7.26 | 7.24 | 7.50 | 7.30 | 10.30 | 10.22 | 9.02 | 10.52 | 8.77 | 10.83 |
| or | 9.50 | 9.57 | 9.54 | 9.79 | 8.90 | 9.32 | 9.54 | 8.93 | 9.55 | 8.96 | 8.31 | 9.90 | 9.88 | 9.57 | 10.11 | 10.12 |
| ab | 30.59 | 30.64 | 31.80 | 32.27 | 31.43 | 31.85 | 31.81 | 31.75 | 31.42 | 31.85 | 30.60 | 32.87 | 34.10 | 33.60 | 34.49 | 33.65 |
| an | 27.82 | 25.91 | 26.51 | 26.75 | 27.66 | 27.50 | 28.16 | 27.59 | 27.87 | 27.67 | 27.05 | 27.19 | 26.74 | 26.51 | 26.37 | 26.84 |
| di wo | 4.70 | 5.88 | 5.46 | 4.48 | 4.45 | 4.25 | 3.80 | 4.19 | 3.65 | 4.33 | 4.58 | 3.15 | 3.37 | 3.56 | 3.14 | 2.96 |
| di en | 2.95 | 3.83 | 3.48 | 3.12 | 2.99 | 2.51 | 2.76 | 2.77 | 2.35 | 2.82 | 3.94 | 2.38 | 2.27 | 3.08 | 1.94 | 2.23 |
| di fs | 1.46 | 1.64 | 1.63 | 0.99 | 1.12 | 1.52 | 0.68 | 1.12 | 1.05 | 1.21 | 0.03 | 0.45 | 0.85 | 0.00 | 1.01 | 0.43 |
| hy en | 6.68 | 6.81 | 6.13 | 6.01 | 5.63 | 5.86 | 5.73 | 5.58 | 5.77 | 5.42 | 5.57 | 4.40 | 4.50 | 3.81 | 4.70 | 4.17 |
| hy fs | 3.29 | 2.91 | 2.88 | 1.90 | 2.10 | 3.55 | 1.42 | 2.26 | 2.59 | 2.32 | 0.04 | 0.83 | 1.68 | 0.00 | 2.45 | 0.81 |
| il | 1.83 | 1.68 | 1.62 | 1.58 | 1.62 | 1.54 | 1.56 | 1.62 | 1.53 | 1.54 | 1.39 | 1.42 | 1.40 | 1.28 | 1.28 | 1.30 |
| mt | 5.97 | 6.06 | 5.67 | 6.33 | 6.19 | 5.02 | 6.47 | 6.02 | 5.75 | 5.75 | 7.36 | 6.37 | 5.33 | 4.73 | 4.82 | 5.77 |
| ap | 0.83 | 0.86 | 0.86 | 0.83 | 0.81 | 0.86 | 0.83 | 0.83 | 0.88 | 0.83 | 0.74 | 0.84 | 0.76 | 0.83 | 0.81 | 0.81 |
| he | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 2.42 | 0.00 | 0.00 |
| cc | 0.00 | 0.11 | 0.11 | 0.11 | 0.11 | 0.11 | 0.00 | 0.11 | 0.11 | 0.00 | 0.11 | 0.00 | 0.11 | 0.11 | 0.11 | 0.11 |
| Volume percent phenocrysts | | | | | | | | | | | | | | | | |
| Plagioclase | 28 | 22 | 25 | 27 | 24 | 32 | 38 | 29 | 33 | 33 | 28 | 33 | 35 | 29 | 27 | 25 |
| Olivine | <1 | <1 | – | – | – | – | <1 | – | – | – | – | – | – | – | – | – |
| Orthopyroxene | – | – | – | <1 | 1 | <1 | 2 | 1 | 1 | 2 | 1 | 1 | <1 | 1 | <1 | <1 |
| Clinopyroxene | 10 | 13 | 10 | 8 | 4 | 7 | 7 | 11 | 8 | 7 | 9 | 7 | 10 | 6 | 7 | 8 |
| Opaque Oxides | 3 | 4 | 3 | 3 | 2 | 4 | 5 | 3 | 3 | 3 | 3 | 2 | 4 | 3 | 3 | 3 |
| Hornblende | 1 | <1 | 4 | 6 | 1 | <1 | <1 | 1 | 1 | 3 | 6 | 2 | 3 | 3 | 2 | <1 |
| Total | 42 | 39 | 42 | 44 | 32 | 43 | 52 | 45 | 46 | 48 | 46 | 45 | 52 | 42 | 39 | 36 |

*Samples analysed at Australian Mineral Development Laboratories, Adelaide.

Table 1 (continued). Locality Index (with the BMR sample numbers shown in brackets)

1. Post ?1950-52 lava flow on northwestern flank of volcano (74400057)

2. Western limb of ?1950-52 lava flow, northern flank (74400064)

3. Relatively old lava flow on southern flank (74400069)

4. Relatively old lava flow from head of clastic deposit on southern flank (74400072 - overlies 74400069)

5. Bomb on northern flank (74400078)

6. Near front of the more extensive of the 1966 lava flows on southern flank (74400066)

7. Front of ?1960-62 lava flow, southern flank of volcano (74400061)

8. Near front of 1971 lava flow (74400071)

9. 1964-65 lava flow, southern flank of volcano (74400063)

10. Front of 1972-75 lava flow (in November 1973 - 74400073)

11. Relatively old (pre-1943) lava flow near head of Saua River (74400068)

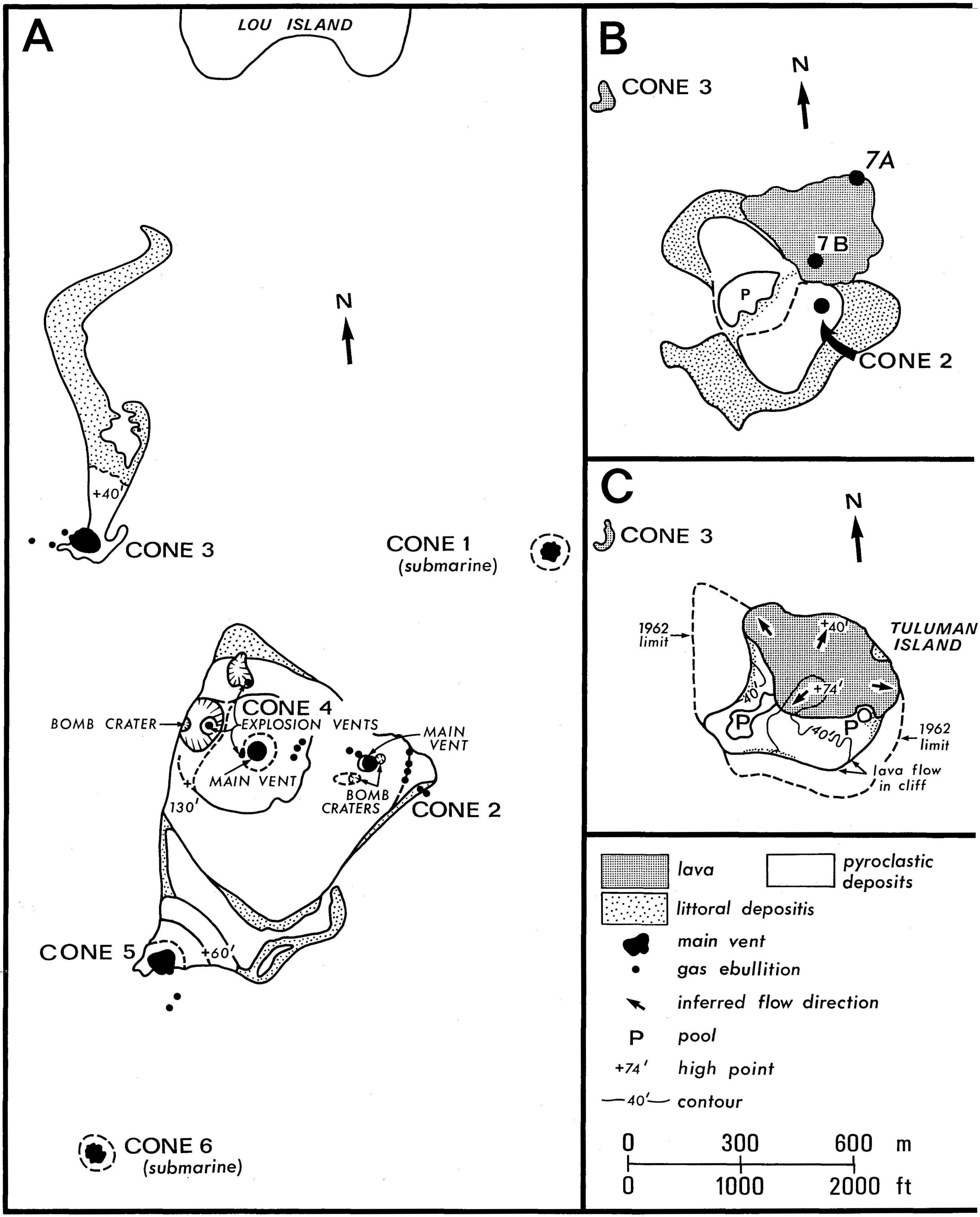
12. Thin, relatively old (pre-1943) lava flow at head of large clastic deposit on northwestern flank of volcano (74400058) - may be now covered by new lava flow observed on 26 June 1975 (see Fig. 4)

13. Pre-1943 flow, southeastern flank (74400070)

14. Pre-1943 lava flow, northwestern flank (74400077)

15. Part of large block on southeastern flank (74400065)

16. Pre-1943 lava flow, northern flank (74400060)


A
LOU ISLAND
N
+40′
CONE 3
CONE 1
(submarine)
CONE 4
BOMB CRATER
EXPLOSION VENTS
MAIN VENT
MAIN VENT
BOMB CRATERS
CONE 2
130′
CONE 5
+60′
CONE 6
(submarine)
B
CONE 3
N
7A
7B
P
CONE 2
C
CONE 3
N
1962 limit
+40′
TULUMAN ISLAND
+74′
40′
P
P
40′
1962 limit
lava flow in cliff
lava
pyroclastic deposits
littoral depositis
main vent
gas ebullition
inferred flow direction
P pool
+74′ high point
40′ contour
0 300 600 m
0 1000 2000 ft

A55/A11/21

## Appendix. Preliminary petrology

by

R.W. Johnson, C.O. McKee, & R.N. England

Chemical and modal analyses and CIPW norms of ten rocks produced during the 1974-75 eruptions of Karkar are given in Table A, where the samples are arranged from left to right in order of eruption. Samples 1 to 8, and 10, are from lava flows, and sample 9 is an ejected block. The ten rocks are believed to be representative of the most important phases of the eruption.

Table A

The rocks are low-silica andesites. They are all especially porphyritic in plagioclase, but also contain some clinopyroxene, olivine, and iron-titanium oxide phenocrysts; sample 1 contains trace amounts of orthopyroxene phenocrysts. In order of eruption, samples 1 to 8 are progressively depleted in total phenocryst and plagioclase phenocryst contents (Table A). Samples 9 and 10, however, the youngest rocks, contain at least as many phenocrysts as samples 6, 7, and 8, and more plagioclase phenocrysts than samples 7 and 8. Clinopyroxene phenocrysts are much more abundant, and olivine phenocrysts slightly more abundant in sample 9 than in the other nine samples. The groundmasses of all the samples contain glass, charged with iron-titanium oxide and clinopyroxene grains and plagioclase laths. The phenocryst and groundmass clinopyroxene is calcium-rich; no pigeonite or groundmass orthopyroxene has been identified.

Chemical analyses of older rocks from Karkar Island (Morgan, 1966, and ten unpublished analyses) suggest that low-silica andesite is the most common rock type on the volcano. In comparison, however, most of the low-silica andesites produced in 1974-75 are slightly more basic than most of

Fig A

Fig B

Fig C

the older rocks, containing less silica (Fig. A)*, and showing higher Mg/Fe values (Fig. B). Sample 9 is the most magnesian and least siliceous of all the analysed rocks from Karkar Island. On the basis of total alkali: $SiO_2$ and $K_2O:SiO_2$ relationships, the rocks of Karkar could be classified as "calcalkaline" although, as pointed out by Morgan (1966), some contain moderate amounts of total-iron similar to those of "tholeiitic" rocks.

In Figure C, the 1974 Karkar rocks are plotted in the normative pseudoquaternary system Olivine - Plagioclase - Clinopyroxene - Quartz. The four diagrams, a to d, are parts of the faces of the tetrahedron, and the samples points are projections from the interior of the tetrahedron onto each face. In Figure Ca,b,d, samples 1 to 5 fall in a cluster, but samples 6, 7 and 8, 10, and 9, form points progressively further from the Pl apex, suggesting plagioclase had a strong influence on the development of the series. This trend of depletion in normative plagioclase corresponds roughly with the depletion in plagioclase phenocrysts mentioned ~~earlier~~ above (cf. Table 1). In Figures Ca,d, the point representing sample 9 is displaced towards the Cpx apex, and in Figure Cb is displaced towards the Ol apex. These divergences correspond with the higher proportions of clinopyroxene and, to a lesser extent, olivine phenocrysts present in sample 9 (Table A).

About one-hundred electron microprobe analyses show that the mean composition of analysed plagioclase phenocrysts in all ten rocks is $An_{88}$, and that the mean plagioclase compositions of each rock fall within the range $An_{84-90}$. However, preliminary attempts to relate the change in compositions of the rocks to the separation (or addition) of different amounts of plagioclase with an average composition of $An_{88}$ have not been

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* The values plotted in Figures A, B, and C, are those obtained after standardising the oxidation state of iron by the method of Irvine and Baragar (1971 - i.e. if $Fe_2O_3 > TiO_2+1.5$, sufficient $Fe_2O_3$ is converted to FeO so that $Fe_2O_3 = TiO_2+1.5$) and recalculating volatile-free to 100 percent. Sample 9 is the most oxidised of the 1974-75 rocks.

wholly successful, although there is little doubt that the evolution of the series has been controlled to a large extent by calcic plagioclase (reasons for the discrepancies will be discussed elsewhere).

The available data do not establish which of the ten rocks, if any, is parental to the series. If the series represents a liquid line of descent from a parental composition represented by one of the older samples 1-5, the later part of the series evolved mainly by the successive separation of plagioclase, producing the compositions represented by samples 6, 7, 8, and 10. Alternatively, the series may have evolved from an alumina-poor parent (such as represented perhaps by sample 10) by the addition of plagioclase, and the magmas were erupted in reverse order. Trends from compositions intermediate along the series, or from compositions not represented by any of the analysed rocks, are also possible. In any of these trends, sample 9 - the only rock not collected from a lava flow - is anomalous because, compared to the other samples, clinopyroxene and olivine, as well as plagioclase, appear to have influenced its position in the series.

References

IRVINE, T.N., & BARAGAR, W.R.A., 1971: A guide to the chemical classification of common volcanic rocks. Can. J. Earth Sci., 8, pp. 523-548

MORGAN, W.R., 1966: A note on the petrology of some lava types from east New Guinea. J. geol. Soc. Aust., 13, pp. 583-591

Figure captions

Figure A. $Na_2O + K_2O$ *versus* $SiO_2$ diagram

Figure B. MgO *versus* $FeO + Fe_2O_3$ diagram

Figure C. Normative pseudo-quaternary system Olivine (Ol) - Plagioclase (Pl) - Clinopyroxene (Cpx) - Quartz (Qtz).

Table A. Chemical analyses,[a] CIPW norms, and modal analyses of ten rocks produced during the 1974-75 ~~during 1974~~ eruptions of Karkar volcano

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 |
|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 53.8 | 53.3 | 54.0 | 53.5 | 53.2 | 54.0 | 54.6 | 54.6 | 53.0 | 54.69 |
| $TiO_2$ | 0.56 | 0.71 | 0.84 | 0.57 | 0.88 | 1.11 | 0.80 | 0.99 | 0.75 | 0.63 |
| $Al_2O_3$ | 19.6 | 19.8 | 19.2 | 19.5 | 19.6 | 17.9 | 16.3 | 16.1 | 15.7 | 15.30 |
| $Fe_2O_3$ | 2.45 | 2.60 | 2.70 | 2.85 | 2.45 | 2.80 | 3.75 | 3.40 | 4.15 | 3.31 |
| FeO | 6.30 | 6.00 | 6.30 | 5.95 | 6.15 | 6.95 | 7.05 | 7.60 | 6.15 | 8.15 |
| MnO | 0.16 | 0.15 | 0.16 | 0.16 | 0.15 | 0.17 | 0.19 | 0.20 | 0.19 | 0.22 |
| MgO | 3.15 | 3.10 | 3.20 | 3.10 | 3.15 | 3.60 | 4.00 | 4.05 | 6.15 | 4.15 |
| CaO | 10.3 | 10.6 | 10.2 | 10.6 | 10.7 | 9.90 | 9.25 | 9.00 | 10.4 | 8.66 |
| $Na_2O$ | 2.50 | 2.50 | 2.55 | 2.50 | 2.50 | 2.65 | 2.70 | 2.75 | 2.20 | 2.97 |
| $K_2O$ | 0.88 | 0.87 | 0.90 | 0.88 | 0.85 | 0.94 | 1.01 | 1.02 | 0.79 | 1.04 |
| $P_2O_5$ | 0.16 | 0.15 | 0.17 | 0.16 | 0.16 | 0.17 | 0.20 | 0.19 | 0.14 | 0.20 |
| $H_2O+$ | <0.01 | <0.01 | <0.01 | <0.01 | <0.01 | 0.01 | <0.01 | 0.01 | 0.02 | 0.06 |
| $H_2O-$ | 0.06 | 0.01 | 0.01 | 0.06 | 0.05 | 0.03 | 0.05 | 0.07 | 0.18 | 0.02 |
| $CO_2$ | 0.05 | 0.05 | <0.05 | 0.05 | 0.05 | 0.05 | 0.05 | 0.05 | 0.05 | 0.05 |
| TOTAL | 99.97 | 99.84 | 100.23 | 99.88 | 99.89 | 100.28 | 99.95 | 100.03 | 99.87 | 99.44 |
| CIPW norms | | | | | | | | | | |
| Q | 7.85 | 7.44 | 8.24 | 7.82 | 7.30 | 7.88 | 9.15 | 8.71 | 7.39 | 7.63 |
| or | 5.20 | 5.15 | 5.31 | 5.21 | 5.03 | 5.54 | 5.97 | 6.03 | 4.68 | 6.18 |
| ab | 21.16 | 21.18 | 21.52 | 21.18 | 21.18 | 22.36 | 22.86 | 23.27 | 18.67 | 25.28 |
| an | 39.70 | 40.31 | 38.20 | 39.46 | 39.81 | 34.09 | 29.41 | 28.59 | 30.73 | 25.51 |
| di wo | 4.21 | 4.62 | 4.66 | 4.95 | 5.00 | 5.62 | 6.22 | 6.06 | 8.26 | 6.72 |
| di en | 1.95 | 2.23 | 2.25 | 2.40 | 2.42 | 2.76 | 3.20 | 3.00 | 5.30 | 3.13 |
| di fs | 2.22 | 2.31 | 2.33 | 2.46 | 2.50 | 2.75 | 2.86 | 2.94 | 2.42 | 3.51 |
| hy en | 5.90 | 5.50 | 5.69 | 5.33 | 5.43 | 6.18 | 6.77 | 7.09 | 10.06 | 7.27 |
| hy fs | 6.71 | 5.68 | 5.90 | 5.48 | 5.61 | 6.15 | 6.03 | 6.95 | 4.59 | 8.16 |
| mt | 3.56 | 3.78 | 3.91 | 4.14 | 3.56 | 4.05 | 5.44 | 4.93 | 6.04 | 4.83 |
| il | 1.06 | 1.35 | 1.59 | 1.08 | 1.67 | 2.10 | 1.52 | 1.88 | 1.43 | 1.20 |
| ap | 0.38 | 0.36 | 0.40 | 0.38 | 0.38 | 0.40 | 0.47 | 0.45 | 0.33 | 0.48 |
| cc | 0.11 | 0.11 | – | 0.11 | 0.11 | 0.11 | 0.11 | 0.11 | 0.11 | 0.11 |
| Volume percent phenocrysts[b] | | | | | | | | | | |
| Plagioclase | 37 | 34 | 33.5 | 31.5 | 29.5 | 19.5 | 12 | 9 | 14.5 | 18.5 |
| Olivine | 0.5 | 0.5 | <0.5 | 0.5 | 1 | 0.5 | <0.5 | 1 | 2 | 0.5 |
| Orthopyroxene | <0.5 | – | – | – | – | – | – | – | – | – |
| Clinopyroxene | 1.5 | 1.5 | 0.5 | 1 | 1.5 | 0.5 | 0.5 | 1 | 10 | 1.5 |
| Fe-Ti oxides | <0.5 | 0.5 | <0.5 | <0.5 | <0.5 | 0.5 | <0.5 | <0.5 | <0.5 | <0.5 |
| TOTAL | 39 | 36.5 | 34 | 33 | 32 | 21 | 12.5 | 11 | 26.5 | 20.5 |

Table A. (Cont'd)

Rock type and locality description

1. Lava flow front, mid-late February, 1974, east base of Bagiai (74710001)[c].

2. Lava flow front, early March, 1974, east base of Bagiai (74710002).

3. Lava flow front, late March, 1974, northern part of caldera, about 1 km from vent (74710003).

4. Lava flow edge, early April, 1974, east base of Bagiai (74710004).

5. Lava flow front, late May, 1974, southern part of caldera, about 300 m from vent (74710005).

6. Lava flow front, early June, 1974, southern part of caldera, about 1.2 km from vent (74710006).

7. Lava flow front, late July, 1974, southern part of caldera, about 1.1 km from vent (74710008).

8. Lava flow front, late July - early August, 1974, southwestern part of caldera, about 300 m from vent (74710009).

9. Ejected block, late July - early August, 1974, base of Bagiai (74710010).

10. Lava flow front, mid-January, 1975, west base of Bagiai (75710015).

a. Supplied by Australian Mineral Development Laboratories, Adelaide

b. Determined by C.O. McKee

c. Bureau of Mineral Resources registered number

Table 1. Analysis of gas condensate collected on Bagiai cone 22 April, 1972*.

| Temperature | ph | $SO_4$ | Cl | Ca | Mg | Na | K | Li |
|---|---|---|---|---|---|---|---|---|
| | | (in ppm) | | | | | | |
| 85°C | 5.1 | 37 | 3 | 1.0 | 0.28 | 5.04 | 0.68 | trace |

*This analysis is included in Report 73/19 (unpublished) of the Geological Survey of Papua New Guinea entitled "Gas condensate ~~XXXXXXXX~~ collection programme - progress report October, 1973", by I.H. Crick.

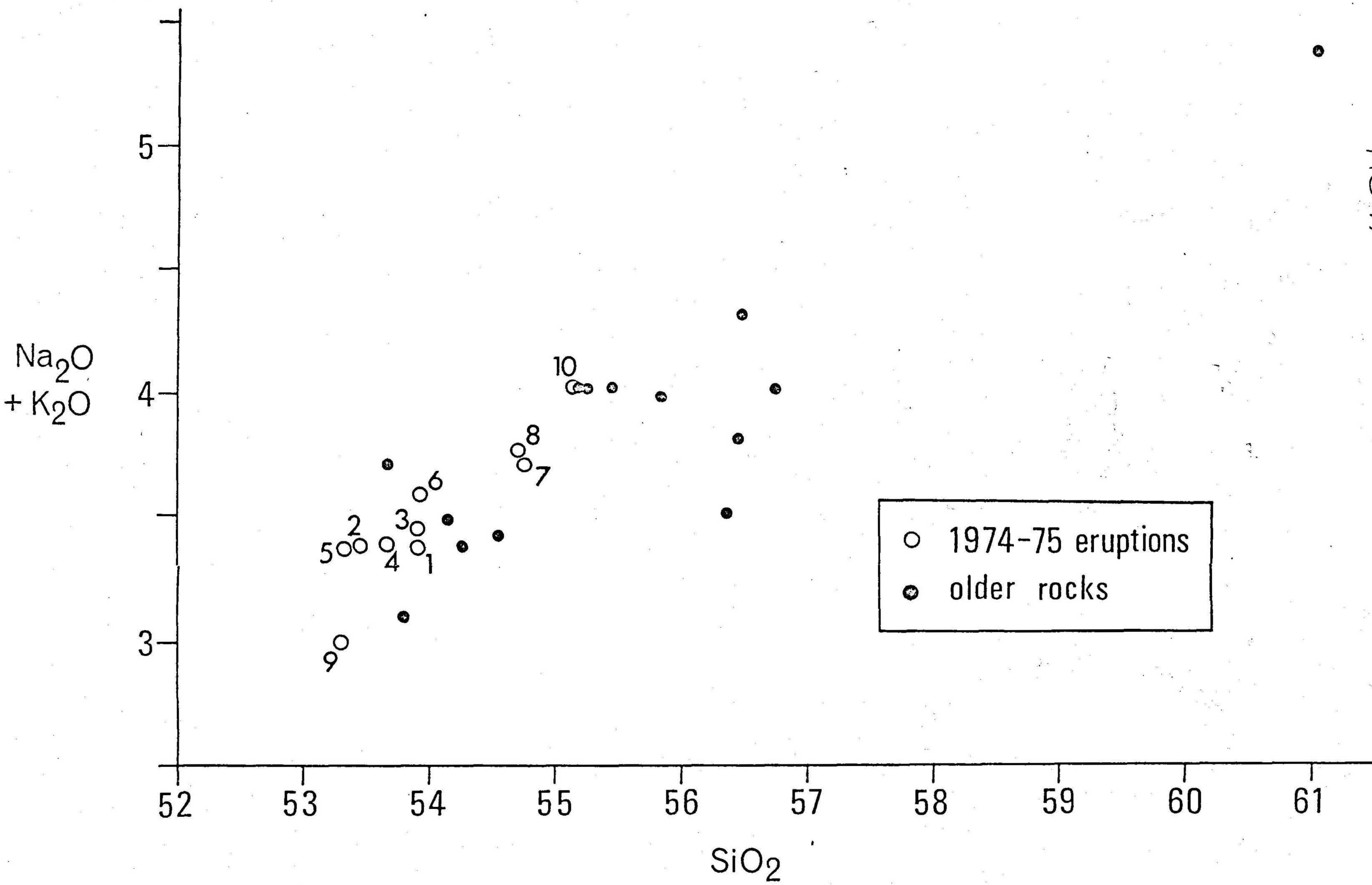
$Na_2O$
$+ K_2O$
5
4
3
52
53
54
55
56
57
58
59
60
61
$SiO_2$
1974-75 eruptions
older rocks
1
2
3
4
5
6
7
8
9
10

FIG A

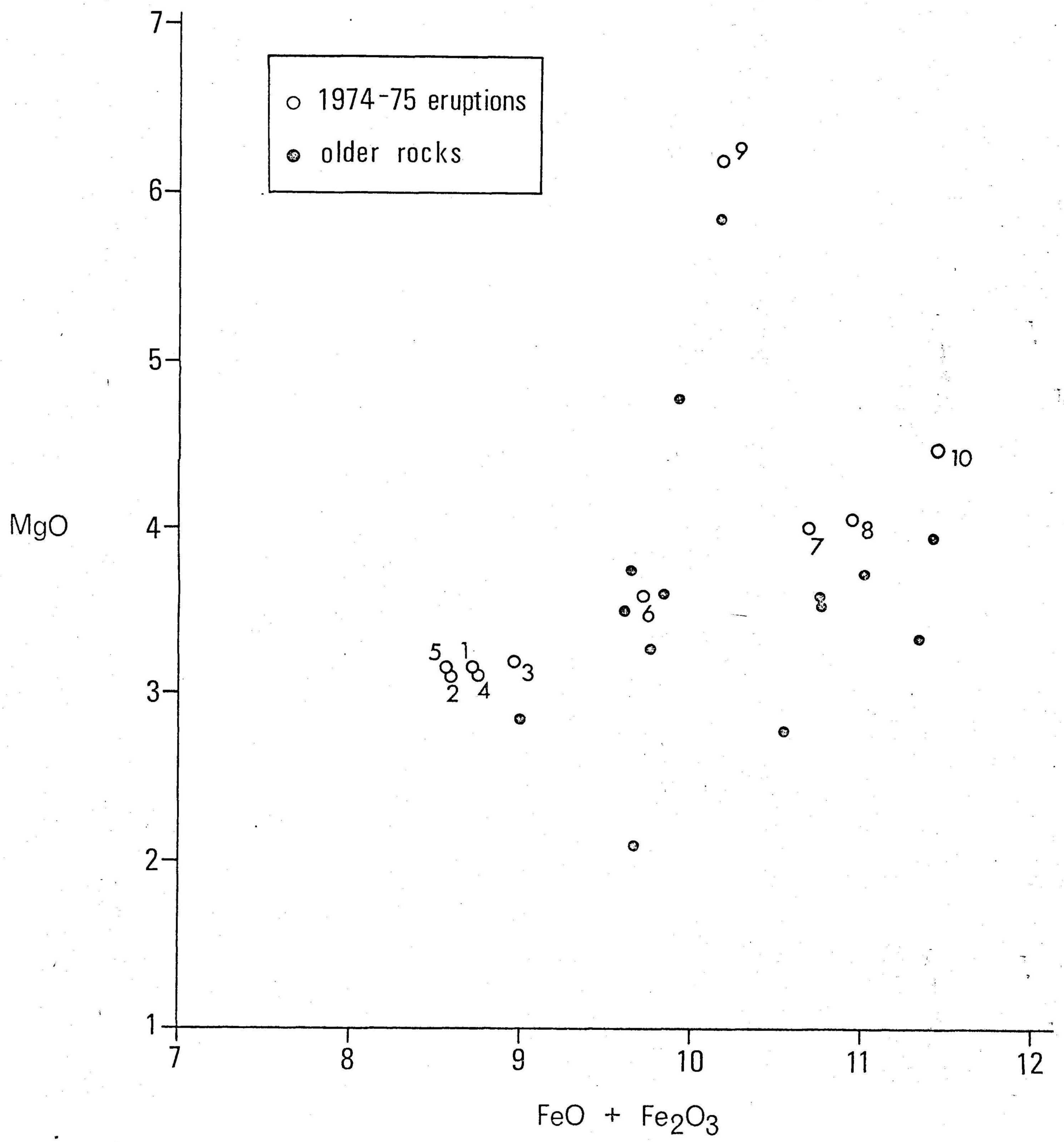
1974-75 eruptions
older rocks
MgO
FeO + Fe2O3
1
2
3
4
5
6
7
8
9
10
11
12

Fig. 13

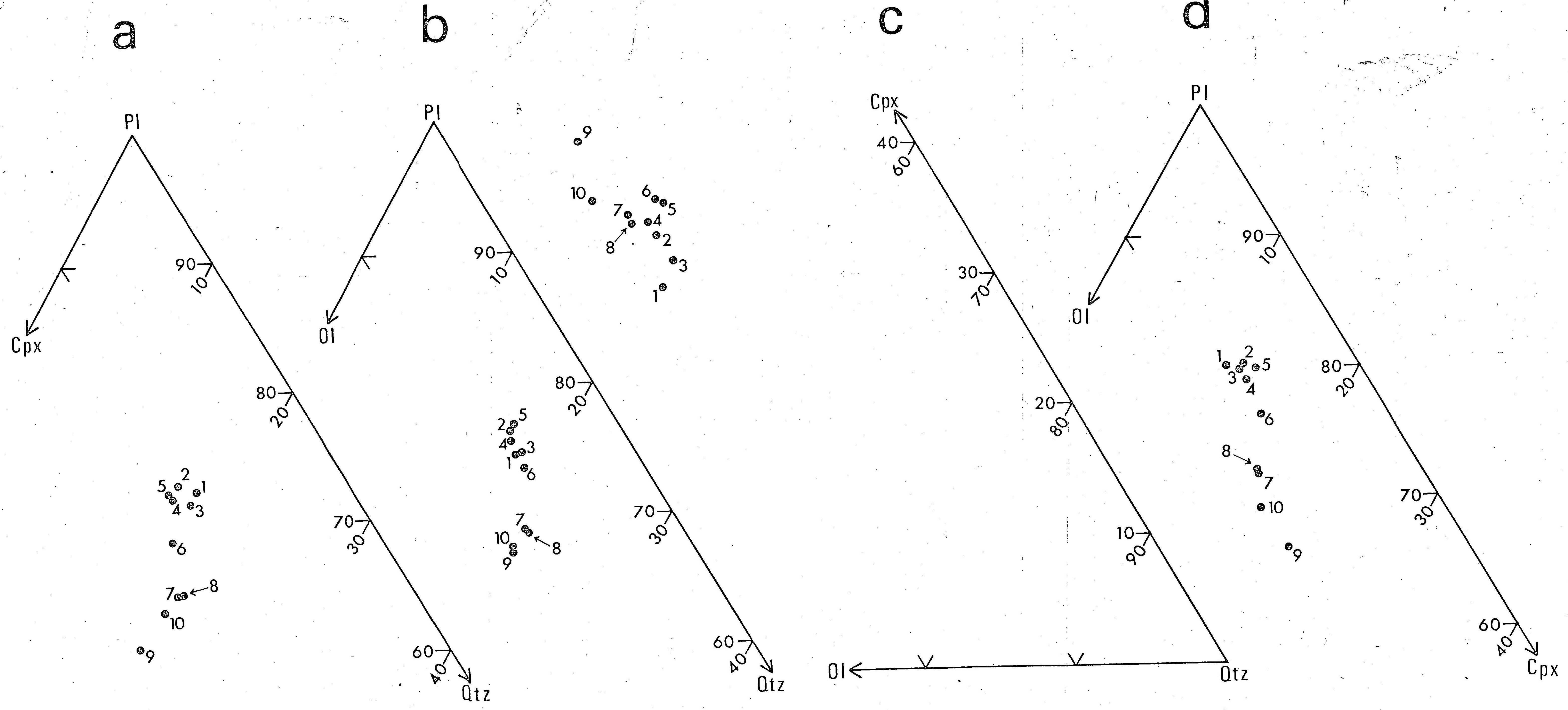


Fig. C