THE GEOLOGY AND TECTONICS OF THE BACAN REGION, EASTERN INDONESIA

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We shall not cease from exploration
And the end of all our exploring
Will be to arrive where we started
And know the place for the first time.

Four Quartets, Little Gidding (T.S. Elliot, 1942)
ABSTRACT

Bacan is located in the zone of convergence between the Australian, Philippine Sea and Eurasian Plates. This study has established a new stratigraphy for Bacan, based on geological mapping, aerial photography, new biostratigraphic dating, K-Ar and Ar-Ar dating, petrography, mineral and whole rock geochemistry.

The oldest rocks in Bacan belong to the Sibela Continental Suite and are of probable Palaeozoic age. The complex includes continental phyllites, schists and gneisses of upper amphibolite-lower granulite facies, with fabrics typical of dynamo-thermal metamorphism. The protoliths are mostly arc-related pelites, derived from the Australian craton and deposited on an active margin. Isotopic dating yields extremely young ages due to interaction with hydrothermal fluids.

Juxtaposed against the continental rocks is the mostly unmetamorphosed, arc-related Sibela Ophiolite. It is metamorphosed locally along shear zones to upper amphibolite facies. Isotopic dating yielded a Cretaceous age with an Oligocene-Miocene overprint. This ophiolite was probably derived from the Philippine Sea Plate.

The Saleh Complex consists of metabasites and foliated metasedimentary rocks. Arc-related calc-alkaline metabasites have suffered lower greenschist facies metamorphism (~250-360°C, ~4 kb) and may be cogenetic with the Bacan and South Bacan Formations. The metasedimentary rocks may be related to the Sibela Continental Suite.

In north Bacan, the oldest formation is the Upper Eocene Bacan Formation which comprises interbedded basic-intermediate arc volcanic and turbiditic volcaniclastic rocks, metamorphosed under conditions transitional between the prehnite-pumpellyite and pumpellyite-actinolite facies (250-330°C, ~2 kb). The oldest rocks in south Bacan belong to the Lower Miocene South Bacan Formation; an interbedded arc volcanic and turbidite sequence, metamorphosed to prehnite-pumpellyite facies (~240-330°C, ~2 kb). The Oligocene Tawali Formation is the oldest formation on Kasiruta, consisting of arc lavas and volcaniclastic turbidites, metamorphosed to zeolite facies (~180°C, <2 kb). The Bacan, South Bacan and Tawali Formations represent different parts of an arc, active from Late Eocene until Early Miocene, resulting from northward subduction of the Indo-Australian Plate under the Philippine Sea Plate.

There is a major Early Miocene unconformity, due to the collision of Australian continent with the Philippine Sea Plate, above which the Lower-Middle Miocene Ruta Limestone was deposited.
These shallow marine limestones include four microfacies: open platform, platform build up (patch reef), tidal bar and foreslope talus. The Ruta Formation was locally interrupted by sudden influxes of volcaniclastic material, forming the Amasing Formation, which contains three facies: shallow marine, shoal or estuarine, and beach deposits.

The Upper Miocene-Pleistocene Kaputusan Formation, composed of the Goro-goro, Pacitak and Mandioli Members, rests unconformably above the Ruta and Amasing Formations. The Goro-goro Member consists of arc andesites, with magma diversification achieved mainly by fractionation. It includes four eruption centres, erupting from Late Miocene to Pleistocene, with the oldest in south and the youngest in north Bacan. The Upper Miocene-Upper Pliocene Pacitak Member consists of reworked pyroclastic and volcaniclastic material, deposited in a shallow marine environment. The Upper Miocene-Lower Pliocene Mandioli Member consists of limestones which formed fringing coastal reefs. This formation was produced by eastward subduction of the Molucca Sea Plate. Quaternary arc basalts are related to movement along the Sorong Fault Zone.
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CHAPTER ONE
INTRODUCTION: Background, Aims, Methods and Terminology

1.1 GEOGRAPHIC SETTING
The Republic of Indonesia consists of 18,508 islands distributed in an area ~5200 km in length. It stretches between 95°E - 141°E longitude and 6°N - 12°S latitude (Fig. 1.1). Bacan is one of these islands situated in NE Indonesia, south of the Philippines and north of Australia. Politically, Bacan is part of Maluku province and the North Maluku administrative area (kabupaten) which includes the districts (kecamatans) of Bacan, Gane Barat and Kayoa.

1.1.1 Location, Settlement and Culture
Bacan and the surrounding islands of Kasiruta, Obit, Mandioli, Muari, Lata-lata, Saleh Lamo and Saleh Kecil, are located between 127°E-128°E longitude and 0°10’S-1°S latitude and constitute an area of ~3600 km². This area is referred to in this thesis as the Bacan Region and is approximately the same size as Kent.

The island group is sparsely populated, with most of the inhabitants living on the coast. Most of the villagers depend on the slash-burn agriculture. The once powerful Kingdom of Bacan, with whom Magellan’s expedition traded (Roditi, 1972; Muller, 1990), still survives but the native Bacanese are now the minority, living only around the town of Labuha.

Islam is the dominant religion in the area. It is, however, not uncommon to find a village with both Moslems and Christians co-existing. Although religion plays a major part in the lives of villagers, animistic beliefs are often accommodated in the religion. Natural forces such as the sea, mountains, and rocks are often considered sacred. Local folklore tells of a King of Bacan moving an island south of Kasiruta to the Amasing River; an indigenous concept of plate tectonics.

1.1.2 Physiography, Morphology and Relief
Physiographically the Bacan Region can be divided into eight principal areas: South Bacan, the Sibela Mountains, North Bacan, Kasiruta, Obit, Mandioli and the Saleh Isles (Fig. 1.2).

Both South and North Bacan have rugged volcanic topography with steep ridges and valleys. Typically the ridges reach ~500 m. Rivers generally have successive water falls which, although beautiful to look at, present problems for fieldwork. On the coast steep beaches are common, often dropping beyond sight within 5 metres of the shore. This is apparent where the coast line is straight and fault-controlled.
Figure 1.1. Geography of the Bacan Region, eastern Indonesia.
The central part of Bacan includes the very steep Sibela Mountains (gradient ~33%) whose summit is the highest point in the region (2111 m) and almost impossible to climb. Kasiruta is characterised by high plateaus of volcanic rocks with some karstic topography in the northern part of the island. Mandioli Island has low karstic topography, while the Saleh Isles form low-lying smooth hills.

To the north and east of Bacan is the Maluku Sea and to the south is the Obi Strait. Around Bacan are some very deep basins (Hamilton, 1974; Mammersickx et al., 1976) including the >4000 m deep Bacan Basin, the Bacan Strait (492 m), the Patinti Strait (1120 m) and Wayaua Bay (1033 m).

1.1.3 Vegetation and Climate
The Bacan Region is covered mostly by primary tropical rain forest. In low-lying coastal areas there are coconut, cocoa, cassava and clove plantations. Logging of the rain forest occurs only in the flatter parts of the region, such as Mandioli, southern Kasiruta and along the flood-plains of major rivers in North and South Bacan. Secondary rain forest reclaims the logged areas after about 2 years.

Cloves (Eugenia aromatica) and durian (Durian bethinus) are well known (Muller, 1990) and common animals include deer, wild boars, black macaques (imported from Sulawesi, by the Sultan of Bacan, in exchange for clove trees), iguana-like reptiles, numerous species of birds, snakes and allegedly crocodiles.

The climate is typical of a tropical region, with temperatures between 26°C-35°C and high humidity. There are only two seasons, wet and dry, the former between April and August, while the latter is between September and March. Field work during the wet season is not advisable due to the common flooding of rivers.

1.1.4 Access
Documentation was arranged through GRDC (Geological Research and Development Centre) in Bandung, with further permits obtained in Ambon, Ternate and Labuha. There are frequent domestic flights from Jakarta and Bandung to Ambon, the capital of Maluku. From Ambon there is a daily flight to Ternate and a biweekly flight to Labuha, the capital of Bacan. Daily passenger
Figure 1.2 Physiography and simplified topography of the Bacan Region.
boats leave Ternate's Bastiong Harbour to Babang where there are many buses connecting it with Labuha by the only road in Bacan. Food and fuel can be bought in Labuha, whereas equipment must be transported from London, Bandung or Ambon.

Geological investigations utilised exposures along rivers, coasts and occasionally logging tracks. River traverses may be difficult or impossible in some areas due to ruggedness of terrain, river depth, and availability of logistical support. Traverses started from a village, where porters were hired to help carry provisions and equipment, and normally lasted for 3 or 4 days.

The intricate shape of Bacan provides an excellent opportunity for coastal work. A dug-out canoe with a 25hp outboard engine was hired in Labuha for transportation between villages and for coastal work. This work was sometimes hampered by the inability of the canoe to land, mostly because of rough seas, waves breaking against cliffs or because of engine problems.

1.2 RESEARCH OBJECTIVES AND THESIS PLAN
This multidisciplinary study is concerned with the history of the zone of convergence between the Australian, Philippine Sea and Eurasian Plates. The Bacan Region includes rocks of Australian (Hamilton, 1979; Morris et al., 1983; Hall et al., 1988a) and Philippine Sea Plate origin and is therefore an ideal area to study the processes involved in the tectonic development of a complex boundary.

The geology and tectonic development of Bacan will contribute to an understanding of the tectonic development of eastern Indonesia and may provide a modern analogue of older complex tectonic areas, such as the Western Cordillera (e.g. Jones et al., 1982; Silver & Smith, 1983) and the Aegean (e.g. Robertson & Dixon, 1984).

To this end, the aims of the project were:
[1] To identify and map lithologic units and establish a stratigraphy for the region.
[2] To interpret the depositional setting of the units, sedimentologically and/or chemically.
[3] To assess the timing of deposition and hiatuses between units, and analyse their significance in a regional tectonic context.
[4] To reconstruct the tectonic development of the region, including identifying the different tectonostratigraphic affinities and the timing of their amalgamation.

This thesis will be divided into three parts: part one (Chapters One and Two) introduces the region
and reviews previous geological research in the region; part two (*Chapters Three to Ten*) describes the stratigraphy, lithology, petrology and geochemistry, and is followed by an interpretation of the depositional environment, metamorphic history, tectonic setting and significance of each formation, starting with the oldest rocks and ending with the youngest; part three (*Chapter Eleven*) proposes a tectonic reconstruction of the Bacan Region and assesses its implications for regional and global tectonics.

**1.3 METHODS**

**1.3.1 Field Work**

Geological mapping was carried out during October-November 1989 and July-October 1990, as part of the University of London Sorong Fault Zone Project in co-operation with GRDC. Lithostratigraphic formations were identified during this fieldwork.

Aerial photographic enlargements were used for location in the forest, and 1:250,000 and 1:100,000 topographic maps were used for navigational purposes and coastal location. Precise location is often difficult due to heavy vegetation, changes in river directions and unreliable topographic maps. This problem was minimised by frequent use of river bearings and correlation between these, aerial photographs and maps.

Additional geological data has been provided by other members of the Sorong Fault Zone Project team who worked on previous and concurrent surveys. Fig. 1.3 shows the locations of traverses undertaken in this study.

**1.3.2 Laboratory Work**

Photogeological interpretation was used to map the distribution of formations, based on the lithology, age determination and aerial photographic characteristics of units, constrained by field data. Aerial photographic interpretation followed the guidance of Perry *et al.* (1979), Pandey (1987) and Drury (1987).

Approximately 1200 samples were collected and nearly all have been sectioned and analysed petrographically. The specific aim of the petrographic study was to identify and classify the samples, texturally and mineralogically, and to characterise the formations.

Based on the petrographic work, 78 samples were selected for microprobe analysis for mineral and glass chemistry determination. Samples were prepared as polished thin sections coated with
graphite. Microprobe analyses were made on a JEOL 733 Superprobe Electron Microprobe at Birkbeck College, London, coupled with a Link Analytical Energy Dispersive System AN 10000/55S and LEMAS stage control. An accelerating potential of 15 kEv beam and a probe current of 10 Na were used, with a minimum beam spot size of 1.0 μm. Elements routinely analysed were SiO$_2$, TiO$_2$, Al$_2$O$_3$, FeO, Fe$_2$O$_3$, MnO, NiO, CrO$_2$, MgO, CaO, K$_2$O and Na$_2$O. Natural silicates and pure metal were used as standards, and during analyses cobalt was used as the primary reference standard. Correction procedures were carried out using the ZAF correction programme supplied by Link (iterative spectrum stripping technique). Stoichiometric calculation was achieved using Pascal programmes developed by Dr Robert Hall. The results of analyses are presented in Appendix A; these are grouped into formations.

A selection of samples from different formations were chemically analysed using the X-ray Fluorescence technique (XRF) for major and trace elements. Samples were chosen on the basis of freshness and any alteration present was avoided as far as possible. Rock crushing used a hardened steel jaw crusher followed by WC Tema mill at Birkbeck College, London. Major element analyses used fused glass discs. Trace element analyses were performed on pressed powder pellets. Disc and pellet preparation followed the standard procedures of RHBNC, London. Analyses were performed using a Philips PW1400 X-Ray Spectrometer at RHBNC. Tungsten anode tube was used for major elements and Sc, La, Ba, V, Nd, Ce, Cr, Ni, Cu and Zn trace elements. Rhodium tube was used for Pb, Th, Rb, Sr, Y, Zr, Nb, Cl and Ga. Details of instrumental conditions are given in Thirlwall & Marriner (1986), from which the 2σ errors were calculated to be: ± 1 ppm for most elements, ± 0.3 ppm for Nb, ± 0.2 ppm for Rb, ± 0.6 ppm for Y, ± 1.3 ppm for La and Ce, ± 2.0 ppm for Cu, V and Zn, ± 3.0 ppm for Ba, ± 6.0 ppm for Cr and ± 12 ppm for Cl. Data was processed using the VAX-VMS computer system. Appendix B shows the result of XRF analyses and includes the calculation for 2σ errors based on the triplicate analyses on sample SM10.

Some samples were analysed for silica using the method of Shapiro & Brannock (1962) and other major and trace elements using the Induced Coupled Plasma-arc (ICP) technique. Samples were fused in lithium metaborate and the resultant glass dissolved in HCl; the solution reacted with ammonium molybdate and then 4-amino 3-hydroxynaphthlene-1 sulphuric acid. The intensity of colour of the resultant blue solution was measured, at 650 nm wave length, on a Ultraviolet Visible spectrophotometer and the SiO$_2$ value calibrated against standard rocks from National Institute of Mining, South Africa. For the other major, trace and REE elements, the sample was dissolved in heated perchloric and hydrofluoric acids and then measured on ICP spectrometer. The
Fig. 1.3 Localiton of river traverses and logging tracks on Bacan. The entire coast has been investigated during this project.
results were calibrated against standard rocks (Walsh, 1980; Walsh et al., 1981). The rubidium content was determined by emission spectrophotometry relative to a standard Rb solution. The zirconium content was determined by treating the sample with HCl or HSO₄ followed by repeated HSO₄ evaporations until dissolution. The solution is then reacted with xylenol orange and the resulting red colour is measured at 525 nm on a UV/VIS spectrometer, calibrated against a standard. Gravimetric methods were employed for CO₂ and H₂O analyses. The sample was heated in a 1000°C tube furnace, and the evolved gas swept by a stream of nitrogen across a tube containing magnesium perchlorate and soda lime. Increasing weight of these tubes correspond to the H₂O⁺ and CO₂ content of the sample respectively. The results of these analyses are presented in Appendix C.

Sample selection procedure for K-Ar and Ar-Ar isotopic dating was based on their petrography and the size of samples. A detailed description of methods, instrumental calibration, sample descriptions and discussion of isotopic age results is given in Appendix D. There are mineral and whole rock chemical analyses of all these samples.

Lithofacies analyses of deep-marine clastic sedimentary rocks follow the model of Pickering et al. (1989), whereas those of carbonate rocks follow Flügel (1982).

1.4 TERMINOLOGY AND NOMENCLATURE

[1] Sample names. All samples collected were identified by one or two letters followed by a number and occasionally another letter. The first letter identifies the sample location: B for Bacan, Kasiruta and Mandioli; S for Saleh Islands. The second letter identifies the geologist who collected the samples: A for Prof M.Audley-Charles (1984), G for Dr G.J.Nichols (1990), J for Dr J.Ali (1989 and 1990), L for Dr L.Garvie (1987), M for J.F.A Malaihollo (1989 and 1990), P for Dr P.Ballantyne (1987), R for Dr R.Hall (1987 and 1989), S for Dr S.J.Roberts (1990) and T for Dr T.Charlton (1989). The only exceptions are samples collected by Dr Hall in 1984 which do not have a second letter. The numbers were assigned in chronological order, and any letter following the number denotes variations of samples collected from a single outcrop/locality. An example is: BT36A = Sample A collected by Dr T.Charlton in 1989 from the thirty-sixth locality in Bacan.

[2] Geographic names. All localities are reported using their Indonesian names

<table>
<thead>
<tr>
<th>Code</th>
<th>Name</th>
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<tr>
<td>P</td>
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Introduction

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Structure. All planar surface attitudes are reported in a form of dip direction/dip angle.
CHAPTER TWO
REVIEW OF PREVIOUS WORK, REGIONAL GEOLOGY AND TECTONIC FRAMEWORK

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2.1.1 The Early Dutch Expeditions and Work during WWII
2.1.2 Recent Indonesian Work
2.1.3 Geochemical Surveys
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2.2 REGIONAL GEOLOGY AND TECTONIC FRAMEWORK
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2.3 SUMMARY
CHAPTER TWO
REVIEW OF PREVIOUS WORK, REGIONAL GEOLOGY AND TECTONIC FRAMEWORK

2.1 REVIEW OF PREVIOUS WORK IN THE BACAN REGION
Perhaps due to its inaccessible location, studies of the Bacan region are scarce. The earliest work in the region was undertaken by Dutch geologists working as part of the Netherlands East Indies (as Indonesia was known) Geological Survey. During WWII, Japanese and American workers studied the area in connection with the war effort. Subsequent publications by Dutch geologists are based on these early studies. Indonesian geologists were the first to systematically map and study the region for its mineral resources. Later studies have included investigations of geophysics, geochemistry and the tectonic evolution of the region. Work by earth scientists from the University of London and the Geological Research and Development Centre in Bandung have constituted the most recent comprehensive studies in the region. This chapter will review all this research and outline a generalised tectonic framework of the Bacan region.

2.1.1 The Early Dutch Expeditions and Work during WWII
The earliest known investigation on Bacan was a study by Retgers in 1895. Brouwer (1923) mentioned this work when describing the metamorphic rocks of the Sibela Mountains. A study by Wichmann in 1899 was also mentioned by Brouwer (1923) in his review of the geology of Bacan. These two publications were unobtainable during this literature review. The earliest work obtained is by Martin (1903), who described the presence of Early Tertiary limestones containing foraminifera from Bacan and Obi.

A study by Bucking in 1911 was referred to by Rutten (1923), who attributed him with the description of the lithologies of Bacan. Verbeek in 1908 wrote a treatise on the geology of the Moluccas based on his reconnaissance work in 1899. Brouwer (1923) credited Verbeek with the first geological map of Bacan. This map, modified by Brouwer, is reproduced as Fig.2.1a.

The first major investigation of Bacan was published by Brouwer (1923) based on his field work in 1915. In this work he identified the lithologies of Bacan as schists and eruptive rocks, overlain by andesite and basalts of Tertiary and Quaternary ages. Based on the lithological similarities between the southern and northern Bacan and their differences with the Sibela Mountain, he also speculated that Bacan was composed of three different islands which have been united in "relatively short age". He attributed the presence of elevated coral reefs and drowned trees to the recent uplift and subsidence of parts of Bacan. Rutten (1923) wrote a summary of the geology of
Review of Previous Work, Regional Geology and Tectonic Framework

the Dutch East Indies which includes Bacan. In this review he remarked on the occurrence of the Sibela metamorphic rocks, the coal-bearing Neogene rocks in the Amasing River and the widespread presence of volcanic rocks in Bacan.

During World War II, when the Moluccas was occupied by the Japanese forces, Bessho (1944) published a study on the mineral exploration and reconnaissance geological surveys in the Halmahera Region. In 1944 the Military Geology Unit of the U.S. Geological Survey produced a geological sketch map of Halmahera as part of a "terrain intelligence study". These two work, were unobtainable during this review.

Van Bemmelen (1949) published a review of the work of the Geological Survey of the Netherlands Indies from its inception in 1850, with additional unpublished data from oil companies. He proposed a stratigraphy of the Halmahera Region, which included Bacan, based mainly on the work of Brouwer (1923). He considered the "pre-Jurassic" metamorphic crystalline basement complex and granites of the Sibela Mountains to be the oldest rocks in the region. He records Cretaceous tuffaceous marl, limestone and chert with contemporaneous basic, ultrabasic, abyssal and hypabyssal intrusions. The Eocene tuffaceous sandstone, marls, limestone with andesitic, dacitic tuffs and lava flows were thought to be unconformably above the Cretaceous sequence. These Eocene rocks were said to be cut by intermediate abyssal and hypabyssal intrusions. Young Palaeogene-old Neogene limestone, tuffaceous sandstone, conglomerate and breccia with occasional lignites overlie the Eocene rocks. He attributed the young Neogene volcanic and tuffaceous rocks to volcanism in the "Ternate Zone" and traced this zone from the northwest arm of Halmahera, through Bacan to the Bird's Head of Irian Jaya.

Neumann van Padang (1951) identified eight active volcanoes in the Halmahera Region. Verstappen (1960, 1964) made aerial photographic interpretations of Halmahera, in particular the geomorphological features of the west Halmahera Arc, based on fieldwork in 1956 and 1958. He recognised the main geomorphological features of the Halmahera region; the deeply dissected undulating upland of the eastern arm; the east tilted southern arm; the complex volcanic graben of the northern arm; and the greatly uplifted horst of the Sibela Mountains. He correlated the east arm of Halmahera with the islands of Gebe and Waigeo as a non-volcanic province, whereas west Halmahera, Bacan and the Bird's Head of Irian Jaya are linked as a volcanic province.
Figure 2.1 Simplified Geological Maps of the Bacan Region
2.1.2 Recent Indonesian Work

The earliest mentioned Indonesian worker in the Bacan region is Muchsin (1976) who published a report on the occurrence of geothermal fields in North Moluccas. In Yasin (1980), this work is said to connect the hot springs in Bacan with faulting. As part of the Five Year Development Programme, the Indonesian Geological Research and Development Centre (previously Geological Survey of Indonesia) conducted a series of mapping programmes on a 1:250,000 scale. Supriatna (1980), Apandi & Sudana (1980) and Yasin (1978, 1980) mapped northern Halmahera and Morotai, central Halmahera and southern Halmahera and Bacan respectively. A simplified version of Yasin’s map is reproduced as Fig.2.1b.

Following this work, Sukamto et al. (1981) divided the Halmahera region into 3 provinces: An Eastern Halmahera-Waigeo Province; a Western Halmahera-Obi Province (including Bacan), which corresponds to Van Bemmelen’s (1949) Ternate Zone Volcanism; and a Talaud-Tifore Province, a zone between the Halmahera and Sangihe Troughs. They concluded that the Eastern Halmahera Province is characterised by an ophiolitic sequence with an imbricated Late Cretaceous deep marine and Paleocene-Eocene trench and trench slope deposit. Overlying these sequences are Neogene marine clastic and carbonate rocks with compressional style tectonics. This province is correlated with the islands of Gag, Gebe and Waigeo. The Western Halmahera-Obi Province consists of Oligocene volcanic arc rocks, overlain by Neogene volcanic arc sequences. There are two types of basement assigned to this province: the regional metamorphic rocks of continental affinity, exposed in central Bacan; and the ophiolite exposed in Obi. This province is said to be dominated by extensional tectonics. The Talaud-Tifore Province is characterised by an ophiolite sequence and intensely tectonised melange, and is dominated by compressional style tectonics. The boundary between the Eastern and Western Halmahera Provinces is an intensely folded and faulted zone called the Median Tectonic Line.

Soeria Atmadja (1981) reported the results of petrological study of the rocks from the Halmahera Region. He concluded that the Miocene volcanic rocks of Western Halmahera Province of Sukamto et al. (1981) have calc-alkaline affinities, and the ophiolite of Obi shows island arc affinities. The Eastern Halmahera ophiolite represents a subduction zone ophiolite. He then referred to the Eastern and Western Halmahera Provinces as a paired metamorphic belt. Although he characterised the low-pressure, high-temperature arc rocks of Western Halmahera, he failed to describe the high-pressure, low-temperature subduction zone metamorphic terrain of Eastern Halmahera, which is an essential component of a paired belt. He is also correlating the pre-Mesozoic ophiolite of Eastern Halmahera and the Oligocene-Miocene volcanics of Western Halmahera.
Silitonga et al. (1981) summarised the results of a collaborative study by the German BGR and the Indonesian Bureau of Mineral Resources study on mineral prospects in the Bacan region. This study incorporated a photo-geological survey and established a stratigraphy of Bacan including: a metamorphic basement, older volcanic rocks, older sedimentary rocks, younger volcanic rocks, younger sedimentary rocks and the youngest volcanic rocks. Although this stratigraphy appears simple, most of the units can be correlated with the formations in this study. They also identified Bacan as broken into fault-bounded horst and graben blocks, which are associated with Quaternary volcanism and rapid uplift of Bacan. Their geological map is reproduced as Fig.2.1c.

Pudjowalujo & Bering (1982) published 44 multi-element anomaly maps from the Kaputusan area in Bacan. In this publication they reported a copper-gold anomaly discovered during the German-Indonesian work in 1977-1979. They attributed the deposit to possible en-echelon faults of the main deep seated fault extending from northern Halmahera to southern Bacan. This fault is now capped by the Quaternary volcanoes of the Halmahera Arc. Pudjowalujo & Surjono (1982) reported the same discovery, with the addition of joint analyses in the Kaputusan area, and link the presence of sulphide mineralisation with the joint development.

Hartono & Tjokrosapoetro (1984) divided the Indonesian archipelago into thirteen terranes, two of which are the Bacan and the Halmahera-northern Irian Jaya arc. They assigned the continental rocks in Bacan to the Australian Continent and the ultrabasic rocks in Bacan to the Halmahera-northern Irian Jaya terrane. The latter terrane is composed of ophiolite, Paleogene-Miocene island arc volcanics, intrusives and sedimentary rocks succeeded by Miocene limestone.

2.1.3 Geochemical Surveys
Morris et al. (1983) conducted a reconnaissance geochemical survey of the present Halmahera arc. They divided the arc into three segments: [1] the active arc north of Bacan, consisting of calc-alkaline basalts, basaltic andesites and andesites, characterised by high $\text{Al}_2\text{O}_3$, small to moderate Fe enrichment, depletion in HFS elements and Pb isotope compositions suggesting contamination by oceanic sediment; [2] largely dacitic southern Bacan volcanic rocks, characterised by high concentrations of Rb and Cs, steep alkali enrichment trends and high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, indicating interaction with continental crust; and [3] biotite-bearing lavas from extinct volcanoes, dominantly andesites with some shoshonites, along the Sorong Fault, distinguished by their high $\text{Al}_2\text{O}_3$, low $\text{TiO}_2$, high total alkalis, no Fe enrichment trend, strong Ba and Sr enrichment, and attributed to mantle-derived magmas erupted along a transform plate boundary.
Tera et al. (1986) suggested that the short half-life and the low level of $^{10}$Be in the present Halmahera Arc rocks indicates that young sediments (<10 Ma) are not being incorporated in the arc magmas. Based on Cs and B-isotope contents, Palmer (1991) shows that volcanic rocks of the present Halmahera Arc, including Bacan, have been influenced by fluids derived from dehydration and/or melting of the subducted slab.

Bering et al. (1985) produced an unpublished report on the Cu-Au porphyry of five highly tectonised areas in the Bacan region. Of these five areas, the Kaputusan area is considered to have most potential with ~70 tonnes of ore containing 0.30% of Cu and 0.18-0.23 g/ton Au. The ore is a tonalite porphyry with hydrothermal and propylitic alteration interspersed with more recent quartz-sericite alteration. Similar types of Au mineralisation led them to suggest that the island chain of northern Moluccas represents, both tectonically and metallo-genetically, a link between the metalliferous provinces of New Guinea and the Philippines. Associated with this report is a photo-geological map of Bacan prepared by Mollat, which is similar to that of Silitonga et al. (1981).

In the GEOSEA conference in 1987, Pringle (1989) presented a paper on the epithermal gold mineralisation in the Bacan region. From stream sediment samples, he was able to distinguish six main geochemical associations: Au-(Cu-Pb-Zn-Ag); (Au-Cu); Cu-(Au-Zn); (Au-Pb-Zn-Ag-As-Ni-Hg); (Au-Pb-Hg) and As-(Au-Pb-Ag). The majority of these associations are associated with Tertiary andesitic lavas and volcanioclastic rocks of the Bacan Formation.

### 2.1.4 Geophysical Surveys and Tectonic Studies

Hatherton & Dickinson (1969) published the first contoured map of earthquake hypocentres in Indonesia, and Fitch (1970, 1972) and Fitch & Molnar (1970) related earthquake focal mechanisms to regional tectonics in eastern Indonesia. These studies showed that the foci beneath Molucca Sea Plate define two Benioff zones which are dipping away from the central axis of the Molucca Sea and defining an inverted U-shaped plate dipping beneath the flanking arcs: the Sangihe Arc to the west and the Halmahera Arc to the east. These arcs are bordered by troughs up to 3 km deep at the edges of the Molucca Sea. In the centre of the Molucca Sea is a topographic high, which locally rises above the sea in the Talaud, Mayu, Mare and Tifore islands. These are sites of shallow thrust-type earthquakes (Fitch, 1972).

The Molucca Sea Collision Zone was recognised by Silver & Moore (1978) as the world's only example of an arc-arc collision. Based on seismic reflection profiles, they interpreted the
topographic high in the centre of Molucca Sea Plate as the deformed Molucca Sea Plate obducted onto the two colliding fore-arc. Due to this deformation, the trough to the west of Halmahera is not a deep, narrow bathymetric trench which is an expression of a fossil subduction zone, but an expression of arc-directed thrusting of the collision complex. Silver & Moore (1978) and Sukamto et al. (1981) indicated that the northern boundary of the Halmahera Plate is an oblique intersection of the Halmahera Trough and the Philippine Trench, implying that Halmahera is on a different plate than the Philippine Sea Plate (Fig.2.3c).

Undoubtedly the most widely quoted tectonic work is the review of the Indonesian geology and tectonics by Hamilton (1979), in which he described the Halmahera Region as being located in a particularly chaotic tectonic setting with poorly known geology. Whilst admitting this, he went on to interpret the eastern arm of Halmahera as a subduction melange and the western arm to be a magmatic arc province, whereas the granitic, gneisses and "contact-metamorphic schists" of Bacan might represent intrusive centres within the magmatic arc or a continental fragment torn from New Guinea. He assigned a Palaeozoic age to these metamorphic rocks and assumed that they were transported from northwest New Guinea along the Sorong Fault Zone. The occurrence of melange in eastern Halmahera and a volcanic arc in west Halmahera is interpreted to be a paired belt belonging to a west-dipping subduction system. The present east-dipping subduction under Halmahera represents subduction polarity reversal. The southern termination of the Philippine Trench is connected, by a thrust or strike-slip fault, to the Sorong Fault Zone defining the eastern boundary of the Halmahera Plate. The western boundary of the Halmahera Plate is the Halmahera Trough, while the Sorong Fault Zone defines its southern boundary. However, he did not define the northern boundary of the plate. His Halmahera Plate includes Halmahera, Bacan and Waigeo (Fig.2.3a).

Using the spatial distribution of earthquake hypocentres and their focal mechanism solutions, Cardwell et al. (1980) constructed the three dimensional shape of the Molucca Sea Plate (Fig.2.2) and the nature of plate boundaries in the region. They suggested that most of the deformation in the area is taken up by the Philippine Fault and subduction of the Molucca Sea Plate. They inferred that the Philippine Trench is terminated by a transform fault south of 2°N, explaining the lack of seismicity. The amount of Philippine Sea Plate lithosphere subducted under the Philippine Trench is <200 km. They also postulated that the subduction of the Philippine Sea Plate in the Halmahera Region is young, with only ~150 km of westward subducted lithosphere shown at 5°N. Cardwell et al. (1980) concluded that the northern continuation of the ~680 km subduction under the Sangihe arc is terminated when the Talaud-East Mindanao island arc collided with the Sangihe-
Figure 2.2. Three dimensional shape of the Molucca Sea (after Cardwell et al., 1980).
McCaffrey et al. (1980a) confirmed the existence of the Molucca Sea collision complex from the thick low-velocity layer revealed by seismic refraction profiles and the large negative free air gravity anomaly. The difference in depth to the basement on either side of the topographic high in the centre of the Molucca Sea Plate, suggests that this high is fault controlled, consistent with the asymmetric shape of the regional gravity anomaly over the basin. From seismic wave travel times, they concluded that the collision complex consist of kilometre scale blocks in a homogeneous matrix. The central Molucca Sea high is therefore interpreted to be an ophiolitic sequence thrust up at a steep angle. This is further supported by McCaffrey et al. (1980b) who used the micro-earthquake survey to determine that the central Molucca Sea shows scattered normal and strike-slip fault mechanisms dominating the upper 20 km of crust, while reverse fault mechanisms characterise the earthquakes below this depth. The collision complex is interpreted as an amalgamation of accretionary wedges of the two opposing arcs. Shortening is accommodated by high-angle reverse and strike-slip faulting at the basement of central Molucca Sea high, rather than by displacement between the Molucca Sea Plate and the colliding arcs (McCaffrey 1982, 1983, 1987; McCaffrey et al., 1983). McCaffrey (1982) located Halmahera within the margins of the Philippine Sea Plate by proposing a transform fault connecting the southern Philippine Trench with the Halmahera Trench (Fig.2.3d).

Following his earlier fault plane solution work, McCaffrey (1991) postulated that the inverted U-shape of the Molucca Sea Plate increases its own buoyancy, thus preventing it from sinking out of the way of the encroaching island arcs. Consequently, the central ridge of the Molucca Sea Plate is being lifted by high angle thrust faults that extend up to 15 km into the upper mantle, exposing the ophiolite sequence.

Using multichannel seismic reflection profiles, Letouzey et al. (1983) proposed the presence of NW-SE late Neogene-present compressional structures in southern Halmahera Region, several sinistral, transcurrent and reverse faults south of Halmahera, and the possibility of remnants of Molucca Sea Plate outcrops in eastern Sulawesi and Obi due to collision.

Moore & Silver (1983) proposed that the Philippine Trench is actively propagating southward from its present termination at 2°N in response to the collision between eastern and western Mindanao and the subsequent strike-slip motion. They described Talaud as the northern limit of the present
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Halmahera fore-arc and postulated that the Snellius Ridge, northeast of Halmahera, is the northward extension of the Halmahera arc. East Mindanao and Halmahera arc are connected by the strike-slip Philippine Fault, putting them in on a different plate from the Philippine Sea Plate (Fig.2.3e).

Pigram & Panggabean (1984) suggested that the Australian Continent begin to rift at ~230 Ma and sea floor spreading was initiated at ~185 Ma in Papua New Guinea and ~170 Ma in Irian Jaya. This resulted in separation of microcontinents, which are the Banggai-Sula, Buton, Northern Irian Jaya and Bacan-Obi microcontinents, from the main Australian Plate. Based on stratigraphical similarities, they attributed the Obi-Bacan microcontinent to be a rifted part of central Papua New Guinea.

2.1.5 Recent University of London Work

The most comprehensive land-based studies in the Halmahera Region have been those of scientists from the University of London and Geological Research and Development Centre in Bandung. These are the results of a programme of research, of which this work is a part, which started in 1984 and continued in 1987, 1988, 1989, 1990 and 1992. These multidisciplinary studies are concerned with the tectonic evolution of the region, and include aerial photographic, sedimentological, stratigraphical, biostratigraphical, petrological, geochemical, isotopic geochemistry, radiometric dating, palaeomagnetic and marine geology studies and gravity surveys.

The first paper published from these works is by Hall (1987), concerning the evolution of plate boundaries in the region. In this paper he proposed that Halmahera and eastern Mindanao were part of a single plate since Late Eocene-Early Oligocene, both having a Late Cretaceous-early Tertiary arc and fore-arc basement complex. Hall reported no evidence of the Late Oligocene-Miocene arc favoured by Sukamto et al. (1981). The succeeding stage of volcanism in the area is related to the eastward subduction of Molucca Sea Plate under Halmahera starting in the Pliocene. This subduction is attributed to the diachronous collision at the western edge of the Philippine Sea Plate, beginning in Mindanao in Late Miocene, which impeded the movement of the Philippine Sea Plate, causing further motion to be taken by the strike-slip motion along the Philippine Fault, subduction under the Philippine Trench and subduction of the Molucca Sea Plate beneath Halmahera. In the Pleistocene there was a period of volcanic quiescence associated with the westward shift of the arc. The present arc is built upon the deformed Pliocene arc. His plate boundary configuration is presented here as Fig.2.3f.
Figure 2.3. Maps showing the different models of the present day tectonic setting of the Halmahera region.
Following this is a series of papers describing the geology of the area (Hall et al., 1988a; 1988b). In the first paper they described the nature of the basement rocks of the Halmahera Region. Eastern Halmahera has a basement of dismembered ophiolite with slices of Mesozoic and Eocene sedimentary rocks unconformably overlain by Middle Oligocene and younger sedimentary and volcanic rocks. Stratigraphical and petrological similarities between eastern Halmahera and the Marianas fore-arc led them to interpret eastern Halmahera as a pre-Oligocene fore-arc lacking an accretionary prism. The Mesozoic and Eocene sedimentary rocks were imbricated with the ophiolite during the Late Eocene plate reorganisation event. The eastern Halmahera basement can be traced into eastern Mindanao, and is probably related to similar terranes within or around the present Philippine Sea Plate. The pre-Oligocene arc is believed to form part of the basement of western Halmahera. In contrast, the basement rocks of southern Bacan were recognised to be of continental metamorphic character with associated deformed ophiolite which is different from that of eastern Halmahera. They interpreted the basement of Bacan to be part of the Australian continental margin separated from the Halmahera basement by a splay of the Sorong Fault zone, with the ophiolite representing magmatism in the fault zone. Their geological map of Bacan is presented here as Fig.2.1d.

In the second paper Hall et al. (1988b) established a new stratigraphy and proposed a geological history of the Halmahera region. After the cessation of volcanic activity in the Eocene, the former fore-arc of eastern Halmahera was uplifted and then deeply eroded in the Palaeogene. Slow subsidence followed in the middle-late Oligocene and by the end of Miocene the area was dominated by shallow water carbonate deposition. Subsidence quickened in the Early Pliocene resulting in the deposition of volcanic debris from the Pliocene arc in western Halmahera which is built on the Early Tertiary arc. This rapid subsidence in the back-arc region was associated with the initiation of the Molucca Sea Plate subduction under Halmahera. Differential subsidence on NW-SE and NE-SW faults led to the formation of deep basins adjacent to eastern Halmahera. A major deformation event in the Pleistocene resulted in cessation of volcanism and shortening between eastern and western Halmahera. This deformation event is interpreted to be related to the interaction of the Molucca Sea Plate with either a fragment of the Australian continent and/or the subducted Philippine Sea Plate beneath eastern Halmahera. The present active arc has been shifted westward and is built on the deformed Pliocene arc.

Several papers succeeded these initial reports to elaborate the processes involved. Ballantyne (1990, 1991a, 1991b, 1992) and Ballantyne & Hall (1989, 1990) wrote series of papers on the petrology and geochemistry of the eastern Halmahera ophiolite. These papers suggested that the
ophiolite has a polygenetic origin, including rocks of boninitic, island arc and within-plate affinity. The boninites represented incipient supra-subduction magmatism, succeeded by island arc volcanism. The within-plate rocks are interpreted as remnants of an accreted sea mount. The petrology and chemistry of these rocks reinforced earlier interpretations that the Halmahera ophiolite (now imbricated as part of the fore-arc) is similar to the Mariana fore-arc. Currently, the northern part of eastern Halmahera is part of the Philippine Sea Plate fore-arc, while the southern part of it is the back-arc of the Halmahera arc (Hall, 1990).

Hakim (1989) studied the petrology, geochemistry and low temperature alteration of the volcanic rocks in western Halmahera. The Late Cretaceous-Eocene Oha Formation is interpreted as the arc related to the eastern Halmahera fore-arc. These rocks are arc tholeiites with olivine, plagioclase and clinopyroxene fractionation trends which subsequently suffered zeolite to sub-greenschist facies burial metamorphism. The Weda Group volcanic rocks are the product of Molucca Sea Plate subduction under Halmahera. The rocks are mostly calc-alkaline andesites with plagioclase, pyroxene, hornblende and magnetite fractionation. They have suffered very-low grade metamorphism, typical of geothermal environments. Hakim & Hall (1991) suggested that the Oha Formation is a result of subduction within the Pacific and may be correlated with similar basement rocks in the Philippine Sea region.

Hall et al. (1990) noted that the basement of Halmahera is similar to submarine plateaus and ridges in the Philippine Sea Plate and the basement of eastern Philippine. These similarities indicate the existence of an extensive region of Late Cretaceous-Eocene arc crust built upon Mesozoic ophiolites. This thickened crust was subsequently fragmented by spreading in the West Philippine Central Basin and back-arc spreading in the Eastern Philippine Sea. They suggested that a proto-Philippine archipelago, with short-lived intra-oceanic arcs, may be the closest modern analogue.

Hall et al. (1991) identified the basement rocks of the southern Molucca Sea region to include ophiolitic, arc volcanic and continental rocks. The ophiolites are associated with arc and fore-arc complexes, and are interpreted to be part of the Philippine Sea Plate. The volcanic basement has an island arc character of probable Late Cretaceous-Paleogene age, overlain by Eocene limestones and an Oligocene rift sequence, including pillow basalt and volcanioclastic turbidites. The distribution of the Eocene-Oligocene sequences indicate a pre-Middle-Late Eocene amalgamation of ophiolitic and arc terranes. Continental crust probably arrived in Late Palaeogene-Early Neogene, due to collision of the Australian Margin with Pacific arc-ophiolite terranes or by terrane
movement along the Sorong Fault Zone. The Bacan region is underlain by all three types of basement.

The petrography and provenance of Neogene sandstones was studied by Kusnama (1990). He was able to identify the provenance of sandstones of southwest and southeast Halmahera. The former is dominated by arc detritus, whereas the latter is characterised by ophiolitic detritus. This led him to conclude that the eastern and western arms of Halmahera were separated before the Pliocene. The Pliocene arenite of central Halmahera is dominated by volcaniclastic debris from the Halmahera arc. Based on the absence of quartz in the sandstone detritus, Nichols et al. (1991) suggested that the Halmahera Region has not been affected by continental input throughout the basin’s history. They attributed the changes in Halmahera sandstone petrography to evolution of the arc and not the degree of dissection of the arc massif. Hall & Nichols (1990a) and Nichols & Hall (1991) proposed that the Halmahera back-arc formed due to subsidence of thickened crust of imbricated older arc and ophiolite rocks.

Following the R.R.S Darwin Cruise in Indonesia, Nichols et al. (1990) described the southern termination of the Philippine Trench based on GLORIA side-scan sonar and single channel seismic profiles. They suggested that the movement of the southward propagating trench was impeded where it met a thickened, elevated oceanic plateau (the East Morotai Plateau). The convergence was then transferred via a broad NE-SW zone of transpressional dextral strike-slip across northern Halmahera into the Moluccas Sea Collision Zone. To the south of this strike-slip zone is the Eastern Halmahera-Waigeo Ophiolitic Terrane, which lies between the Philippine Sea Plate and the Sorong Fault Zone. The western part of Halmahera is therefore considered to lie in a diffuse boundary zone at the margin of the Philippine Sea Plate (Fig.2.3g).

Hall & Nichols (1990b) postulated that the heterogeneous character of the Philippine Sea Plate strongly influenced the position, nature and evolution of the plate boundaries. In the Halmahera Region, further convergence between the Philippine Sea Plate and Eurasia will result in the Philippine Trench propagating southward, meeting the Sorong Fault Zone, or the development of a new trench to the east of the East Morotai Plateau. Either way the East Halmahera-Waigeo Terrane will be accreted to the Eurasian Margin.

Milsom et al. (1992) used the termination of the Philippine Trench as an example of the development of a trench along strike. They argued that if propagation of a trench could be influenced by nature of the lithosphere at distance of 100 km or more along the trend, the
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Philippine Trench would have propagated east of the East Morotai Plateau. The fact that it did not suggested that propagation was determined by immediate and local weaknesses and not by regional factors.

2.2 REGIONAL GEOLOGY AND TECTONIC FRAMEWORK

The tectonic complexity and lack of geological data from the Bacan region are apparent from this review. There are now, however, some identifiable geological and tectonic consistencies which provide the framework for further study (Fig.2.4).

2.2.1 Halmahera and The Philippine Sea Plate

To the east of the Bacan region is the southwest arm of Halmahera. The oldest rocks observed are the Late Cretaceous Oha Fm volcanic sequence interpreted to be the arc of a west-facing subduction zone (Hakim & Hall, 1991). Unconformably upon it are Late Neogene volcanic arc and associated back-arc volcaniclastic rocks of the Weda Group, products of the Molucca Sea Plate subduction under Halmahera (Hall, 1987; Hall et al., 1988a; 1988b; Nichols & Hall, 1991; Nichols et al., 1991).

The eastern arm of Halmahera consists of a disrupted polygenetic ophiolite sequence with its associated Late Cretaceous-Eocene fore-arc deposit (Hall et al., 1988a; Ballantyne, 1990; Ballantyne & Hall, 1990; Hall, 1990; Hall et al., 1990; Ballantyne 1991a,b; Hall et al., 1991).

To the east of Halmahera is the Philippine Sea Plate, which is subducting westward under the Philippine archipelago and northernmost Halmahera. In the process the plate is rotating clockwise with respect to Eurasia (Haston & Fuller, 1991; Hall et al., 1993). This rotation has taken place since the Eocene, with a total clockwise rotation of ~90°. The trench east of Halmahera is very young with less than 150 km of subducted lithosphere (Hamilton, 1979; Cardwell et al., 1980). The trench terminates south of 2°50'N (Nichols et al., 1990; Milsom et al., 1992), where the accretionary prism becomes less deformed. There is no physiographic or seismic evidence linking the trench and other plate boundaries. Nichols et al. (1990) linked the southern termination of the Philippine Trench with the Molucca Sea Collision Zone via a NE-SW dextral strike-slip zone. The Halmahera Region is, therefore, in the process of amalgamation from the Philippine Sea Plate to the Eurasian Plate (Hall & Nichols, 1990b; Nichols et al., 1990).

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Figure 2.4. Present day tectonic map of Eastern Indonesia; modified from Hamilton (1979), Moore & Silver (1983), Nichols et al. (1990) and Milsom et al. (1992).
2.2.2 The Halmahera Arc
To the north of Bacan is the active volcanic arc formed by the eastward-subducting Molucca Sea Plate under Halmahera (Hatherton & Dickinson, 1969; Katili, 1975; Cardwell et al., 1980; McCaffrey, 1982). The present northern boundary of this arc is the volcano of Rurukumu, in northwest Halmahera and its present southern end is now located north of Bacan at Makian. To the south, the Sorong Fault Zone terminates the Neogene arc, and forms the southern boundary of the Philippine Sea Plate.

Hall (1987) suggested the collision of East and West Mindanao, in the Pliocene, to be the trigger for subduction under Halmahera to absorb the convergence between the Philippine Sea Plate and Eurasia. The volcanic rocks associated with this subduction have been found throughout the western arm and central Halmahera, the Bacan region, Obi and the island arc chain west of Halmahera.

2.2.3 The Molucca Sea Plate
A thick fore-arc apron deposited on the Molucca Sea Plate is present to the west of Kasiruta, deduced from the seismic reflection profiles of McCaffrey et al. (1980a). The age of the Molucca Sea Plate is unknown. To the north of the Molucca Sea Plate is the Philippine archipelago whilst to the south of it is the Sorong Fault Zone. The Molucca Sea Plate itself is currently being subducted on both east and west sides under Halmahera and the Sangihe Arcs respectively, and has no spreading centre (Hatherton & Dickinson, 1969; Cardwell et al., 1980; McCaffrey et al., 1980a). The plate is, therefore, sinking on both sides causing the two overriding arcs to collide (Silver & Moore, 1978; Hamilton, 1979), forming a central ridge, exposing the islands of Mayu, Tifore and Talaud (Silver & Moore, 1978; Moore & Silver, 1983). The northward continuation of this thrust zone is interpreted as the Philippine Fault Zone by some (Cardwell et al., 1980), although other workers have doubted this (e.g. Moore & Silver, 1983; Hall & Nichols, 1990b; Pubellier et al., 1991). Fig.2.5 is a three-dimensional cartoon diagram of the Molucca Sea Plate, showing it as a buffer zone between the collision of Eurasian and Philippine Sea Plates.

2.2.4 Sorong Fault Zone
The Sorong Fault Zone is a major sinistral strike-slip system, running east-west from the Bird's Head of Irian Jaya to eastern Sulawesi. This fault zone forms the boundary between the Molucca Sea Plate and the Philippine Sea Plate to the north and the Australian Continental Margin of Irian Jaya, Seram and Buru to the south (Hamilton, 1979; Letouzey et al., 1983; Pigram & Panggabean, 1984; Dow & Sukamto, 1984).
Figure 2.5. Three dimensional plate configuration in the Bacan region (after Hall et al., 1992).
To the south of the Bacan region there is a strand of the Molucca-Sorong Fault Zone, separating it from Obi (Hamilton, 1979; Morris et al., 1983; Letouzey et al., 1983; Dow & Sukamto, 1984). Strands of the Sorong Fault Zone are shown to cut in a WNW-ESE direction, forming the northern and southern boundaries of the Weda Bay, up to central Halmahera (Hall & Nichols, 1990b). The origin of Weda Bay is interpreted to be related to movement of different strands of the Sorong Fault Zone.

2.3 SUMMARY
From the work done in the region, it is clear that the Bacan region occupies an important position in the complex plate boundaries. It is located in the convergent zone of the Australian, Pacific and Eurasian Plates (Fig.2.5). Within this unstable triple junction zone (Nichols et al., 1990) a number of small plates and crustal fragments (terranes) of continental, oceanic and arc affinities are in a process of amalgamation (Hall, 1987; Hall & Nichols, 1990b). The Bacan region contains rocks of all of these affinities, thus its tectonic history becomes central in understanding the processes involved the evolution of a collision zone.

Despite this, the geology of the Bacan region is relatively poorly studied. Most of the publications in the area are reviews based on early data. There are only a handful of workers who have conducted field work in the region, and of these many are concerned only with mineral exploitation. Limited surveys and difficulties of differentiating volcanic rocks of different ages led to a variety of stratigraphic divisions, geological maps and subsequently inaccurate models of tectonic development. Fig.2.6 shows a comparison of the stratigraphy erected by Van Bemmelen (1949), Yasin (1980), Silitonga et al. (1981) and Hall et al. (1988a, 1988b).

A more thorough stratigraphy and geological understanding of the Bacan region is necessary before any model of the tectonic evolution of the region is attempted. This study provides the basis for such modelling.
<table>
<thead>
<tr>
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</tr>
</thead>
<tbody>
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<td>Pleistocene</td>
<td>N 22-23</td>
<td>Reef and Volcanics</td>
<td>Reef Alluvium Volcanic</td>
<td>Young Sediment</td>
<td>Pliocene - Pleistocene Volcanics</td>
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<td>Young Neogene</td>
<td>Obit Fm</td>
<td>Young Volcanic</td>
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<td>5</td>
<td>N 17</td>
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<td>10</td>
<td>N 16</td>
<td></td>
<td>Ruta Fm</td>
<td>Older Sediment</td>
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<td></td>
<td>Amasing Fm</td>
<td>Older Sediment</td>
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<td></td>
<td>N 9-11</td>
<td></td>
<td>Tuffaceous sst, Litoral-neritic</td>
<td>Older Volcanic</td>
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<tr>
<td>20</td>
<td>N 7-8</td>
<td>Old Neogene (and ?Younger Palaeogene)</td>
<td>Bacan Fm</td>
<td>Volunteer with tuff intercalations.</td>
<td>Older Volcanics</td>
</tr>
<tr>
<td>30</td>
<td>P 4-6</td>
<td>Limestone, Tuffaceous sandstone, Conglomerate, Breccia and some Lignite</td>
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<td>Volcanic and sedimentary rocks</td>
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<td>P 20-22</td>
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</tbody>
</table>

Figure 2.6 The different stratigraphies for the Bacan Region. Van Bemmelen and Hall et al., include the Halmahera Region.
CHAPTER THREE
METAMORPHIC COMPLEXES:
THE SIBELA AND SALEH METAMORPHIC COMPLEXES

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CHAPTER THREE
METAMORPHIC COMPLEXES:
THE SIBELA AND SALEH METAMORPHIC COMPLEXES

3.1 INTRODUCTION
Although both extensive and commonly cited, the metamorphic complexes are nevertheless the least well studied of all the rocks in the Bacan Region. They include high mountainous areas covered by impenetrable rain forest, and these physical problems coupled with structural, lithological and mineralogical complexities place limitations on field and laboratory studies. This chapter includes brief descriptions to characterise these complexes, including: [1] summary of mineral assemblages; [2] chemical analyses of minerals; [3] chemical analyses of whole rocks and [4] new isotopic data concerning the age of these complexes. A detailed study of metamorphic history has not been attempted.

In this chapter, the term Metamorphic Complex is defined as a mappable area of rocks having characteristics of dynamothermal metamorphism. In the Bacan Region there are two such metamorphic complexes: the Sibela and the Saleh Metamorphic Complexes (Fig.3.1).

3.2 SYNONYMY
The first mention of a metamorphic complex in the Sibela Mountains is by Brouwer (1923), who cited the occurrence of gneisses, mica schists, epidote chlorite schists and amphibolite, diorites, gabbros, peridotite, serpentinite, andesite and basalt. In more recent years, Silitonga et al. (1981) mapped both the Sibela Mountains and the Saleh Complex as 'schist or phyllite'. The Sibela Metamorphic Complex was formally named by Yasin (1980), although he did not differentiate between the Continental, the Ophiolitic and the Saleh Metamorphic Complexes. Hall et al. (1988a) divided the Sibela Complex into the Continental and Basic/Ultrabasic Basement, and interpreted the Saleh Complex as the upper part of the Sibela Basic/Ultrabasic Basement.

3.3 THE SIBELA METAMORPHIC COMPLEX, CONTINENTAL SUITE
The Sibela Complex is divided into two sub-complexes, these are rock suites with: [1] continental and [2] ophiolitic affinities. On aerial photographs, the Sibela Metamorphic Complex is seen to form a high, elongate ridge with a central summit, having a parallel drainage pattern producing a distinctive morphology. The continental suite differs from the ophiolitic in that the latter is lower in elevation and has a component of dendritic drainage pattern. The continental suite is typified by strong penetrative fabrics, observed in both outcrops and thin sections, indicative of a polyphase deformation and recrystallization history. Due to the limited time for survey and
Figure 3.1 Map showing the distribution of the Sibela and the Saleh Metamorphic Complexes.
difficulties of the terrain, structural analysis of the Sibela Complex was not conducted. Aerial photographic studies, however, clearly show that the complex is an uplifted horst. The contact between the continental and ophiolitic rocks is interpreted to be a possible thrust fault.

3.3.1 Lithofacies

Type Section

S. Penambuan is the type section for the continental suite and boulders of garnetiferous mica schist and garnet amphibolite gneiss are common. Garnet porphyroblasts are up to 5 mm across; other porphyroblasts include staurolite and kyanite. The rocks contain quartz segregations which are flattened in the plane of schistosity.

High temperature, medium pressure metamorphism of upper amphibolite facies is inferred from the mineralogy of this metamorphic complex. Low grade alteration is absent in most rocks. Most of the samples analysed show intense foliation, with both L and crenulated S fabrics.

The continental suite includes calcareous phyllite, quartzo-feldspathic phyllite, amphibole quartz phyllite, staurolite garnet mica schist, kyanite garnet mica schist, quartzo-feldspathic gneiss, with minor talc marble and quartzite.

Thickness

Based on the height of the Sibela Mountains, the thickness of the continental suite is estimated to be at least 2000 m.

Lateral Extent and Variations

The continental suite dominates the western half of Sibela Mountains in the central block of Bacan. Outside the region, similar rocks are known to the south from P. Tapas, off the northwest coast of P. Obi.

3.3.2 Petrography and Mineral Chemistry

Only limited petrological and mineralogical studies have been conducted on this metamorphic complex, mainly to gain an insight into conditions of metamorphism and to assess the nature of the protoliths. Fig.3.2 shows the locality of samples used in analytical work.

3.3.2.1 Petrography

Most of the rocks in the Sibela Complex are pelites, interlayered with quartzo-feldspathic layers
Figure 3.2 Location of samples from Metamorphic Complexes used in analytical work

Key:
- Saleh Complex
- Sibela Ophiolite
- Sibela Continental

*Italics = Not In Situ*
Metamorphic Complexes

and calcareous quartzo-feldspathic rocks. There are some rare carbonate-rich rocks and metabasites. Appendix E contains brief mineralogical assemblages of the Continental Suite. Mineral assemblages in this suite include:

[1] pelites (Plate 3.1a,b):
   [a] garnet + staurolite + biotite + muscovite + quartz + plagioclase + graphite + iron ore,
   [b] garnet + staurolite + kyanite + muscovite + quartz + plagioclase + graphite + iron ore,
   [c] garnet + staurolite + kyanite + biotite + muscovite + quartz + plagioclase + graphite + iron ore.

[2] quartzose and quartzo-feldspathic rocks (Plate 3.1c,d):
   muscovite + quartz ± plagioclase + tourmaline + iron ore.

[3] calc-quartzo-feldspathic rocks (Plate 3.2a,b):
   [a] biotite + plagioclase + epidote + sphene + quartz + apatite + iron ore (+ retrograde chlorite).
   [b] biotite + muscovite + plagioclase + epidote + quartz + apatite + iron ore (+ retrograde chlorite).

[4] impure carbonates (Plate 3.2c):
   calcite + talc ± chondrodite.

[5] metabasites:
   hornblende + garnet + plagioclase + quartz + iron ore.

Sigmoidal internal mineral fabrics are common in the thin sections, indicating porphyroblast growth during gradual rotation, possibly due to simple shearing. Poikiloblasts are often rotated, and some garnets exhibit snowball texture (Plate 3.2d).

3.3.2.2 Mineral Chemistry

Five samples were chosen for mineral chemical analysis with mineral assemblages given in parentheses corresponding to those listed above.

Pelites: B22 [1b], B78 [1a], B105 [1a].
Calcereous quartzo-feldspathic rocks: B79 [3a], B102 [3b].

Plagioclase and Orthoclase

Plagioclase is present in B78, B79 and B102, commonly as matrix and rarely as porphyroblast. All plagioclases are unzoned sodic varieties with compositions in the range An_{17-31} (Fig.3.3a). There is no significant difference in plagioclase composition between pelitic and calc-silicate rocks. A single orthoclase analysis was obtained from B79.

Sheet silicates

Muscovite is present in B22, B78, B105 and B102. It is commonly intimately intergrown with fine elongate opaque grains (iron ore) and analyses of these grains have apparently higher Fe_{2}O_{3} and
Plate 3.1a,b. PPL and XPL of graphitic metapelites (B84); note the presence of garnet and staurolite porphyroblasts in a muscovite, biotite, quartz and plagioclase matrix.

Plate 3.1c,d. PPL and XPL of tourmaline-bearing quartzo-feldspathic gneiss (B5).
Scale Bar = 0.5 mm for all photomicrographs.
Plate 3.2a,b. PPL and XPL of calc-quartz-feldspathic gneiss (B102); note the presence of epidote and plagioclase porphyroblasts in a biotite and quartz matrix.

Plate 3.2c. XPL of impure carbonates (B1); note pyrope garnet porphyroblasts in a talc and carbonate matrix.

Plate 3.2d. Garnet with snowball texture (B2); mica inclusions in garnet. Scale Bar = 0.5 mm for all photomicrographs.
lower Na₂O contents. Compositions are comparable to those of end-member muscovite compositions with low celadonite contents (Deer et al., 1992). Biotite occurs as part of the matrix in B78, B79, B102 and B105 and has relatively high Mg/Fe ratios. Poorly crystalline chlorite is present in B79 and B102 as a retrograde mineral replacing biotite.

**Staurolite**

Staurolite occurs in B22 and B78 as idioblastic porphyroblasts with a well developed cleavage. The analyses are closely similar to staurolite from a staurolite mica schist (Deer et al., 1992).

**Garnet**

Large garnet poikiloblasts, with mica and graphite inclusions, is present only in the pelites B22, B78 and B105. When compared to almandine garnet from a quartz biotite gneiss (Deer et al., 1992), these have a lower MgO and a higher CaO contents. Ferry & Spear (1978) utilise the partitioning of Fe and Mg in biotite and garnet as a geothermometer and this system is relatively insensitive to pressure. Using their method the temperature calculated for the pelitic rocks is ~505°C at 5 kb.

**Kyanite**

Probable kyanite is present in one sample (B22) but is entirely replaced by a fine grained sheet silicate, probably retrograde pyrophyllite.

**Fe-Ti Oxides and Iron Ore**

Iron oxides are present in B22 and B78. Fig.3.3b shows that they are close to pure ilmenite in composition. Fine disseminated specular iron ore is present in most samples as a reaction product surrounding porphyroblasts. These are similar to graphite.

**Sphene**

Sphene occurs in B79; and these have a lower SiO₂ and higher TiO₂ contents compared to Coombs et al. (1976). They are similar to the ophiolitic sphenes (see below).

**Epidote**

Granular epidote occurs in the calc-silicate rocks B79 and B102. These have a lower Al₂O₃ and higher Fe₂O₃ compared to epidote from an amphibolite (Deer et al., 1992) and are similar to the ophiolitic epidote (e.g. BM572). Epidote and sphene do not coexist with staurolite or garnet.
Figure 3.3a Ternary diagram showing all feldspar compositions from the Sibela Continental Suite.

Figure 3.3b. Ternary diagram showing all Fe-Ti oxides and iron ore compositions from the Sibela Continental Suite.
Metamorphic Complexes

Quartz
Polycrystalline quartz is a ubiquitous mineral, occurring in all samples in the matrix.

Graphite
Graphite is present in almost all samples, either in the matrix or as a reaction product surrounding large porphyroblasts. These are impossible to analyse.

3.3.3 Implications
The presence of staurolite ± garnet ± kyanite is typical of an amphibolite facies regional metamorphism of pelitic sediments (Winkler, 1979; Yardley, 1989; Deer et al., 1992). The variation in mineralogy reflects small rock compositional differences, both for whole samples and within layers. Many of the pelites have quartz-rich and quartz-poor layers. Minor quartzite and marble layers and samples demonstrate sandy and carbonate/marly protoliths within the pelites.

The assemblage of chlorite, biotite, muscovite, almandine garnet, staurolite ± kyanite indicates medium grade metamorphism of Barrovian type. In compositional terms there is a rather restricted range and on the ACFK diagram, these rocks plot in a limited volume between epidote, muscovite and biotite. Kyanite appears rarely in thin layers without garnet which are presumably Fe-poor compared to other Fe-rich layers which lack kyanite and contain staurolite + garnet + biotite. The presence of kyanite and staurolite suggests pressures of at least ~4-5 kb at temperatures of ~500°C indicated by biotite-garnet equilibria. Chlorite and pyrophyllite are the products of retrograde alteration of biotite and kyanite associated with post peak-metamorphic deformation.

3.3.3 Whole Rock Chemistry
Three in situ samples were analysed for whole rock chemistry (B22, B79 and B102). The aims of this study were: [1] to determine nature and composition of the protoliths and [2] to assess the original depositional setting.

3.3.3.1 Major Elements
Van de Kamp (1968) attempted to decipher the nature of the protoliths from a study of whole rock chemistry of medium grade metamorphic rocks. The Sibela Metacontinental major element results are compared to that study in Table 3.1.

The principal difference between the Sibela pelites and calc-silicate rocks is that the latter have lower Al and higher Ca and Na. All have higher Si than the protoliths analysed by Van de Kamp.
Metamorphic Complexes

The Sibela pelites differ from Van de Kamp's average pelite in having higher Si and lower Fe, Mg, Ca and Na. The high SiO₂ suggests high quartz in the protolith, suggesting a cratonic provenance (Dickinson & Suczek, 1979).

Table 3.1 Comparison of Metacontinental major elements with other gneisses and pelites. FeO* is the total Fe content (FeO + Fe₂O₃ x 0.998).

<table>
<thead>
<tr>
<th>Wt%</th>
<th>B22 Pelite</th>
<th>B79 Calc-silicate</th>
<th>B102 Calc-silicate</th>
<th>Biotite Gneiss</th>
<th>Carbonate-silicate gneiss</th>
<th>Average pelites</th>
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<tr>
<td>SiO₂</td>
<td>65.96</td>
<td>67.04</td>
<td>66.45</td>
<td>60.76</td>
<td>61.76</td>
<td>58.64</td>
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<td>TiO₂</td>
<td>0.87</td>
<td>0.48</td>
<td>0.28</td>
<td>0.85</td>
<td>0.69</td>
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<td>Al₂O₃</td>
<td>21.18</td>
<td>16.05</td>
<td>18.43</td>
<td>15.45</td>
<td>15.54</td>
<td>21.12</td>
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<tr>
<td>FeO*</td>
<td>5.18</td>
<td>3.64</td>
<td>2.29</td>
<td>7.17</td>
<td>6.81</td>
<td>8.48</td>
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<td>MnO</td>
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<td>0.11</td>
<td>0.10</td>
<td>0.13</td>
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<td>MgO</td>
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<td>1.70</td>
<td>0.98</td>
<td>3.80</td>
<td>4.92</td>
<td>2.81</td>
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<td>CaO</td>
<td>0.46</td>
<td>4.11</td>
<td>5.06</td>
<td>5.36</td>
<td>5.08</td>
<td>1.54</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.49</td>
<td>3.80</td>
<td>5.04</td>
<td>3.62</td>
<td>2.94</td>
<td>1.72</td>
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<td>K₂O</td>
<td>3.59</td>
<td>2.83</td>
<td>0.98</td>
<td>2.46</td>
<td>1.80</td>
<td>3.80</td>
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<tr>
<td>P₂O₅</td>
<td>0.18</td>
<td>0.14</td>
<td>0.12</td>
<td>0.21</td>
<td>0.24</td>
<td>0.21</td>
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</tbody>
</table>

3.3.3.2 Trace Elements

Van de Kamp (1968) also used trace element ratios to distinguish different protoliths. Table 3.2 compares the Sibela Metacontinental lithologies with his average shale, carbonate, greywacke and para-amphibolite.

Table 3.2 Comparison of Sibela Metacontinental trace elements with other sedimentary rocks and gneisses.

<table>
<thead>
<tr>
<th></th>
<th>B22 Pelite</th>
<th>B79 Calc-Silicate</th>
<th>B102 Calc-Silicate</th>
<th>Shale</th>
<th>CaCO₃</th>
<th>Greywacke</th>
<th>Para-Amphibolite</th>
</tr>
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<tr>
<td>Rb/Sr</td>
<td>1.61</td>
<td>0.20</td>
<td>0.02</td>
<td>0.50</td>
<td>0.008</td>
<td>-</td>
<td>0.11</td>
</tr>
<tr>
<td>La/Ce</td>
<td>0.46</td>
<td>0.53</td>
<td>0.44</td>
<td>0.39-0.72</td>
<td>0.50</td>
<td>0.73</td>
<td>0.17</td>
</tr>
<tr>
<td>Th (ppm)</td>
<td>19.5</td>
<td>10.5</td>
<td>1.0</td>
<td>12</td>
<td>1.7</td>
<td>-</td>
<td>4.5</td>
</tr>
</tbody>
</table>
Metamorphic Complexes

The Rb/Sr ratios vary widely within the continental suite, probably due to mobility of both of these elements. The La/Ce ratios are within the ranges of both shale and carbonate protoliths, although they are considerably higher than para-amphibolites. The Th content varies widely, with B22 and B79 having values closest to the average shale, and B102 having a value close to carbonate rocks.

3.3.3.3 Tectonic Discrimination Diagrams

There are many published tectonic discrimination diagrams to relate sedimentary provenance to tectonic settings based on the whole rock chemistry of the sediments. Although most of the diagrams are based on sands, Roser & Korsch (1986) suggested that similar discrimination techniques can be applied to pelites. Mortimer & Roser (1992) indicated that Ti, Zr, La, Ce and Y are among the elements resistant to mobility during metamorphism, and used discriminant diagrams to decipher the protolith of rocks metamorphosed to greenschist facies.

Roser & Korsch (1986) used variation of the $K_2O/Na_2O$ ratio with $SiO_2$ to distinguish Arc, Active Continental Margin (ACM) and Passive Margin (PM) rocks. Fig.3.4a shows the Sibela Metacontinental rocks plot in the ACM, PM and Arc fields. Copeland & Condie (1986) utilised the covariation of $TiO_2$ with $FeO^*+MgO$ to characterise cratonic, extensional continental, continental arc and oceanic arc basins. On this diagram (Fig.3.4b) the metacontinental rocks plot near the cratonic field, except B22 which plots in the oceanic arc basin. Bhatia & Crook (1986) used the trace element ratios ($Ti/Zr$, $La/Sc$, $La/Y$ and $Sc/Cr$) to develop similar diagrams. Fig.3.4c shows that on the $Ti/Zr-La/Sc$ diagram the Sibela Metacontinental rocks plot in the continental arc field, except B102 which plots in the oceanic island arc field. On $La/Y-Sc/Cr$ diagram (Fig.3.4d), only B22 plots in the continental arc field, the others have higher $Sc/Cr$ ratios compared to the defined fields. Mortimer & Roser (1992) proposed that $La$ and $Ce$ are enriched in continent-derived sediments, whereas $Y$ and $V$ in oceanic arc-derived sediments. Using their $La/Y$ against $Ce/V$ covariation diagram (Fig.3.4e), the Sibela Metacontinental Suite plots in the continental field, except B102 which plots in the arc field. Roser & Korsch (1988) calculated discriminant functions based on major element data to decipher the provenance of sediments. On their diagram (Fig.3.4f), the metacontinental suite plots in the 'mostly andesitic protolith' field.

Additionally, Bhatia & Crook (1986) introduced ternary diagrams using trace elements to decipher tectonic setting of sediments. On the $La-Th-Sc$ (Fig.3.5a) and the $Th-Sc-Zr/10$ (Fig.3.5b) diagrams, the metacontinental rocks plot consistently in the continental arc field, except B102 which plots in the oceanic arc field.
Figures 3.4a-f. Covariations of major and trace elements to decipher the tectonic setting of the Sibela Continental Suite protoliths. See text for details.
Figures 3.5a-b. Trace element ternary diagrams to determine the tectonic setting of the Sibela Continental Suite protoliths. After Bhatia & Crook (1986).
3.3.4 Implications

The major element results indicate that the protoliths were pelitic, quartz-rich sediments, possibly derived from a cratonic source. The trace element geochemistry supports this, with the suggestion that B102 contained some carbonate component. These results are consistent with derivation from the Australian continent (especially considering the lack of quartz in both ophiolitic and arc terranes) but with an original depositional setting on an active continental margin (?continental arc).

3.3.4 Age Determination

Attempts were made to date two samples from the Sibela metacontinental suite by the Ar-Ar and K-Ar methods. Both samples analysed, a calc-quartzo-feldspathic schist (B102) and a staurolite-garnet-mica schist (B22), yielded <1 Ma ages. The Ar-Ar step heating profile (Tables 3.3 and 3.4) does not show a diffusion loss profile (indicating episodic \(^{40}\text{Ar}^*\) [radiogenic \(^{40}\text{Ar}\)] loss; McDougall & Harrison, 1988; Ross & Sharp, 1988) or profile with anomalous plunges (indicating thermally overprinted samples; Berger, 1975), suggesting that either the rocks have lost their argon completely or the assumed initial value for \(^{40}\text{Ar} / ^{36}\text{Ar}\) is invalid.

In order for the samples to have lost argon completely, from a near pure muscovite, the rock must have passed the blocking temperature of muscovite (~350°C) recently. As neither sample has any indication (field, petrographic or mineral chemical evidence) of thermal overprint, there is no reason to believe that these samples have been affected by any thermal event recently. On the other hand, it is possible that there would be no observable effect of heating an already medium-high grade metamorphic rock at temperatures not significantly greater than 350°C. However, further evidence that argon loss is not the cause for the young age is the fact that rocks from the Ophiolite Complex, located close to these samples (e.g B102 and B103) have retained their argon (see Section 3.4.4).

The alternative is that the assumed initial value for \(^{40}\text{Ar} / ^{36}\text{Ar}\) is incorrect. The young ages are, therefore, suggested to be the result of interaction of rocks with a low temperature (~100°C) hydrothermal fluid containing large amounts of fractionated Ar having a low \(^{40}\text{Ar} / ^{36}\text{Ar}\) ratio. This would mean that there need not have been \(^{40}\text{Ar}^*\) loss and the temperature of the fluid could have been lower than that required to completely reset the muscovite system (the extremely young age excludes the possibility of partial resetting). The hydrothermal fluid may affect the samples by adding fractionated argon between the muscovite layers (cf. Lippolt et al., 1990). Alternatively a hydrothermal fluid may have entered the muscovite system in the past, in the form of fluid
inclusions (Turner, 1988) with the effect of lowering the apparent age dramatically. Miller et al. (1991) described hornblendes from a polymetamorphic suite whose argon had been flushed out by low temperature (sub-blocking temperature) fluid reaction, significantly lowering their K-Ar ages. Conceivably similar processes could have affected the muscovites in the Sibela Continental Suite. The negative $^4\text{Ar}^*$ values, the low $^{40}\text{Ar}/^{36}\text{Ar}$ ratios in both samples (less than the assumed value of 295.5), and the high $^{38}\text{Cl}$ value in the first step heating result of B22 are further evidence of $^{36}\text{Ar}$ enrichment.

There are currently nineteen hot springs around the Sibela Mountains (Yasin, 1978) with emerging water temperatures ~100°C. These are located along obvious faults for which there is often evidence of recent movements. There are also common mylonite and minor fault zones within the Sibela rocks which could allow fluid access.

These results have a wider implication for other metamorphic rocks in the region. In the Bird’s Head of Irian Jaya (Pigram & Davies, 1987) high grade metamorphic rocks have yielded Neogene ages attributed to Neogene metamorphism although the very young metamorphic ages require improbably high uplift rates. On Seram and Buru (Linthout et al., 1991) Palaeozoic metamorphic rocks have young ages attributed in part to Neogene resetting due to obduction of a hot ophiolite. However, Linthout et al. (1991) reported that where some of their samples were collected there were no ophiolitic rocks or even ophiolitic clasts in younger sediments to indicate the presence of a former ophiolite. They ascribed these young ages to reheating due to hydrothermal activity. Obduction of a hot ophiolite also requires young oceanic crust for which there is no evidence other than the metamorphic ages and indirect arguments about the age of the North Banda Sea, for which various ages have been suggested, ranging from Cretaceous to Miocene. Circulation of hydrothermal fluids or thermal overprint may explain these young ages, particularly credible in view of the recent volcanic activity and fault movements in the region.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Material</th>
<th>Grain size (μ)</th>
<th>%K (1σ error)</th>
<th>Wt for Ar (g)</th>
<th>$^4\text{Ar}^*$ (nl/g, 1σ error)</th>
<th>$^{36}\text{Ar}_{un}$</th>
<th>Age (Ma, 2σ error)</th>
</tr>
</thead>
<tbody>
<tr>
<td>B102</td>
<td>muscovite</td>
<td>125-250</td>
<td>4.466 ± 1.00%</td>
<td>0.0821</td>
<td>&lt; LOD</td>
<td>-</td>
<td>&lt;0.21</td>
</tr>
<tr>
<td>B22</td>
<td>muscovite</td>
<td>125-250</td>
<td>4.0</td>
<td>0.0233</td>
<td>&lt; LOD</td>
<td>-</td>
<td>-0.3 ± -0.8*</td>
</tr>
</tbody>
</table>

* denotes integrated age from Ar-Ar analysis. LOD = Limit of Detection.
Metamorphic Complexes

Based on correlation with Australian continental material, Hamilton (1979) assigned the continental suite a Palaeozoic age. Although this seems plausible, radiometric dating has so far not confirmed this suggestion.

Table 3.4 Summary of Ar-Ar step heating results from the Sibela Metamorphic Complex, Continental Suite.

<table>
<thead>
<tr>
<th>Temp (°C)</th>
<th>(^{39})Ar/(^{37})Ar</th>
<th>(^{39})Ar/(^{40})Ar</th>
<th>(^{39})Ar/(^{38})Ar</th>
<th>Ca</th>
<th>(^{39})Ar/(^{39})Ar</th>
<th>Age (Ma)</th>
<th>Age Err (Ma)</th>
</tr>
</thead>
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<tr>
<td>650</td>
<td>0.028</td>
<td>97.8</td>
<td>0.5</td>
<td>3.1</td>
<td>9.2</td>
<td></td>
<td></td>
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<tr>
<td>730</td>
<td>0.002</td>
<td>11.4</td>
<td>-3.3</td>
<td>1.1</td>
<td>10.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>760</td>
<td>0.011</td>
<td>168.3</td>
<td>-2.1</td>
<td>2.4</td>
<td>8.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>795</td>
<td>0.007</td>
<td>147.6</td>
<td>-1.5</td>
<td>1.9</td>
<td>11.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>875</td>
<td>0.013</td>
<td>176.4</td>
<td>-2.3</td>
<td>1.1</td>
<td>21.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>945</td>
<td>0.007</td>
<td>107.4</td>
<td>-0.4</td>
<td>2.7</td>
<td>11.0</td>
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<td></td>
</tr>
<tr>
<td>1020</td>
<td>0.012</td>
<td>99.9</td>
<td>0.0</td>
<td>2.9</td>
<td>10.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1100</td>
<td>0.010</td>
<td>106.2</td>
<td>0.3</td>
<td>2.2</td>
<td>13.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1340</td>
<td>0.006</td>
<td>111.9</td>
<td>31.7</td>
<td>9.2</td>
<td>3.0</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**3.4 THE SIBELA METAMORPHIC COMPLEX, OPHIOLITIC SUITE**

This complex includes peridotites and gabbroic rocks considered to represent an incomplete ophiolite which has been metamorphosed to some degree. Many of the rocks appear to retain original magmatic textures. Some of the rocks are characterised by penetrative fabrics indicating polyphase deformation and recrystallization. The Sibela ophiolitic rocks are quite unlike the east Halmahera ophiolite in the nature and degree of deformation (Hall et al., 1988a) and are also very different in fabric and mineralogy to the rare metabasites of the Sibela Continental Suite. The main aim of this study was to determine the age of the ophiolitic rocks, the character of metamorphism and its significance.

**3.4.1 Lithofacies**

**Type Section**

S. Ra is the type section for the ophiolitic suite, where outcrops of layered metagabbros, amphibolites and serpenitnized harzburgite are present. Some metagabbros exhibit flaser textures, forming augen gneiss (Plate 3.3a). Complexly sheared metagabbros are also found with fabrics which appear to be have been acquired at or near magmatic temperature (Hall et al., 1988a; Plate 3.3b). Small scale deformation forming S and Z folds and boudinage structures were locally observed (Plate 3.3c). Hornblende porphyroblasts are up to 20 mm long. Compositional banding is defined by feldspar.

Hall et al. (1988a) reported, in S. Ra, lateral transitions from deformed basic rocks into less deformed equivalents which are intruded by undeformed gabbros and basalts, diorites and
Plate 3.3a. Flaser gabbro at S. Ra.
Plate 3.3b. Sheared gabbro at S. Ra.
Plate 3.3c. S and Z foliations in gabbro at S. Ra.
Plate 3.3d. Deformed multiple intrusions of metagabbros. From Hall et al. (1988a).
Metamorphic Complexes

Microdiorites (Plate 3.3d). They concluded that the deformed rocks were the product of active shear zones which were repeatedly intruded by basic magma, and therefore suggested that the ophiolitic rocks represent magmatism in a splay of the Sorong Fault system.

**Thickness**

Based on the height of the northeastern part of the Sibela Mountains (1291 m), the thickness of the ophiolitic suite is at least 1000 m.

**Lateral Extent and Variations**

The ophiolitic suite crops out in the northeastern part of the Sibela Mountains (Fig.3.1). Samples were collected from S. Mandawong, S. Gandasuli, S. Ra and S. Tawali and along the eastern coast of the Sibela Mountains.

Outside the Bacan region, ophiolitic rocks associated with arc-related sedimentary rocks form the basement of east Halmahera, Waigeo, Gag and Gebe. In all these islands, the ophiolitic rocks are situated to the north of the Sorong Fault Zone and now form part of the Philippine Sea Plate. Within the Sorong Fault Zone similar ophiolitic rocks with Cretaceous-Eocene sedimentary rocks form the basement of northern and central Obi.

### 3.4.2 Petrography & Mineral Chemistry

The main purpose of the petrographical study was to gain an insight into the metamorphic condition and tectonic setting of the ophiolite. Fig.3.2 shows the location of samples used for analytical work.

#### 3.4.2.1 Petrography

Five samples were chosen for mineral chemical study. B32 is a hornblende schist, B97 and B103 are peridotites, BM571 is a hornblendite and BM572 is an amphibolite. Appendix E contains brief mineralogical assemblages of rocks from the Ophiolitic Suite. The mineral assemblages of the five analysed rocks are:

- **B32** (mylonitized metabasalt) amphibole + plagioclase + iron ore + sphene (chlorite + prehnite) (Plate 3.4a,b)
- **B97** (metamorphosed peridotite) possibly primary clinopyroxene + amphibole + plagioclase + iron ore (+ sphene + epidote segregations + chlorite)
- **B103** (metamorphosed peridotite) possibly primary clinopyroxene + amphibole + iron ore; high T granoblastic texture
- **BM571** (?magmatic hornblendite) hornblende
- **BM572** (cumulate) hornblende + plagioclase + opaque (+ sphene + epidote + mica + chlorite) (Plate 3.4c,d)

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Plate 3.4a,b. PPL and XPL of mylonitic gabbro (B32); amphibole, plagioclase, and chlorite vein.

Scale for all photomicrographs are 0.5 mm.

Plate 3.4c,d. PPL and XPL of cumulate amphibolite (BM572); amphibole, plagioclase and iron oxide.
Plate 3.5a,b. PPL and XPL of dunite (B49); serpentine and iron oxide.

Scale for all photomicrographs are 0.5 mm.

Plate 3.5c,d. PPL and XPL of harzburgite (B77); relict orthopyroxene, serpentine, chlorite and prehnite.
Metamorphic Complexes

In addition, there are other ultrabasic and basic rocks, including serpentinized dunite (Plate 3.5a,b), harzburgite (Plate 3.5c,d), lherzolite and wehrlite; amphibole gneiss and epidote amphibole schist.

3.4.2.2 Mineral Chemistry

Amphiboles

Amphiboles are the most common minerals and are mostly tschermakitic and pargasitic hornblende, with minor ferroan pargasite and edenitic hornblende. Four inclusion analyses in BM571 are ferroan actinolite and tremolite.

Raase (1974) used the covariation of Al\textsuperscript{iv} with Si, and the Ti content of hornblende from rocks of known metamorphic conditions to estimate P and T. Fig.3.6a,b demonstrate that most amphiboles from the ophiolite plot at ~5 kb, with Ti contents typical of low temperature amphibolite facies. When the NaM4 content of the amphibole is plotted against the Al\textsuperscript{iv} + Fe\textsuperscript{iv} + Ti + Cr content (after Laird & Albee, 1981), amphiboles in B32 plot in the garnet zone (Fig3.6c).

Plagioclase

Plagioclase is present in B32 and BM572. The composition of plagioclase in each sample is: B32 An\textsuperscript{52}Ab\textsuperscript{48}Or\textsuperscript{0} (n=4); BM572 An\textsuperscript{98}Ab\textsuperscript{4}Or\textsuperscript{0} (n=4). The very calcic compositions suggest primary magmatic compositions. It is not clear if these are genuine magmatic plagioclases (suggested texturally in BM572 where the plagioclase appears to be an intercumulus phase) or are plagioclase that have recrystallized under metamorphic conditions in which calcic plagioclase is stable or has failed to re-equilibrate (B32 has a mylonitic fabric). From experimental results, Plyusnina (1982) used the compositions of coexisting plagioclase and amphibole to estimate PT and these suggest conditions of ~625°C and >8 kb pressure for B32. Blundy & Holland (1990) also utilised the composition of coexisting plagioclase and calcic amphibole as a semi-empirical geothermometer. Using their method at 1 kb, B32 yielded a temperature of 1041°C, at 5 kb it is 969°C, and at 8 kb it is 916°C, which are consistent with magmatic crystallization or near solidus temperatures.

Pyroxene

Only B103 contains pyroxene in the section probed and this has an average composition of En\textsuperscript{2}Fs\textsuperscript{5}Wo\textsuperscript{9} (n=6). These clinopyroxenes are augite and salite, and are comparable to those of Mevel et al. (1978) and Hawkins & Evans (1983) in gabbros. Nisbet & Pearce (1977) introduced a tectonic discrimination diagram based on analyses of volcanic clinopyroxenes from known tectonic settings and in this type of diagram the ophiolite pyroxenes plot in the VAB (Volcanic Arc Basalt) field (Fig.3.7a).
Metamorphic Complexes

**Fe-Ti minerals**

Five Fe-Ti minerals are present: rutile (B32), pyrite (B97), titanomagnetite (B97 and BM572), magnetite (BM572) and chrome spinel (B103). Rutile occurs as anhedral-euhedral porphyroblasts. Titanomagnetite occurs as inclusions in amphiboles, whereas magnetite is observed as a reaction product, surrounded by amphiboles and plagioclase. Chrome spinel is reddish-brown and occurs as anhedral grains with high Cr# \([\text{Cr}/(\text{Cr}+\text{Al})]\) and low Mg# \([\text{Mg}/(\text{Mg}+\text{Fe}^{2+})]\). They resemble spinels of the east Halmahera harzburgites which are highly depleted and are thought to represent the residuum of two stages of partial melting (Ballantyne, 1991b); extraction of MORB was followed by remelting in a hydrous environment such as a supra-subduction zone setting.

Dick & Bullen (1984) proposed that increasing Cr# of spinel reflects increasing degrees of partial melting. They divided Cr-spinels of alpine-type peridotites and associated basalt into three types: type I (Cr# < 60) similar to MORB; type II (transitional) formed in intra-oceanic island arc; type III (Cr# > 60) analogous to arc volcanics, continental intrusives and oceanic plateau basalts. The Cr# of the Sibela Ophiolite spinel are 0.75-0.87, suggesting type II or III origin. Arai (1992) used the TiO₂ contents, Cr# and Fe³# \([\text{Fe}^{3+}/(\text{Cr}+\text{Al}+\text{Fe}^{2+})]\) of spinel for tectonic discrimination and suggested that when spinel survived metamorphism, it can be used to determine original tectonic setting. Figs.3.7b,c show that the Sibela Ophiolite chrome spinel plots in Arai’s island arc and intraplate basalt fields.

**Sphene**

Euhedral-subhedral sphene is present in B32, B97 and BM572 and is similar to those reported in Deer et al. (1992) and Coombs et al. (1976). Spear (1981) suggested that the presence of sphene in an amphibolite (i.e. B32, B97) indicates high P, low T and high \(f_0_2\) conditions, typical of low grade amphibolite facies.

**Chlorite**

Chlorite is present in B32, B97 and BM572 where it is present in veins or as a replacement of amphibole. Following Hey (1954), these are pychnochlorite (B32 and BM572) and clinochlore (B97) with high Mg content (especially B97 and B32), which reflects the whole rock composition.

**Epidote**

Epidote is present in B97 and BM572, replacing pyroxene, and as segregations. All epidote \((n=8)\) has high Fe³⁺ contents, which may be related to bulk rock composition, and are similar to epidote values of Coombs et al. (1976), Nakajima et al. (1977) and Deer et al. (1992).
Figures 3.6a,b. Covariations of Si with Al'' and Ti in amphibole. After Raase (1974).
Figure 3.6c. Covariation of NaM4 content in calcic amphibole with Al'' + Fe'' + Ti + Cr. After Laird & Albee (1981).
Figure 3.7a. Clinopyroxene tectonic discrimination diagram. After Nisbett & Pearce (1977).
Figure 3.7b,c. Cr-spinel tectonic discrimination diagrams. After Arai (1992).
Quartz and Calcite
Both quartz and calcite are present as veins and as replacement of plagioclase.

Mica
Sericite is present in BM572 as a replacement of plagioclase.

3.4.2.3 Implications
The petrographic study identified components of an incomplete ophiolite. Pyroxene and chrome-spinel compositions suggest that the ophiolite may have formed in an arc-related setting. The ophiolitic rocks do not have the same metamorphic history as the continental rocks, from which they differ in terms of mineralogy, texture as well as rock chemistry. They lack obvious medium to high grade metamorphic minerals, such as garnet, found in the Sibela Continental Suite; where metamorphic minerals are present they are of lower grade i.e. amphibole, sphene, epidote and chlorite, some of which occur in veins; harzburgites are serpentinised but otherwise unmetamorphosed; the ophiolitic rocks generally lack penetrative fabrics; some rocks display igneous textures (cumulate, gabbroic and high T granoblastic textures interpreted as acquired during cooling).

The Sibela Ophiolitic Suite has not suffered regional metamorphism like the Continental Suite, but locally displays metamorphic effects typical of shear zones, with metamorphic conditions of ~5 kb, ~1000°C. The high temperature metamorphic features are consistent with local metamorphism in an unusual setting (?fracture zone, spreading centre or forearc) although there is nothing diagnostic. Later low T metamorphism may be related to emplacement or young volcanic activity, as discussed below.

3.4.3 Whole Rock Chemistry
Five samples (B32, B97, B103, BM571 and BM572) were selected for whole rock chemical analysis by XRF. Only B97 was in situ, the others are float samples collected from large boulders near outcrops.

3.4.3.1 Major Elements
Only BM572 has >2% ignition loss (LOI) and therefore all rocks are considered fresh on the basis of criteria proposed by Le Bas et al. (1986). All samples are ultramafic, but only B97 and BM572 are ultrabasic. B32 is chemically similar to an olivine gabbro, but the remaining samples differ considerably from gabbros, norites and peridotites of ophiolites or other basic-ultrabasic igneous
Metamorphic Complexes

complexes. B103 and BM571 have SiO₂ contents of ~48-49 wt % whereas B97 and BM572 have <45 wt % SiO₂ but all have high MgO + FeO* + CaO.

Typical peridotites (e.g. Best, 1982; Wilson, 1989) have <2.0 wt% Al₂O₃ and ~40 wt% MgO whereas B97 and B103 have considerably higher Al₂O₃ and lower MgO. B97 is similar to a picrodolerite. Despite the difference in petrography, B103 is similar to BM571, both having high SiO₂, MgO and CaO together with low TiO₂, Al₂O₃, and FeO* contents compared to the other samples, and these resemble olivine pyroxenites. Table 3.5 compares Sibela ophiolite samples with chemically similar ultramafic rocks.

Table 3.5 Comparison of the Sibela Ophiolite major elements with other ultrabasic rocks. Sources are: olivine gabbro (Kuno, 1967); picro-dolerite (Drever & Johnston, 1967); pyroxene hornblendite and olivine pyroxenite (Taylor, 1967).

<table>
<thead>
<tr>
<th>Wt%</th>
<th>B32 schist</th>
<th>Gabbro</th>
<th>B97 peridotite</th>
<th>Picrodolerite</th>
<th>BM572 cumulate</th>
<th>H‘blende</th>
<th>B103 peridotite</th>
<th>BM571 amphibolite</th>
<th>Olivine px-nite</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>45.67</td>
<td>40.73</td>
<td>44.79</td>
<td>44.84</td>
<td>40.04</td>
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<td>49.70</td>
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<td>6.54</td>
<td>9.93</td>
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</tr>
<tr>
<td>FeO*</td>
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<td>14.8</td>
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<tr>
<td>MgO</td>
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<td>CaO</td>
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<td>13.77</td>
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<td>9.36</td>
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<td>11.6</td>
<td>20.06</td>
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</tr>
<tr>
<td>Na₂O</td>
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<tr>
<td>K₂O</td>
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<td>P₂O₅</td>
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<td>-</td>
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<td>-</td>
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<tr>
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<td>2.0</td>
<td>0.74</td>
<td>1.27</td>
<td>0.7</td>
</tr>
</tbody>
</table>

BM572 has a texture and mineral compositions indicating a cumulate origin and least squares mixing calculations demonstrate that compositions of B97, BM572, B103 and BM571 are similar to different cumulate types. BM572 can be modelled by a mixture of amphibole + plagioclase + iron oxide; B97 can be obtained by a combination of amphibole + clinopyroxene + plagioclase; B103 and BM571 are similar and they are mixtures of amphibole and clinopyroxene, with high proportion of amphibole for BM571 and pyroxene for B103.
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B97, B103 and BM571 are therefore also interpreted to be cumulates, formed by fractional crystallisation, which means that their geochemistry does not represent the composition of magmatic liquid, but rather reflects the modal abundance of cumulus phases. Serri (1981) suggested that ophiolites with wide spectrum of fractionated cumulate rocks (such as the Sibela Ophiolite) can be divided into high-Ti and low-Ti types, the former indicating Mid Ocean Ridge (MOR) affinity, whereas the latter could have a MOR, Island Arc (IA) or Supra Subduction Zone (SSZ) origin. Fig.3.8 shows that the Sibela rocks lie near the boundary between the two types of ophiolite.

The dominant cumulus phase in all the cumulates is amphibole (± clinopyroxene), with intercumulus calcic plagioclase and magnetite suggesting crystallisation of magmatic amphibole, probably indicating hydrous conditions. These cumulates differ from oceanic cumulates (e.g. Elthon et al., 1992) in having amphibole, as opposed to olivine, as the primary cumulate phase and are similar to the type III arc cumulates of Beard (1986).

3.4.3.2 Trace Elements

B32 shows an anomalously high concentrations of incompatible elements (i.e. Nb, Th, LIL, HFS, HREE and LREE) and low compatible elements (i.e. Ni and Sc). This is consistent with derivation of B32 from a liquid enriched in incompatible elements, indicating partitioning of incompatible elements into coexisting melt.

Trace element ratios are often used as a means for deciphering magma genesis and ultimately tectonic setting of volcanic rocks. Only B32 is a possible candidate for this type of approach, assuming it is close to the magma composition. It does fall within the compositional limits suggested by Pearce & Cann (1973) for the Ti-Zr-Y method (CaO + MgO < 20 wt%). A comparison of the values of B32 and N-MORB, IAT, Back Arc Tholeiite (BAT) and tholeiitic Ocean Island Basalt (OIB) is provided in Table 3.6.

Both the trace elements and their ratios suggests that B32 is closest to BAT. The difference in Sr and La may be attributed to the mobility of these elements during lower temperature metamorphism. The lower Nb and higher Nd contents may be a reflection of the degree of fractionation or contamination of magma source.

The trace element values of the cumulates are controlled by the original cumulus mineralogy. Within the cumulates there is a trend of decreasing MgO, CaO/Al₂O₃ and increasing HFS (Ti, Zr
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and Y), from B97, BM571, B103 to BM572, indicating increasing fractional crystallisation. B103 and BM571 have a high Sc/Ga ratio (19.9 and 12.8 respectively) compared to the Darvel Bay (~9.5; Omang, 1993) and Halmahera harzburgites (~9.0; Ballantyne, 1990). B97 and BM572 Sc/Ga ratio (7.5 and 7.2) is similar to Halmahera Iherzolite (~7.0; Ballantyne, 1990).

Table 3.6 Comparison of the ophiolite gabbro trace elements and their ratios with rocks from known tectonic settings. Data for comparison are from Wilson (1989).

<table>
<thead>
<tr>
<th>Element (ppm)</th>
<th>B32</th>
<th>N-MORB</th>
<th>IAT</th>
<th>BAT</th>
<th>OIB</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sr</td>
<td>742.4</td>
<td>124</td>
<td>200</td>
<td>212</td>
<td>290</td>
</tr>
<tr>
<td>Rb</td>
<td>6.0</td>
<td>1.0</td>
<td>4.6</td>
<td>6.0</td>
<td>7.5</td>
</tr>
<tr>
<td>Zr</td>
<td>133.7</td>
<td>85</td>
<td>22</td>
<td>130</td>
<td>149</td>
</tr>
<tr>
<td>Nb</td>
<td>2.3</td>
<td>3.1</td>
<td>0.7</td>
<td>8.0</td>
<td>13</td>
</tr>
<tr>
<td>Y</td>
<td>27.3</td>
<td>29</td>
<td>12</td>
<td>30</td>
<td>26</td>
</tr>
<tr>
<td>La</td>
<td>24.1</td>
<td>3.0</td>
<td>1.3</td>
<td>7.83</td>
<td>9</td>
</tr>
<tr>
<td>Nd</td>
<td>31.9</td>
<td>7.7</td>
<td>3.4</td>
<td>13.1</td>
<td>19</td>
</tr>
<tr>
<td>Zr/Rb</td>
<td>22.3</td>
<td>85</td>
<td>4.8</td>
<td>21.7</td>
<td>19.9</td>
</tr>
<tr>
<td>Zr/Nb</td>
<td>58.1</td>
<td>24.3</td>
<td>31.4</td>
<td>16.3</td>
<td>11.5</td>
</tr>
<tr>
<td>Zr/Y</td>
<td>4.9</td>
<td>2.9</td>
<td>1.8</td>
<td>4.3</td>
<td>5.7</td>
</tr>
<tr>
<td>Y/Nb</td>
<td>11.9</td>
<td>9.4</td>
<td>17.1</td>
<td>3.8</td>
<td>2</td>
</tr>
<tr>
<td>La/Nb</td>
<td>10.5</td>
<td>1.0</td>
<td>1.9</td>
<td>1.0</td>
<td>0.7</td>
</tr>
<tr>
<td>Ba/La</td>
<td>5.7</td>
<td>4.0</td>
<td>84.6</td>
<td>9.8</td>
<td>11.1</td>
</tr>
</tbody>
</table>

All cumulates are distinctive in having lower Pb, Cl, Sr and Ba, and higher Ni, Cr (especially B103 and BM571), V (except B97) and Sc than B32. Some depletions may be due to plagioclase (Sr and Ba substituting for Ca) abundance in B32, whereas the enrichment is because of preferential substitution of Ni, V and Sc in the cumulate pyroxene and presence of Cr-spinel in some cumulates.

3.4.3.3 Tectonic Discrimination Diagrams

Several tectonomagmatic diagrams have been constructed to elucidate the tectonic setting of ancient volcanic suites, based on immobile trace element covariations. Although these discrimination diagrams were intended to decipher the formation setting of basaltic rocks, they
may be applied to the Sibela Ophiolite gabbro, assuming it represents a magma composition and the trace elements were immobile during metamorphism.

On the Ti-Zr-Y diagram (Pearce & Cann, 1973), B32 plots in Calc-Alkaline Basalt field (CAB) (Fig.3.9a). On the Nb-Zr-Y diagram (Meschede, 1986) B32 plots in field of VAB and Within Plate Basalt (WPB) (Fig.3.9b). The TiO$_2$-MnO-P$_2$O$_5$ (Mullen, 1983) is the only major element tectonomagmatic diagram; on this diagram, B32 plots in the CAB field (Fig.3.9c).

Pearce (1982) introduced a diagram in which trace element values are normalised against MORB. Using this diagram the ophiolitic rocks can be divided into two groups: B32 and the cumulates (Fig.3.10a). B32 shows LIL enrichment, Nb and Cr depletion, which is characteristic of island arc basalts. The cumulates all have similar patterns: fairly flat Sr-Th distribution; depletion of Th-Y, having a concave-up shape with Nb and Zr as minimas; enrichment of Sc and Cr, except BM572 which is depleted in Cr. Fig.3.10b is a chondrite normalised diagram (Sun, 1980), showing the cumulates, having Ba, Nb, REE (La, Ce and Nd) and HFSE (P, Zr and Ti) depletions and Sr enrichment. B32 is enriched in all elements compared to the other samples and shows Nb depletion, Th enrichment and a concave down Nb-Ti pattern, typical of arc volcanics. All samples are likely to be cogenetic as the pattern for B32 is the mirror image of the cumulates, particularly in LRE, HRE and HFS elements. This suggests that B32 represents a higher level of the ophiolite.

Pearce (1982) used the covariation of immobile trace elements which are unaffected by alteration and AFC processes (e.g Cr against Y and Ce/Sr; Ti/Y against Nb/Y; and Ti against Zr) to characterise volcanic rocks according to their tectonic setting. In these diagrams (Figs.3.11a-d) B32 consistently plots in the VAB field. Fig.3.11e is a similar diagram using Zr/Y against Zr (Pearce & Norry, 1979), showing B32 plotting in field which contains WPB and MORB.

3.4.3.4 Implications
All samples analysed are plutonic ultramafic rocks, interpreted as representing the lower crustal part of an ophiolite sequence (metagabbro and cumulates). Only B32 (metagabbro) is a non-cumulate rock, and from spider and tectonic discrimination diagrams, it most likely formed in an arc setting, similar to many ophiolites. Elemental distribution suggests that B32 is cogenetic with the cumulates.

Despite the different types of cumulates, all of them share amphibole as one of the main cumulus phases, which indicates crystallisation in a hydrous environment, suggesting an arc origin. Within
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Figure 3.8. Covariation of TiO$_2$ and FeO*/MgO. After Serri (1981).

Figure 3.9a-c. Ternary tectonic discrimination diagrams applied to B32. See text for details.
Figure 3.10a-b. Spider diagrams for the Sibela Ophiolitic Suite's cumulates and metagabbro. After Pearce (1982) and Sun (1980).
the cumulates, there is a possible fractionation trend from B97, BM571, B103 to BM572. The east Halmahera ophiolite was formed in a tectonic setting normally associated with initiation of an arc (SSZ; Ballantyne, 1990). The abundance of cumulate rocks suggests that the Sibela ophiolite may be a more mature (fractionated) version of the Halmahera one.

### 3.4.4 Age Determination

Two amphibolites from the oceanic suite yielded K-Ar ages of 46.40 ± 3.50 Ma (BM572) and 90.60 ± 10.90 Ma (B32). The older sample was also analysed using the Ar-Ar step heating technique and showed plateau ages of 38.0 ± 4 Ma and 799.0 ± 5 Ma (Fig.3.12a). Three additional samples were also analysed using the Ar-Ar method. Sample BM571 yielded a plateau age (>50% of total $^{39}$Ar released) of 31.0 ± 6 Ma (Fig.3.12b); B97 produced a plateau age of 85.0 ± 12 Ma (Fig.3.12c) and B103 29.0 ± 9 Ma (Fig.3.12d).

All of the samples analysed with the $^{39}$Ar/$^{40}$Ar technique exhibited U-shape spectra indicative of excess $^{40}$Ar (Lanphere & Dalrymple, 1976; McDougall & Harrison, 1988). Plateau ages were calculated from the minima of the U-shaped spectra, and are older than the time at which the closure temperature was reached (Lanphere & Dalrymple, 1976). Excess argon may be introduced to the system from chlorite and mica formed during healing of post-metamorphic fractures (Blanckenburg & Villa, 1988), which is consistent with mineral chemical studies. The K-Ar dates are interpreted to reflect excess argon and possible multiple thermal events and therefore probably do not record meaningful ages.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Material</th>
<th>Grain size (µ)</th>
<th>$%K$ (1σ error)</th>
<th>Wt for Ar (g)</th>
<th>$^{40}$Ar*/(nl/g, 1σ error)</th>
<th>$^{40}$Ar$_{an}$</th>
<th>Age (Ma, 2σ error)</th>
</tr>
</thead>
<tbody>
<tr>
<td>B32</td>
<td>amph</td>
<td>125-250</td>
<td>0.467 ± 3.43%</td>
<td>0.2079</td>
<td>1.6875 ± 5.16%</td>
<td>83.34</td>
<td>90.6 ± 10.9</td>
</tr>
<tr>
<td>BM572</td>
<td>amph</td>
<td>125-250</td>
<td>0.231 ± 3.46%</td>
<td>1.1672</td>
<td>0.4218 ± 1.59%</td>
<td>54.95</td>
<td>46.4 ± 3.5</td>
</tr>
</tbody>
</table>

The plateau ages give an average young plateau age of 25-37 Ma and an old plateau age of 94-97 Ma. The young plateau age is interpreted to be related to the Oligocene volcanism between ~40 and ~25 Ma (see Chapter Six: The Tawali Formation), whereas the old plateau age can be correlated with the thermal events associated with volcanic activity affecting the eastern Halmahera Ophiolite (80-94 Ma; Ballantyne, 1990). The older plateau age strongly supports the link between the east Halmahera and the Sibela Ophiolites.
Figure 3.12a-d. Ar-Ar step heating spectra for B32, B97, B103 and BM571.
Table 3.8 Summary of the Ar-Ar step heating results from the Sibela Metamorphic Complex, Ophiolitic Suite.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Material</th>
<th>K (%)</th>
<th>Plateau Age (Ma; 1σ error)</th>
<th>Plateau Age (Ma; 1σ error)</th>
<th>Integrated Age (Ma; 1σ error)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BM57</td>
<td>Amphibole</td>
<td>0.12</td>
<td>31 ± 6 (steps 3,4,5)</td>
<td>56.0 ± 8.0</td>
<td></td>
</tr>
<tr>
<td>B32</td>
<td>Amphibole</td>
<td>0.38</td>
<td>38 ± 4 (steps 3,4,5,6)</td>
<td>99 ± 5 (steps 7,8)</td>
<td>107 ± 6.0</td>
</tr>
<tr>
<td>B97</td>
<td>Amphibole</td>
<td>0.12</td>
<td>85 ± 12 (steps 3,4,5,6)</td>
<td>98.0 ± 12.0</td>
<td></td>
</tr>
<tr>
<td>B103</td>
<td>Amphibole</td>
<td>0.08</td>
<td>29 ± 9 (steps 5,6)</td>
<td>46.0 ± 12.0</td>
<td></td>
</tr>
</tbody>
</table>

3.5 THE SALEH METAMORPHIC COMPLEX

There are outcrops of metamorphosed sedimentary and volcanic rocks on the small Saleh Islands and on the SE tip of north Bacan which are physically separated and different in character from metamorphic rocks of the Sibela Complex. Hall et al. (1988a) suggested that the Saleh metamorphic rocks could represent parts of the upper levels of an ophiolite complex and be equivalent to the Sibela ophiolitic rocks. Because of the differences these rocks are here assigned to a separate unit, the Saleh Metamorphic Complex.

3.5.1 Lithofacies

On aerial photographs, the Saleh Metamorphic Complex is characterised by low, subdued hills, with few rivers. This complex consists of metavolcanic rocks of prehnite-pumpellyite facies, separated by a fault from phyllites and quartzite. The latter show polyphase folding with foliations dipping NE at 30-40°.

Type Locality

The type locality for the metabasites is P. Saleh Lamo where there are outcrops of unfoliated but thoroughly jointed basic metavolcanic rocks with quartz veins. Samples collected from the central part of the island, appear to be amygdaloidal pillow basalt with green psammitic interpillow material overlain by phyllites. Pillows are right way up and deformed in places (Plate 3.6a).

The adjacent island of Saleh Kecil is the type section for the metasedimentary rocks where there are sequences of well foliated, friable graphitic phyllites and quartzite, interbedded with highly foliated metavolcanic rocks. Kink folding occurs in the phyllites. Cutting through the quartzites are quartz and epidote veins.
Plate 3.6a. Deformed pillow lavas at P. Saleh Kecil (from Hall et al., 1988a).

Plate 3.6b. Foliated metasedimentary rock associated with pillow lavas at P. Saleh Kecil (from Hall et al., 1988a).
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The metasedimentary rocks have a pronounced foliation (Plate 3.6b) whereas the metavolcanic rocks, especially in the larger outcrops, are typically massive. This may be due to the resistance of the latter to fabric development and to the mineralogical differences between the two types of rocks. In the southern part of Saleh Kecil, this complex is intruded by amphibole-rich diorite (Chapter Nine).

Lower and Upper Boundaries
Contacts with other formations cannot be seen. However aerial photographic interpretation suggests that the outcrops in northern P. Bacan are in fault (possibly steep normal) contact with the Bacan Formation.

Thickness
Based on the height of P. Saleh Lamo, where nearly horizontal metabasite rocks are exposed, the thickness of this complex is estimated to be at least 100 m.

Lateral Extent and Variations
This metamorphic complex is seen in both P. Saleh Lamo and P. Saleh Kecil and the southeastern tip of northern P. Bacan, where there are well foliated calcareous metaturfs, phyllites and metabasites. These low grade metamorphic rocks are cut by veins of calcite and chalcedony or quartz. Pillowed metabasites are observed at Tg. Semola in Halmahera. Fig. 3.1 shows the distribution of the Saleh Metamorphic Complex in the region.

3.5.2 Metabasite Petrography and Mineral Chemistry
3.5.2.1 Petrography
Unless stated, all modes are percentages based on visual estimation. The metabasites are composed of albite, clinopyroxene, and titanomagnetite phenocrysts in a glassy matrix. Intergrowths of pyroxene and plagioclase phenocrysts are locally present. Pumpellyite, prehnite, amphiboles and occasionally epidote replace pyroxene. Pumpellyite, chlorite and smectite replace the matrix. Trachytic and seriate textures may be preserved (Plate 3.7a,b). Veins are filled by calcite and quartz whereas the amygdales are filled with quartz, calcite and chlorite. Foliation in all samples is defined by alignment of elongate amphiboles. Interpillow psammites are dominated by devitrified glass, with some plagioclase and opaque detritus. Veins cutting through these samples are filled with calcite (Plate 3.7c,d).
Plate 3.7a,b. PPL and XPL of Saleh metabasite (SM10); note bent plagioclase laths, opaque phenocrysts and epidote replacing pyroxene.

Plate 3.7c,d. PPL and XPL of Saleh psammite (SR15); contact between fine and coarse grained psammite. Plagioclase and opaque grains in a micritic matrix. Scale for all photomicrographs = 0.5 mm.
3.5.2.2 Mineral Chemistry

Five representative metabasites (SM10, SM12, SR13, SR17, SR20) were analysed for their mineral chemistry contents, to assess the metamorphic conditions and tectonic setting of this complex. Fig. 3.2 shows the locality of samples used for analytical studies.

**Plagioclase**

Plagioclase is a ubiquitous phenocryst and microphenocryst phase, forming euhedral laths up to 0.4 mm across, with complex twinning. Zoning was not observed and most laths appear to be strained and occasionally bent. In places these contain opaque, pyroxene and pumpellyite inclusions.

The phenocryst compositions (Fig. 3.13a) are mostly An$_{0.5}$Ab$_{0.4}$Or$_{0.1}$ (n=19), except SR17 which contain three analyses of An$_{0.20}$. The microphenocrysts are An$_{0.2}$Ab$_{0.97}$Or$_{0.1}$ (n=11). Compositional consistency and lack of zoning suggest that the plagioclase has been albitised from an originally higher An composition. The higher An analyses are from microphenocrysts forming the trachytic groundmass texture and may represent incompletely albitised calcic plagioclase.

**Pyroxene**

Clinopyroxenes are abundant, occurring both as phenocrysts and as inclusions in plagioclase. Phenocrysts are anhedral-subhedral. Locally these form glomerocrysts and occasionally they are corroded. Neither twinning nor zoning is observed. They are mostly fresh although in places they are replaced by epidote, prehnite, pumpellyite and amphibole.

The composition of the phenocrysts is En$_{36.49}$Fs$_{10.24}$Wo$_{15.46}$ (average En$_{49}$Fs$_{51}$Wo$_{42}$; n=16), and one inclusion analysis is En$_{35}$Fs$_{23}$Wo$_{41}$. They all plot in the field of augite and salite. Fig. 3.13b is a tectonic discrimination diagram modified from Nisbet & Pearce (1977) suggesting the pyroxene phenocrysts are of VAB origin.

**Fe-Ti Oxide**

Oxides occur mostly as phenocrysts. They are predominantly anhedral with occasional small squared euhedral crystals. All analyses fall between magnetite and ulvöspinel. The amount of TiO$_2$ varies widely (0-16 wt%) even within a sample. There is no systematic variation observed in terms of crystal shape, size or texture. Inclusions tend to be higher in TiO$_2$ compared to the phenocrysts.
Figure 3.13a Ternary diagram showing all plagioclase compositions from the Saleh Metabasite.

Figure 3.13b. Clinopyroxene tectonic discrimination diagram for Saleh Metabasite. After Nisbett & Pearce (1977).
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Glass
The matrix is mostly devitrified glass which has changed to quartz. Relict glasses (SR17GM1-4) are rich in SiO₂ and Al₂O₃ with variable amounts of CaO, Na₂O and K₂O. The presence of 'fresh' glass is interpreted to reflect incomplete alteration, consistent with the sample's preserved trachytic texture and mineral assemblage (quartz + chlorite + calcite).

Quartz and Calcite
Quartz is a ubiquitous secondary mineral replacing glass, plagioclase and filling veins. Calcite is observed in SM10 and SR17. It occurs either in veins or replacing plagioclase.

Chlorite
Chlorite, in SM12 and SR17, replaces glass and plagioclase. It is poorly crystalline, greenish-yellow, with anomalous interference colours. Chlorite is distinguished from smectite by its low SiO₂ content (27-37 wt%) and lack of K₂O. It is similar to those reported by Offler & Aguirre (1984).

Smectite
Smectite occurs only in SM12 replacing the groundmass. It is cryptocrystalline, clear in PPL, yellow-brown in XPL with a moderate birefringence. Following Andrew (1980), it is classified as K-rich celadonite.

Pumpellyite
Pumpellyite occurs in SM10, SM12, SR13, replacing pyroxene which often occur as inclusions in plagioclase, and in SM10 and SR20 replacing groundmass. It occurs as radial, high birefringence grains. All analyses are high Fe-variety (>10 wt% FeO*), similar to those reported by Arvin (1982), with lower Al₂O₃ contents than those of Offler & Aguirre (1984). These plot in Kawachi's (1975) Upper Wakatipu III field.

Epidote
Only SR20 contains epidote, replacing pyroxene and filling veins. It is small, anhedral-subhedral, clear in PPL, while in XPL it is yellow-green with high birefringence. No chemical difference was detected between the two types. Distinction from pumpellyite is based on the lack of MgO in epidote, amount of Fe₂O₃ (epidote: 11-13 wt%, pumpellyite: 13-21 wt%), and total analysis (epidote ~95 wt%, pumpellyite ~90 wt%). Epidotes are similar to those reported by Deer et al. (1992), Coombs et al. (1976), Nakajima et al. (1977) and Offler & Aguirre (1984).
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**Prehnite**
Prehnite was found only in SM12, coexisting with pumpellyite and replacing pyroxene. It is tabular with moderate relief and strong birefringence.

**Sphene**
Sphene occurs as a replacement of Fe-Ti oxide in SR20. It has a high Al$_2$O$_3$ (3.75 wt%) and low TiO$_2$ contents (32.8 wt%) compared those of Offler & Aguirre (1984), but similar to those of Coombs *et al.* (1976). The presence of sphene indicates very low $X_{CO_2}$ in the metamorphic fluids (Coombs *et al.*, 1976).

**Amphibole**
Amphiboles are of secondary origin, replacing pyroxene. They are actinolite, actinolitic hornblende and magnesio-hornblende, similar to those of Coombs *et al.* (1976).

3.5.2.3 Implications
Textures and mineral chemistry suggest that the Saleh metabasites have arc characteristics. Although there are small differences in the secondary minerals, an assemblage of albite + quartz + calcite ± (chlorite, smectite, pumpellyite, prehnite, epidote, actinolite, sphene) is recognized. The reaction pumpellyite + chlorite + quartz = chabazite + tremolite + H$_2$O and pumpellyite + heulandite + quartz = epidote + actinolite + chlorite + H$_2$O defines the transition between pumpellyite-actinolite to greenschist facies metamorphism (Liou *et al.*, 1987). The rocks of the Saleh Metamorphic Complex therefore indicates conditions transitional between greenschist and pumpellyite-actinolite facies of metamorphism with temperatures of 250-360°C and pressures of ~4 kb, suggested by the assemblage of albite + quartz + chlorite + pumpellyite + epidote + actinolite + sphene (Coombs *et al.*, 1976; Liou *et al.*, 1987).

3.5.3 Metabasite Whole Rock Chemistry
Four metabasite samples (SM10, SM12, SR17 and SR20) were analysed using the XRF technique to determine the tectonic setting of the complex. Only sample SM12 was not *in situ*, but was collected from an angular block ~1 m across, indicating that it had not travelled far. Triplicate analyses of sample SM10 were performed to monitor the reproducibility of the method (App. B).

3.5.3.1 Major Elements
The seven analyses shows that the LOI ranges from 2.52-3.28 wt% indicating alteration (Le Bas *et al.*, 1986). Plots of alkalis versus silica indicate that they are calc-alkaline andesites (Fig.3.14a,b;
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Le Bas & Streckeisen, 1991; Basaltic Volcanism Study Project, 1981). Table 3.9 is a comparison of the major elements of the Saleh Complex with a High-K Calc-Alkaline andesite from Rinjani Mountain in the Sunda Arc (HKCA), an IAT from the South Sandwich Arc (data from Wilson, 1989) and a Back Arc Basin Basalt (BABB) from the Mariana Trough (DSDP analysis 456A-3, Wood et al., 1981).

The Saleh metabasites are comparable to HKCA and IAT, with a slightly higher Na$_2$O values. The K$_2$O content of the Saleh Complex varies widely, which may be an indication of the degree of alteration that they have suffered. Although the Saleh metabasites are more acidic than the BABB, they have similar Al$_2$O$_3$ and MnO contents. The lower MgO and CaO, and higher FeO$^*$ contents of the Saleh metabasite indicate that it is more differentiated than BABB.

Table 3.9 Comparison of the Saleh Complex major elements with other andesites from known tectonic settings; (range of values are in brackets).

<table>
<thead>
<tr>
<th>Wt%</th>
<th>SALEH</th>
<th>HKCA</th>
<th>IAT</th>
<th>BABB</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO$_2$</td>
<td>55.24 (53.07-58.23)</td>
<td>55.49</td>
<td>57.10</td>
<td>50.7</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>0.90 (0.569-1.036)</td>
<td>0.91</td>
<td>0.92</td>
<td>1.18</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>16.32 (15.37-18.34)</td>
<td>18.45</td>
<td>16.15</td>
<td>16.6</td>
</tr>
<tr>
<td>FeO$^*$</td>
<td>9.42 (7.02-10.31)</td>
<td>8.32</td>
<td>9.13</td>
<td>8.87</td>
</tr>
<tr>
<td>MnO</td>
<td>0.154 (0.129-0.185)</td>
<td>0.16</td>
<td>0.18</td>
<td>0.16</td>
</tr>
<tr>
<td>MgO</td>
<td>3.40 (2.56-4.84)</td>
<td>3.10</td>
<td>3.38</td>
<td>5.23</td>
</tr>
<tr>
<td>CaO</td>
<td>6.66 (3.78-9.33)</td>
<td>7.47</td>
<td>8.47</td>
<td>11.23</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>5.85 (3.49-7.46)</td>
<td>4.09</td>
<td>3.56</td>
<td>3.27</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>0.89 (0.095-2.513)</td>
<td>1.60</td>
<td>0.32</td>
<td>0.62</td>
</tr>
<tr>
<td>P$_2$O$_5$</td>
<td>0.286 (0.106-0.351)</td>
<td>0.28</td>
<td>0.15</td>
<td>0.11</td>
</tr>
</tbody>
</table>

3.5.3.2 Trace Elements

The major element results suggest an arc-related origin. To clarify this, comparison of selected trace elements ratios with IAT, HKCA and MORB (data source as above; Wilson, 1989) and Back Arc Basin Basalt (data as above, Wood et al., 1981) is shown in Table 3.10.
Figure 3.14a. Saleh Metabasite in a volcanic rock classification diagram based on total alkali and silica contents (after Le Bas & Streckeisen, 1991).

Figure 3.14b. Saleh Metabasite in a volcanic rock classification diagram based on potassium and silica contents (after Basaltic Volcanism Study Project, 1981).
Metamorphic Complexes

Table 3.10 Comparison of Saleh Complex trace element ratios with rocks from known tectonic settings and the Sibela Ophiolite metagabbro. SM10 values are the average of the three analyses.

<table>
<thead>
<tr>
<th></th>
<th>SM10</th>
<th>SM12</th>
<th>SR17</th>
<th>SR20</th>
<th>HKCA</th>
<th>IAT</th>
<th>BABB</th>
<th>MORB</th>
<th>B32</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zr/Y</td>
<td>4.1</td>
<td>3.6</td>
<td>3.2</td>
<td>4.4</td>
<td>4.5</td>
<td>1.9</td>
<td>3.2</td>
<td>2.9</td>
<td>4.9</td>
</tr>
<tr>
<td>Zr/Rb</td>
<td>79.8</td>
<td>4.8</td>
<td>1.1</td>
<td>6.8</td>
<td>3.8</td>
<td>6.4</td>
<td>10.5</td>
<td>85</td>
<td>22.3</td>
</tr>
<tr>
<td>Cr/Y</td>
<td>0.2</td>
<td>0.3</td>
<td>0.5</td>
<td>0.2</td>
<td>0.9</td>
<td>1.9</td>
<td>4.1</td>
<td>-</td>
<td>0.9</td>
</tr>
<tr>
<td>Sr/Nb</td>
<td>19.3</td>
<td>56.2</td>
<td>216.4</td>
<td>110.1</td>
<td>216.5</td>
<td>-</td>
<td>-</td>
<td>40</td>
<td>322.8</td>
</tr>
</tbody>
</table>

SM10 appears to be different to the other Saleh rocks, particularly in the Zr/Rb and Sr/Nb ratios, which is unexpected in light of similarities in petrography and mineral chemistry. The Zr/Y ratio of all Saleh samples is between that of HKCA and BABB. The Zr/Rb ratio of SM10 is closest to MORB, whereas the others are similar to HKCA and IAT. The lower Cr contents of the Saleh samples result in the lower Cr/Y ratio, and this may reflect a higher degree of fractionation of the Saleh Samples. The Sr/Nb ratios of the Saleh rocks are higher than MORB, but there is no Nb data available for IAT and BABB. The Sr concentration of SM10 (~68 ppm) is lower than anything else; the Sr concentration of the other Saleh samples (140-363 ppm) are within range with HKCA, IAT, BABB and MORB. The Sibela metagabbro clearly differs geochemically from the Saleh metabasite, particularly in the Zr/Rb, Cr/Y and Sr/Nb ratios.

3.5.3.3 Tectonic Discrimination Diagrams

On a Ti-Zr-Y discrimination diagram of Pearce & Cann (1973) the Saleh rocks plot in the Ocean Floor Basalt (OFB), CAB and LKT fields (Fig.3.15a). The Nb-Zr-Y diagram (Meschede, 1986) shows that the Saleh metabasites scatter in the fields of N-MORB, VAB and Within Plate Basalt (WPT) (Fig.3.15b). On the TiO$_2$-MnO-P$_2$O$_5$ (Mullen, 1983), the Saleh metabasites plot in the IAT and CAB fields (Fig.3.15c).

On the MORB normalised diagram (Pearce, 1982), the Saleh Complex shows an arc character (enrichment in LIL relative to HFS, depletion in Nb and Cr). However, the flat Th-Cr pattern and flat LIL pattern for sample SM10 suggests a MORB-like character (Fig.3.16a). When normalised against chondrite (Sun, 1980), the Saleh metabasites are enriched in Rb, Th and K; and depleted in Nb and La, typical of island arc rocks. The elemental distribution from HFSE to REE, however, shows a curving down pattern, characteristic of MORB rocks (Fig.3.16b).
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Figures 3.15a-c. Ternary tectonic discrimination diagrams for Saleh Metabasite. See text for details.

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Figure 3.16a-b. Spider diagrams for the Saleh Metabasite. After Pearce (1982) and Sun (1980).
Metamorphic Complexes

Figs. 3.17a-e show covariations of immobile trace elements (Cr against Y and Ce/Sr; Ti/Y against
Nb/Y; and Zr against Ti and Zr/Y) in which the Saleh Complex plots in the VAB and MORB
fields. Hawkins & Evans (1983) proposed a FeO*/MgO versus wt% TiO₂ diagram to distinguish
back-arc basalt from arc basalts. This diagram will reflect the amount of fractionation with the
assumption that the back-arc basalt will be less differentiated than the arc basalt. On this diagram,
the Saleh Complex plots in the island arc field (Fig. 3.17f).

3.5.3.4 Implications

Major element and trace element ratios suggest an arc character with some MORB and BABB
affinities. Tectonic discrimination diagrams show the Saleh Complex plotting in both MORB and
island arc fields. When normalised against MORB and chondrite, the Saleh Complex shows both
island arc and MORB aspects. The Saleh Complex is therefore, interpreted as erupted in a back-arc
setting, which may include rocks with Island Arc, MORB and BABB characteristics (cf. Saunders
& Tarney, 1984).

3.5.4 Metabasite Age Determination

K-Ar analyses of SM12 yielded an age of 12.0 ± 0.6 Ma, with a duplicate analysis of 10.9 ± 0.8
Ma. The two ages overlap between 11.4-11.7 Ma. The grade of metamorphism and petrographic
differences from Neogene volcanic rocks suggest that the true age is older than Neogene.
Figs. 3.18a,b are isochron diagrams plotted for the Saleh Metamorphic Complex and the Saleh
Intrusive (Chapter Nine). These show that the two groups are related by a thermal event at 11.6-
12.2 Ma, indicating a resetting of the Saleh Metamorphic ages. This thermal event is interpreted
to be related to the initiation of the Halmahera Arc. The age of the complex is still unknown.

Table 3.11 Summary of the K-Ar results from the Saleh Metamorphic Complex.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Material</th>
<th>Grain size (µm)</th>
<th>% K (1σ error)</th>
<th>Wt for Ar (g)</th>
<th>¹⁸⁷Ar* (nl/g, 1σ error)</th>
<th>¹⁸⁷Ar on (%)</th>
<th>Age (Ma, 2σ error)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SM12</td>
<td>whole rock 125-250</td>
<td>0.842 ± 1.00%</td>
<td>0.9867</td>
<td>0.3938 ± 2.20%</td>
<td>65.91</td>
<td>12.0 ± 0.6</td>
<td></td>
</tr>
<tr>
<td>SM12 dp</td>
<td>whole rock 125-250</td>
<td>0.934 ± 1.00%</td>
<td>1.9346</td>
<td>0.3977 ± 3.46%</td>
<td>76.53</td>
<td>10.9 ± 0.8</td>
<td></td>
</tr>
</tbody>
</table>
Figure 3.17a-e. Trace element covariations used to decipher the tectonic setting of the Saleh Metabasite. A-D after Pearce (1982), E after Pearce & Norry (1979).

Figure 3.17f. Covariations of TiO$_2$ and FeO*/MgO to distinguish back-arc from arc setting for the Saleh Metabasite (after Hawkins & Evans, 1983).
Figure 3.18a-b. K-Ar isochron diagrams for the Saleh Metabasite (A after Hayatsu & Carmichael, 1970; B after Harper, 1970).
3.5.5 Metasedimentary Rocks

Petrographically these rocks can be divided into (summary in Appendix E):

1. Phyllites:
   - Foliated metatuff: composed primarily of green amphibole + quartz (~35% each) + epidote (in veins) ± brown biotite ± plagioclase. Grains are angular and poorly sorted, of tuffaceous origin. Quartz appears to be both of detrital and authigenic origin, locally filling the veins (Plate 3.8a,b).
   - Calc-quartz-feldspathic phyllite: calcite + quartz + plagioclase + smectite ± iron ore.
   - Foliated calcareous phyllite: calcite + quartz + amphibole + iron ore porphyroblasts in a micritic matrix.

2. Schist:
   - Augen plagioclase + quartz + brown biotite + iron ore (+ calcite + retrograde chlorite).

3. Calc-quartz-feldspathic gneiss:
   - Alternating layers of epidote-rich quartzo-feldspathic gneiss, with distinct quartz horizons, and amphibole-rich quartzo-feldspathic gneiss. Strong foliation defined by alignment of brown biotite and muscovite in matrix of plagioclase and quartz (Plate 3.8c,d).

These are similar to rocks of the Sibela Continental Suite, particularly in having brown metamorphic biotite which, throughout the whole region, is found only in the Sibela Continental rocks. The presence of biotite indicates a higher grade of metamorphism (above greenschist facies) than that of the metabasite.

Although petrographically and mineralogically the metasedimentary rocks can be correlated with the Sibela Continental Suite, field relations are ambiguous. No contact between the metasedimentary and metabasite rocks was observed.

3.6 SYNTHESIS AND TECTONIC SIGNIFICANCE

3.6.1 The Sibela Metamorphic Complex, Continental Suite

Major and trace element studies have established that the protoliths are pelitic, with minor psammite and marl. Tectonic discrimination diagrams suggests that the original rocks were deposited in a continental arc setting.

High temperature, medium pressure, Barrovian type, dynamothermal metamorphism of upper amphibolite facies is inferred from the mineralogy. This type of metamorphism requires temperatures of ~600°C and pressures of ~5 kb, indicating burial depths of ~15 km. Since the Sibela Mountains presently reach a height of 2100 m above sea level, this represents a minimum uplift of 17 km.
Plate 3.8a, b. PPL and XPL of Saleh phyllite (SM5); amphibole and opaque in a matrix of brown biotite and quartz. Fabric is defined by alignment of mafic minerals.

Plate 3.8c, d. PPL and XPL of Saleh gneiss (BA27); epidote and quartz porphyroblasts in a plagioclase, quartz, brown biotite and muscovite matrix. Note presence of amphibole-rich layers on the top and bottom of photomicrographs. Scale for all photomicrographs = 0.5 mm.
Detritus of the Sibela Metamorphic Complex is found only in the Quaternary deposits, indicating that it has become available for erosion only recently, suggesting significant and extremely fast rates of uplift (~1 mm/yr).

The continental suite is thought to originate from the Australian continental margin (e.g. Hamilton, 1979), although this is based on regional stratigraphic arguments. Metamorphic rocks attributed to the Australian basement are known elsewhere in the region (e.g. Bird's Head, Misool, Obi, Banggai-Sula, Seram and Buru), but high grade rocks similar to those of the Sibela Complex have not been reported. Pigram & Symonds (1991) suggested that all continental fragments represent microcontinents rifted from the Australian craton during Late Mesozoic-Early Tertiary. In the Bacan Region, there is no evidence to support or refute this suggestion.

From geological mapping all that can be said is that the Sibela continental rocks are pre-Quaternary in age. Radiometric dating proved to be a vain effort, as all samples yielded very young ages. These are interpreted to be the result of recent interaction with hydrothermal fluids and this suggestion may account for other unusually young ages from high-grade metamorphic complexes in the region.

The simplest interpretation is that the Sibela Continental Suite is of Australian origin as there is no other source for continental rocks. The original protoliths could represent part of old (?Palaeozoic or older) active margin. During the Mesozoic and Tertiary, these rocks probably formed part of the passive margin of north Australia or fragments rifted from Australia during Gondwana breakup.

3.6.2 The Sibela Metamorphic Complex, Ophiolitic Suite

The northern Sibela Mountains includes ultrabasic and basic plutonic rocks interpreted to be part of a dismembered ophiolite. Pyroxene and chrome spinel mineralogy suggest an island arc or intraplate origin. Petrographical and geochemical analyses demonstrate that most of the samples analysed are cumulates with amphibole as the main cumulus phase, indicating crystallisation in a hydrous environment, implying an arc origin. Tectonic discrimination methods based on trace element analyses of one gabbro indicate a volcanic arc origin.

Some rocks display complex magmatic textures suggesting contemporaneous intrusion and deformation. Most of the rocks are unmetamorphosed although locally there are mylonitic rocks with strong penetrative fabrics. Amphibole and plagioclase geothermometry and geobarometry
Metamorphic Complexes

indicate metamorphic conditions of ~5 kb and ~1000°C. Overall, the field, textural and mineral chemical evidence indicates metamorphism was not on a regional scale and was related to ductile deformation of hot rocks at or near their place of formation, and local recrystallization in shear zones.

Radiometric dating yielded Oligocene-Miocene (25-37 Ma) and Middle Cretaceous (94-97 Ma) Ar-Ar plateau ages. Both of these ages can be correlated with ages recorded by the Philippine Sea Plate, implying that the ophiolitic rocks have been part of the Philippine Sea Plate since at least the Middle Cretaceous. The Oligocene-Miocene ages are interpreted as due to reheating during the ~40-25 Ma Oligocene volcanism and extension event (see Chapter Six). The Middle Cretaceous ages can be correlated with the arc and forearc thermal events recorded radiometrically (80-94 Ma; Ballantyne, 1990) and stratigraphically (Hall et al., 1988a, 1991) by the eastern Halmahera Ophiolite.

3.6.3 Juxtaposition of Continental and Ophiolitic Rocks

Differences in metamorphic character compared to the ophiolite suite indicate metamorphism occurred before amalgamation of continental and ophiolitic rocks. The continental suite is interpreted to be part of the Australian Plate, whereas the ophiolitic rocks represented the Philippine Sea Plate. Regional mapping of the region shows that since the Early Miocene, rocks from these two plates have the same geological history, indicating an Early Miocene collision (Hall et al., 1992; Chapters Four-Six). Morris et al. (1983), deduced the presence of continental crust beneath southern Bacan based on the $^{87}$Sr/$^{86}$Sr and $^{207}$Pb/$^{204}$Pb ratios of Quaternary volcanic rocks. Isotopic evidence shows that Neogene volcanic rocks from south Bacan have a continental signature (E.Forde, 1993, pers.comm), indicating the presence of continental suite in the Bacan region by the Late Miocene (see Chapter Eight: The Kaputusan Formation).

There are three possibilities for the mode of juxtaposition of the continental and ophiolitic rocks: [1] the continental rocks were part of the northern Australian Margin, translated by young strike slip faulting (Sorong Fault Zone) into the region; [2] the Sibela Complex represents a suture zone due to collision between the Australian and Philippine Sea Plates; [3] is a combination of [1] and [2] in which the continental rocks collided with the Philippine Sea Plate and were subsequently translated by strike-slip faulting as part of a composite fragment to their present position. These interpretations will be discussed more fully in Chapter Eleven.
3.6.4 The Saleh Metamorphic Complex

The Saleh Metamorphic Complex includes calc-alkaline metabasite and foliated metasedimentary rocks. Mineral chemistry and trace element geochemistry studies indicate that the metabasite is an arc related sequence, possibly formed in a back-arc. The metabasite has been affected by lower greenschist to upper prehnite-pumpellyite metamorphism, requiring metamorphic conditions of \( \sim 250-360^\circ C, \sim 4 \text{ kb} \).

The Saleh metabasite differs from the Sibela Ophiolite in the following ways: [1] geothermometry indicates temperatures which are considerably lower than that of the Sibela Ophiolite; [2] the Sibela Ophiolite has a strong penetrative fabric in some rocks showing polyphase deformation and recrystallization; in comparison, the Saleh metabasites lack any fabric; [3] the Sibela Ophiolite is dominated by lower crustal rocks whereas the Saleh metabasite has an upper crustal origin; [4] the trace element chemistry, in particular the trace element ratios, is different; [5] the gravity anomaly map shows a high positive anomaly over the Saleh Islands, whereas over the Sibela Complex there is a low positive anomaly (Hall et al., 1992). This evidence imply that the metabasite and the Sibela Ophiolite are not cogenetic, contrary to the view of Hall et al. (1988a).

Similarities in lithological character, grade of metamorphism and whole rock chemistry, suggest the Saleh Metamorphic Complex may be related to the Bacan (Chapter Four) and the South Bacan Formations (Chapter Five). Radiometric dating, however, has not proved this as K-Ar ages of the Saleh Complex, the Bacan and the South Bacan Formations have all been affected by a thermal event at \( \sim 12 \text{ Ma} \), related to the initiation of the Halmahera Arc.

The metasedimentary rocks are highly foliated and show a higher grade of metamorphism than the metabasite. These have a similar character to the Sibela Continental Suite, although they lack typical medium grade metamorphic minerals such as garnet, staurolite and kyanite. They are dominated by phyllites, suggesting a lower grade of metamorphism compared to the Sibela Continental Suite which is dominated by schists and gneisses.

The Saleh Metamorphic Complex is not strictly a complex, but is probably the product of juxtaposition of Early Miocene or older arc rocks with rocks of continental affinity (Sibela Continental Suite).
CHAPTER FOUR
THE BACAN FORMATION

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4.3 AERIAL PHOTOGRAPHY
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CHAPTER FOUR
THE BACAN FORMATION

4.1 INTRODUCTION
This formation consists of interbedded volcaniclastic sandstones and siltstones, chert and mudstone, and feldspar-phyric volcanic rocks with associated breccio-conglomerates. It is metamorphosed to greenschist - prehnite-pumpellyite facies and contains areas of quartz mineralisation. This chapter is concerned with lithological and petrographical observations of sedimentary rocks, the petrology and mineral chemistry of the volcanic rocks, and age determination and structural evidence from the formation.

4.2 SYNONYMY
The term Bacan Formation was introduced by Yasin (1980) without a well-defined type locality. The work in the Halmahera region has demonstrated that rocks mapped previously as Bacan Formation include volcanic and volcaniclastic rocks which are of very different character and age (Hall et al., 1988b; Hall et al., 1992). Even within Bacan itself dating work has shown that similar rocks forming the basement of north and south Bacan include rocks of different ages (see Chapter Five: The South Bacan Formation). Since much of northern Bacan is underlain by rocks which have a uniform character, the Bacan Formation is redefined with north Bacan as the type area.

The Bacan Formation (Fig. 4.1) as defined here corresponds approximately to the areas mapped by Silitonga et al. (1981) as the Older Volcanic Rocks and Older Sedimentary Rocks.

4.3 AERIAL PHOTOGRAPHY
On aerial photographs, this formation is characterised by rugged, pointed, high hills forming continuous ridges. They have multiradial - rectangular drainage patterns. Due to similar hardness the different components of the Bacan Formation are indistinguishable on aerial photographs.

4.4 LITHOFACIES
4.4.1 Type Area
In the type locality, the upper part of S. Amasing, there are feldspar-phyric basaltic andesites interbedded with chert and indurated volcaniclastic turbidites. Grey-green volcanic rocks contain plagioclase, relict clinopyroxene, epidote and chlorite. There are bedded red cherts ~7 km from Amasing Village. They are succeeded upstream by green volcaniclastic sedimentary rocks consisting of extremely well lithified sandstone and siltstone containing grains of basic igneous composition. Sandstones show normal grading and are generally fine grained. Carbonate was
Figure 4.1 Distribution of the Bacan Formation in the Bacan Region.
absent in all samples collected, except near to the contact with the Ruta Limestone. Low grade metamorphism is indicated by the presence of epidote, chlorite and pumpellyite. All of the outcrops are pervasively jointed. In many localities, the formation is cut by diorite sills and dykes. Details of the intrusive rocks are provided in Chapter Nine (the Nusa Babi Monzodiorite). Late-stage pyrite, galena, sulphur and quartz mineralisation occurs near the contact with the intrusive bodies.

**Feldspar-Phyric Basaltic Andesite**

The feldspar-phyric basaltic andesites are ubiquitous throughout north Bacan, and in general they are bedded on a dm-m scale. At the type locality in S. Amasing, plagioclase phenocrysts form laths and are up to 10 mm across. The matrix is a dark glass, microcrystalline or cryptocrystalline material. Locally there are zeolite-filled amygdales, usually about 5-10 mm in diameter. These volcanic rocks are otherwise massive and lack any internal structures (Plate 4.1a). Exposures are typically pervasively jointed, with feldspar veins on a millimetre scale filling some of the cracks.

A peculiar feature of the feldspar-phyric volcanic rocks is a "microconglomerate" appearance, seen best on weathered surfaces. The rock appears to be a micro-conglomerate, although both the "clast" and "matrix" are of the same composition, with the matrix being more epidotised and the clasts being feldspar rich. The texture may be due to hydrothermal alteration.

Pudjowalujo & Surjono (1982) and Pudjowalujo & Bering (1982) reported the occurrence of propylitised volcanic rocks with sericite and silica alteration. These rocks, which are interpreted to be part of the Bacan Formation of this study, contain pyrite, malachite, azurite filled veins.

**Volcaniclastic Rocks**

Volcaniclastic sedimentary rocks within the Bacan Formation are mostly greyish green to greyish blue. There are graded sandstone-siltstone with irregular stratification varying from 3 mm parallel laminations to 70 mm thick beds in S. Amasing (Plate 4.1b). Load and scour structures are present at the base of coarser beds (facies C2.3). Above these beds, there are fine grained sandstones and siltstones with cross lamination, with set heights up to 20 mm, and convolute lamination (facies B2.2). The upper part of the outcrop consists of siltstone-mudstone (facies D2). The section includes Bouma Tcd sequences with possible amalgamation of Tb units. Small scale normal faults forming horst and graben structures were observed within this sequence.
Plate 4.1a Typical exposure of Bacan Formation volcanic rocks, upper parts of S. Kusu-bibi.

Plate 4.1b Typical exposure of Bacan Formation volcanlastic rocks, lower parts of S. Amasing.

Plate 4.1c Typical exposure of Bacan Formation breccio-conglomerate, upper parts of S. Goro-goro.
The Bacan Formation

In many of the localities there are monomict, clast supported breccio-conglomerates (e.g. S. Amasing, S. Goro-goro, S. Aru and S. Kusu-Bibi). The poorly sorted conglomerates form boulders up to 6 m across with angular feldspar-phyric basaltic-andesite clasts, up to 50 mm across. The matrix consists of either metamorphosed fine grained silt-mudstone or volcanic rocks (in which case no clear compositional difference between the matrix and clasts can be detected; Plate 4.1c). Occasionally there are boulders of polymict, clast supported breccio-conglomerate (e.g. S. Nyonyifi), with up to 0.2 m across, angular-subangular sedimentary, andesitic and basaltic clasts. These breccio-conglomerate may represent disorganised gravel (facies A1.1) or lahar deposits.

Chert and Mudstone

The most extensive outcrop of these rocks is seen along the S. Amasing, where there is a 2.5 m thick bedded red chert sequence (bedding thickness ~0.05-0.4 m) interbedded with red mudstone. No internal structures were observed. These rocks are interpreted to be equivalent of facies E2.2 and Bouma Te.

4.4.2 Lower and Upper Boundaries

The lower contact of the Bacan Formation has not been observed. It has an unconformable upper contact with the Ruta Formation in S. Amasing. A fault contact with the Kaputusan Formation is inferred from aerial photographic studies of the Kusu-Bibi and Goro-goro areas.

4.4.3 Thickness

Due to incomplete exposure and faulting, the thickness of this formation is exceedingly difficult to ascertain. Based on the widespread occurrence of the formation and the height of hills composed entirely of this formation (up to 885 m), the thickness of the formation is estimated to be more than 1000 m.

4.4.4 Lateral Extent and Variations

The Bacan Formation is exposed at the northern part of Bacan (Fig.4.1). Lithological similarities to the South Bacan Formation make it possible that the Bacan Formation is exposed in the southern block of Bacan as well. Where the Bacan Formation is exposed, both the volcanic and volcanioclastic rocks are present (e.g. S. Nyonyifi, S. Goro-goro, north cost of Bacan, S. Kusu-Bibi and S. Kaputusan).

At S. Kusu-Bibi, volcanioclastic sandstones are interbedded with siltstones having beds of 20-70 mm thick (thickness increases up section). There are no internal sedimentary structures. Along the
The Bacan Formation

S. Kaputusan, there is a red mudstone outcrop above a pillow basalt. Overlying this is a graded volcaniclastic sandstone and siltstone showing thin-thick parallel lamination, possibly representing Tb. Float of similar volcanic and volcaniclastic rocks were collected from S. Nyonyifi.

Pudjowalujo & Bering (1982) described the presence of recrystallized limestone interbedded with breccio-conglomerate in the upper parts of S. Kaputusan. Although this was not observed, the reported bedding direction of 200/75 is similar to that of volcaniclastic rocks observed in this study in the same river, and therefore the limestone may be part of this formation.

4.4.5 Depositional Environment and Mechanism
The breccio-conglomerates may represent lahars, deposits of high concentration turbidity currents or debris flows. The other volcaniclastic rocks in the Bacan Formation are interpreted as the product of turbidity currents. The presence of the Bouma Tbcde sequence suggests a turbidite origin. Facies B2.2 indicates bed load transport beneath a dilute turbidity current or strong bottom current, most likely by grain flow. Facies C2 includes the classical Bouma sequence and is indicative of a dilute turbidity current deposit by bed load tractional transport. Facies D2 results from a low concentration turbidity current deposit with a grain by grain deposition from suspension. Facies E2.2 reflects grain by grain deposition of organic matter from suspension in a low concentration turbidity current.

4.5 PETROGRAPHY AND MINERAL CHEMISTRY
Unless stated, all percentages are modal proportions.

4.5.1 Volcanic Rocks

Phenocrysts are mostly plagioclase (~20%), clinopyroxene (~10%), opaques and occasional orthopyroxene. Relict olivine phenocrysts are present in some samples. The matrix consists of cryptocrystalline material, showing quenching (BR71), trachytic (BM132), intersertal (BM311) or seriate (BR73) textures. Plagioclase shows polysynthetic twinning and complex zoning patterns. It is albites and sometimes replaced by sercite, calcite or epidote. Pyroxene is replaced by calcite, epidote, pumpellyite, prehnite and zeolite and locally by green hornblende. Titanomagnetite occurs as needles and micro-phenocrysts, rarely replaced by sphene. Pumpellyite, smectite and chlorite replace the matrix. Calcite forms veins. Amygdales are filled by plagioclase, chalcedony, prehnite, zeolites and epidote outlined by pumpellyite and/or chloride (Plate 4.2a-d).
Plate 4.2a,b. PPL and XPL of Bacan Formation metavolcanic rock (BR71, prehnite-pumpellyite facies).

Plate 4.2c,d. PPL and XPL of Bacan Formation volcanic rocks (BM533, pumpellyite-actinolite facies). Scale for all photomicrographs = 0.5 mm.
4.5.1.2 Mineral Chemistry

Nine samples were chosen for mineral chemistry study (BR62, BR68, BR71, BR73, BR238, BM132, BM311, BM433 and BM533). Fig. 4.2 shows the localities of samples. The metamorphic grade ranges from zeolite (BM433) through prehnite-pumpellyite to greenschist facies (BM533).

**Plagioclase**

Phenocrysts occur mostly as sub-euhedral laths, displaying complex, polysynthetic and albite twinning, which lack zoning. These contain many small dark inclusions. Microlites are small laths, often bent. Composition varies in both varieties from Ca-rich to Na-rich. This variation can be detected even within a single thin section (e.g. BR62 and BM132). BM533 contains a Ca-rich variety only. Phenocryst size is unrelated to composition.

Na-rich phenocrysts are $\text{An}_{2.18}\text{Ab}_{8.97}\text{Or}_{0.99}$, with an average of $\text{An}_4\text{Ab}_9\text{Or}_1$ (n=30). These phenocrysts fall in the field of oligoclase-albite with an albite average. Ca-rich phenocrysts range from $\text{An}_{87.100}\text{Ab}_{8.11}\text{Or}_0$, with an average of $\text{An}_{96}\text{Ab}_{10}\text{Or}_0$ (n=8). These are anorthite and bytownite, with an anorthite average.

The microlites also show a bimodal distribution. Na-rich varieties are $\text{An}_{1.32}\text{Ab}_{6.70}\text{Or}_{0.2}$ (mean $\text{An}_{4.9}\text{Ab}_{8}\text{Or}_0$; n=12). Ca-rich microlites are $\text{An}_{86.92}\text{Ab}_{8.19}\text{Or}_0$ (average $\text{An}_{89}\text{Ab}_{11}\text{Or}_0$; n=3). The compositions of the microlites are comparable with the phenocrysts (Fig. 4.3a).

The marked bimodal distribution of compositions suggests that the Na-rich plagioclases are products of albitisation of an original Ca-rich variety. A primary Ca-rich plagioclase with complex twinning and inclusions suggests an arc origin. The absence of zoning may reflect alteration.

**Pyroxene**

Clinopyroxenes occur as phenocrysts in all samples except BR238. They are sub-euhedral, with simple twinning. No zoning was observed. There are no difference in the composition between the large and small phenocrysts. Clinopyroxene compositions are $\text{En}_{38.5}\text{Fs}_{8.36}\text{Wo}_{13.46}$, with an average of $\text{En}_{4}\text{Fs}_{14}\text{Wo}_{42}$. These clinopyroxenes plot in the field of diopside and augite, with one endiopside analysis (Fig. 4.3b). Following Nisbet & Pearce (1977) the pyroxenes plot in the Volcanic Arc Basalt field (Fig. 4.3c). One orthopyroxene ($\text{En}_{37}\text{Fs}_{38}\text{Wo}_{14}$) was found in BM433.
Figure 4.2 Location of Bacan Formation samples used in analytical work.
The Bacan Formation

**Fe-Ti Oxides**

Oxides have square outlines and occur as phenocrysts. There are three types recognised: Fe-rich, Ti-rich and Al-rich. Fe-rich varieties are present in all samples. All analyses plot on the ulvöspinel-magnetite tie-line, except for one ilmenite (Fig.4.3d).

**Apatite**

BM132 contains small, euhedral apatite phenocrysts.

**Glass**

Some relict interstitial glass has been analysed. It is distinguishable from plagioclase by its enrichment in SiO₂, Na₂O and K₂O.

**Phyllosilicates**

Chlorite occurs in BR71, BR73, BR238, BM132, BM433 and BM533 as yellow-green, radial crystals replacing pyroxene, and green poorly crystalline material replacing the groundmass. No obvious chemical differences were detected. Chlorite is distinguished from pumpellyite by its low CaO content (<1 wt%), whereas distinction from smectite is based on SiO₂ content (<40 wt%). They are similar to chlorites from low-grade rocks reported in Deer *et al.* (1992), Offler & Aguirre (1984) and Hakim (1989). The presence of K₂O, Na₂O and CaO in these chlorites indicates minor quantities of illite and smectite within the chlorite layers, typical of sub-greenschist metamorphism (Offler & Aguirre, 1984).

Anomalous yellow-orange, poorly crystalline smectite occurs in BM132, replacing pyroxene. Following Andrew (1980) this smectite is a high Fe, low Mg saponite which reflects non-oxidative diagenetic environment.

BR238 contains flakes of high-Fe and -Mg muscovite. These are similar to phengitic micas reported by Coombs *et al.* (1976) and those of low grade psammitic schists in Deer *et al.* (1992).

**Pumpellyite and Prehnite**

Yellow-orange pumpellyite occurs in vesicles or as a replacement of pyroxene, often forming radial fibres. They are similar to pumpellyites from other low grade meta-volcanic rocks reported by Coombs *et al.* (1976), Nakajima *et al.* (1977), Arvin (1982) and Hakim (1989). The distinction from epidote is based on >0.25 wt% MgO content and 90-96 wt% total. Sample BM132 differs from the rest in having a high Mg content. These pumpellyites plot in the fields of Upper Wakatipu I & II zones of Kawachi (1975), which is of prehnite-pumpellyite facies.
The Bacan Formation

Figure 4.3a. Ternary diagram showing all plagioclase compositions from the Bacan Formation.
Figure 4.3b. Quadrilateral diagram of pyroxenes in the Bacan Formation.
Figure 4.3c. Clinopyroxene tectonic discrimination diagram (after Nisbet & Pearce, 1977) applied to the Bacan Formation.
Figure 4.3d. Ternary diagram showing compositions of Fe-Ti oxides in the Bacan Formation.
Prehnite coexists with pumpellyite except in BM533. It is platy with high birefringence and occurs in vesicles and veins, and as a replacement of pyroxene (BM533). The pyroxene replacement has lower CaO and higher Na₂O contents than the others. Prehnite is distinguished from epidote in having higher SiO₂ (>42 wt%), lower FeO* (<7 wt%), slightly higher CaO (>16 wt%) and lower total (92-96 wt%).

**Epidote**

BR68, BR73, BR238 and BM311 contain epidotes replacing pyroxene, some of which grow in a radial habit. All epidote analyses (n=15) are similar to analyses from low-grade metavolcanic rocks reported by Coombs *et al.* (1976), Nakajima *et al.* (1977) and Deer *et al.* (1992) having Fe⁹⁺/(Fe⁹⁺+Al) ratios of 0.2-0.4. These are also within range with analyses of similar rocks in Offler & Aguirre (1984).

**Zeolite Group**

There are three types of zeolite in these rocks: stilbite, thomsonite and ferrierite. Stilbite occurs as a stubby, almost isotropic mineral in BR73 and BM132. Thomsonite is present as a fibrous, radial mineral in BR68, filling vesicles. Ferrierite in BM433 is a low birefringence mineral replacing plagioclase. The chemical differences between the three are tabulated in Table 4.1.

<table>
<thead>
<tr>
<th></th>
<th>Stilbite</th>
<th>Thomsonite</th>
<th>Ferrierite</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂ (wt%)</td>
<td>50-55</td>
<td>40-43</td>
<td>65-73</td>
</tr>
<tr>
<td>Al₂O₃ (wt%)</td>
<td>15-21</td>
<td>18-20</td>
<td>1.8-17.8</td>
</tr>
<tr>
<td>Ca:Na ratio</td>
<td>4.8-10.2</td>
<td>no Na</td>
<td>0.04-0.44</td>
</tr>
</tbody>
</table>

**Amphiboles**

Actinolite replaces pyroxene in BM533. It occurs as pleochroic green, anhedral crystals of various sizes. Fig.4.4a utilises the covariation of the Na content in amphiboles with the Al⁹⁺ content to estimate the pressure at the time of formation (Brown, 1977) indicates that the Bacan Formation amphiboles formed at ~2 kb. Fig.4.4b,c (from Raase, 1974) shows that the Bacan Formation amphiboles formed at >5 kb. Blundy & Holland’s (1990) geothermometry yielded a temperature of 560°C at 2 kb, and 510°C at 5 kb for the Bacan Formation amphiboles.
The Bacan Formation

Figure 4.4a. Geobarometry using covariation of NaM4 content in calcic amphibole with Al\textsuperscript{vi}. After Brown (1977).

Figures 4.4b,c. Covariations of Si with Al\textsuperscript{vi} and Ti in amphibole. After Raase (1974).
Sphene

BR238 contains dusty sphene, replacing Fe-Ti oxides. The presence of sphene indicates very low XcO$_3$ in the metamorphic fluid (Coombs et al., 1976). The high Al$_2$O$_3$ content may be due to a high confining pressure (Offler & Aguirre, 1984).

Quartz and Carbonate

Quartz occurs as a replacement of plagioclase, as a product of glass devitrification and filling vesicles. Carbonate veins were observed in some samples but were not analysed. In BM433 vesicles filled with smectite are lined by ankerite (FeMnCO$_3$).

4.5.1.3 Implications

A primary mineral assemblage of plagioclase + clinopyroxene + oxides + apatite and the textural features suggest that the Bacan Formation volcanic rocks were erupted in an arc-related setting. Three secondary mineral assemblages are recognised, these are: [1] albite + quartz + carbonate + prehnite + pumpellyte ± chlorite ± zeolite ± smectite; [2] albite + quartz + carbonate + zeolite + prehnite + pumpellyte + epidote ± chlorite ± sphene ± muscovite and [3] quartz + chlorite + pumpellytite + actinolite. Assemblages [1] and [2] are typical of prehnite-pumpellyite facies (e.g. Bishop, 1972; Offler & Aguirre, 1984; Liou et al., 1987). Assemblage [3] is typical of pumpellyite-actinolite facies, marked by the disappearance of prehnite (Bishop, 1972) and is similar to the pumpellyite-actinolite assemblage from the Taveyanne Formation (Coombs et al., 1976) estimated as indicating temperatures of ~250-350°C. Nakajima et al. (1977) divided the pumpellyite-actinolite facies into higher temperature part, which passes into the greenschist and low pressure glaucochane schists facies, having the assemblage pumpellyite + epidote + chlorite + actinolite and low temperature part characterised by the assemblage of pumpellyite + haematite + chlorite + actinolite. The boundary is marked by the coexistence of epidote, having Fe$^{3+}$/(Fe$^{3+}$+Al) = 0.10-0.15, with pumpellyite, actinolite and chlorite. Although assemblage [2] is of prehnite-pumpellyite facies, its higher Fe$^{3+}$/(Fe$^{3+}$+Al) ratios indicates lower temperatures (Nakajima et al., 1977; Liou et al., 1983), possibly indicating a transitional facies from prehnite-pumpellyite to pumpellyite-actinolite.

Prehnite-pumpellyite facies metamorphism requires metamorphic conditions of ~240-330°C and ~2 kb, whereas the conditions for pumpellyite-actinolite facies are ~280-380°C and 2-8 kb (Liou et al., 1987). The lack of fabric development suggests a regional, static metamorphism, due to burial, with metamorphic facies variations attributed to pressure variations, interpreted as reflecting different burial levels. The metamorphic conditions for the Bacan Formation is therefore
interpreted as ~250-330°C, with varying pressures from 2 kb upwards. The higher temperature estimate derived from the amphibole-plagioclase geothermometry may be due to lack of equilibrium between these phases in BM533. The presence of hydrothermal alteration is attributed to local thermal activity possibly due to later volcanism.

4.5.2 Volcaniclastic Rocks

*Conglomerate*

Clasts in the conglomerates are mostly feldspar-phyric andesite and minor quartz diorite. They are petrographically similar to the volcanic rocks described above and the Nusa Babi Monzodiorite *(Chapter Nine)*.

*Sandstone*

The poorly-moderately sorted sandstone is dominated by subrounded-subangular lithic fragments of probable basic composition, with minor plagioclase, pyroxene, opaque, epidote, sphene, chlorite and pumpellyite. Mineral fragments tend to be more angular. Grains are in long contact with each other (Plate 4.3a,b). Plagioclase occurs as laths showing complex twinning and zoning, indicative of an arc volcanic origin. Laths are calcitised and albitised. Pyroxene and opaques occur as broken euhedral crystals. Epidote has a drusy texture and appears to replace pyroxene and plagioclase. Sphene is present in some samples (e.g. BR239-BR241). Chlorite replaces the matrix and the glass in the volcanic clasts. Pumpellyite replaces pyroxene and matrix. The metamorphic mineralogy is identical to the volcanic rocks of the formation and suggest metamorphism of prehnite-pumpellyite to pumpellyite-actinolite facies.

*Mudstone and Chert*

The micritic mudstones include rare planktonic foraminifera and contain layers of fine grains of plagioclase, pyroxene, opaques and minor volcanic lithic fragments. The cherts have a brecciated appearance with interstitial quartz and minor zeolite. There are many euhedral pyrite cubes within the chert. Occasional crustacean excrements (coprolites) were observed (Plate 4.3c,d).

4.6 WHOLE ROCK CHEMISTRY

The purpose of the chemical work was to characterise the formation, decipher its tectonic setting and compare the volcanic rocks to the Saleh Complex and the Oha Volcanics of southwest Halmahera. Two samples were analysed for whole rock chemistry analyses using XRF technique (BR73 and BM433). Four additional samples were analysed using the ICP method.
Plate 4.3a,b. PPL and XPL of Bacan Formation sandstone (BM422).

Plate 4.3c. XPL of Bacan Formation mudstone (BM416). Note the presence of planktonic foraminifera.

Plate 4.3d. XPL of Bacan Formation chert (BM70). Note the presence of coprolites. Scale for all photomicrographs = 0.5 mm.
Except BM132 and BM533, all samples are in situ. Because of differences in analytical precision of the two methods, particularly affecting the trace element determinations, the samples will be treated in groups according to the analytical method.

### 4.6.1 Major Elements

All samples analysed contain high LOI (1.55-2.73 wt%), with only samples BM433 and BM533 unaltered (LOI < 2.5 wt%, H$_2$O < 2 wt%; Cann, 1971; Le Bas et al., 1986). The alkalis K$_2$O, CaO and Na$_2$O; and LIL (Sr and Ba) are susceptible to alteration (e.g. Menzies & Seyfried, 1979; Moody, 1979; Alt et al., 1986). A plot of these elements against Zr (immobile, HFS element), will reflect their mobility relative to the HFS, which is indicative of the degree of alteration. A plot of HFS against another HFS element (Y) will serves as a control to determine whether the rocks are cogenetic. Fig. 4.5a shows that all samples are cogenetic. Most of the alkalis and the LILE correlate well with Zr, except BM132 and BM533 (highest and lowest Zr content), which may reflect alteration and will, therefore, be treated with caution.

Fig. 4.5b shows that BR73, BM311 and BM533 are basalt; BR238 is a basaltic andesite and BM132 and BM433 are basaltic trachy-andesites (Le Bas & Streckeisen, 1991). Using the Harker-type diagrams (Figs. 4.6a-g) it is apparent that SiO$_2$ correlates well with MgO, Na$_2$O, CaO and Fe$_2$O$_3$, suggesting fractional crystallisation as the main mode of magma differentiation. Covariations of Al$_2$O$_3$, TiO$_2$ and FeO*/MgO with SiO$_2$, however show two trends, with the basaltic samples demonstrating decrease of Al, Ti and FeO*/MgO with increasing SiO$_2$, while the basaltic andesites showing the opposite. These trends may be a reflection of source heterogeneity or crustal contamination.

Table 4.2 compares the average analyses by ICP and XRF with MORB, High-K Calc Alkaline (HKCA) and IACA. Due to their close proximity and probable similar ages, the rocks are also compared to the Oha Formation of South Halmahera.

The ICP and XRF methods yield very similar major element results. The alkali contents of the Bacan Formation rocks vary considerably which may be attributable to the low temperature metamorphism the rocks have suffered (Menzies & Seyfried, 1979; Moody, 1979; Alt et al., 1986). In general they are comparable to the IACA average. The Bacan Formation rocks have very low MgO and high FeO* contents, and this may be due to a high degree of olivine fractionation (Roeder & Emslie, 1970). In general, the major elements are comparable to those of the Oha Formation.
The Bacan Formation

Figure 4.5a. Covariations of bulk alkali and LIL contents with HFS (Zr) to determine degree of alteration. Y serves to establish whether the samples are cogenetic.

Figure 4.5b. Classification of Bacan Formation volcanic rocks based on total alkali against SiO$_2$ contents (after Le Bas & Streckeisen, 1991).
The Bacan Formation

Figures 4.6a-g. Harker-type diagrams; covariations of SiO₂ with MgO, Na₂O, CaO, Fe₂O₃, Al₂O₃, TiO₂ and FeO*/MgO.
The Bacan Formation

Table 4.2. Comparison of the average ICP and XRF results from the Bacan Formation with rocks from known tectonic settings and the Oha Formation. (parentheses) = range. Data for MORB, HKCA and IACA are from Wilson (1989), data for OHA are from (Hakim, 1989; Hakim & Hall, 1991).

<table>
<thead>
<tr>
<th>Wt%</th>
<th>ICP</th>
<th>XRF</th>
<th>MORB</th>
<th>HKCA</th>
<th>IACA</th>
<th>OHA</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>51.92 (49.53-54.32)</td>
<td>52.78 (50.81-54.75)</td>
<td>50.67</td>
<td>51.00</td>
<td>49.40</td>
<td>51</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.70 (0.59-0.83)</td>
<td>0.72 (0.69-0.76)</td>
<td>1.28</td>
<td>0.93</td>
<td>0.70</td>
<td>0.8</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>16.07 (49.53-54.32)</td>
<td>16.66 (50.81-54.75)</td>
<td>15.45</td>
<td>13.60</td>
<td>13.29</td>
<td>16.8</td>
</tr>
<tr>
<td>FeO⁺</td>
<td>9.27 (6.75-11.35)</td>
<td>8.65 (8.01-9.18)</td>
<td>9.67</td>
<td>8.11</td>
<td>10.15</td>
<td>7.6</td>
</tr>
<tr>
<td>MnO</td>
<td>0.21 (0.13-0.30)</td>
<td>0.19 (0.18-0.19)</td>
<td>[0.15]</td>
<td>0.14</td>
<td>0.20</td>
<td>-</td>
</tr>
<tr>
<td>MgO</td>
<td>5.42 (3.91-6.52)</td>
<td>5.57 (4.26-6.87)</td>
<td>9.05</td>
<td>12.50</td>
<td>10.44</td>
<td>5.5</td>
</tr>
<tr>
<td>CaO</td>
<td>7.02 (4.37-10.62)</td>
<td>8.17 (5.31-11.03)</td>
<td>11.72</td>
<td>7.92</td>
<td>12.22</td>
<td>4.12</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.64 (0.54-3.63)</td>
<td>4.44 (4.01-4.86)</td>
<td>2.51</td>
<td>2.67</td>
<td>2.16</td>
<td>-</td>
</tr>
<tr>
<td>K₂O</td>
<td>1.43 (0.14-2.93)</td>
<td>1.37 (0.28-2.46)</td>
<td>0.15</td>
<td>2.37</td>
<td>1.06</td>
<td>&gt; 0.9</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.21 (0.19-0.25)</td>
<td>0.16 (0.18-0.20)</td>
<td>0.20</td>
<td>0.59</td>
<td>0.20</td>
<td>-</td>
</tr>
</tbody>
</table>

4.6.2 Trace Elements

To further investigate the fractional crystallisation process, compatible elements (Ni, Cr and Sc) and the ratios of compatible over incompatible elements (Sc/Ba and Ti/Zr) are plotted against SiO₂. These diagrams (Fig.4.7a-e) show a trend which may be an indication of magma replenishment: there is a decrease of Ni, Cr and Sc with SiO₂ up to ~51 wt% SiO₂ and an increase above this value. Decreasing compatible elements with increasing SiO₂ indicates fractionation of olivine, clinopyroxene and plagioclase. BM533 (SiO₂ = 40.65 wt%) is always an outlier in these diagrams, reflecting its altered state.

To clarify the tectonic setting, rocks from this formation are compared to average N-MORB, IACA, BABB and the Oha Formation. Additionally, because of the geographical distribution of the Saleh Metamorphic Complex and its uncertain age (see Section 3.5), the Bacan Formation is also compared with the Saleh Complex.

There are differences in some trace element concentrations measured by ICP and XRF methods, principally Zr, Nb and La. For these elements the XRF results are preferred, although La by XRF may be slightly low (M.F.Thirlwall, pers.comm. 1992).
Figure 4.7a-e Covariations of SiO₂ with Ni, Cr, Sc, Sc/Ba and Ti/Zr.
Table 4.3. Comparison of selected trace elements and their ratios from the Bacan Formation with rocks from known tectonic settings, the Saleh Metamorphic Complex and the Oha Formation. Values in [squared parentheses] are from BAT values of Wilson (1989). Data are from Wilson (1989); Wood et al. (1981) for BABB; and Hakim & Hall (1991) for Oha Formation.

<table>
<thead>
<tr>
<th></th>
<th>ICP</th>
<th>XRF</th>
<th>NMORB</th>
<th>IACA</th>
<th>BABB</th>
<th>Saleh</th>
<th>Oha</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sr</td>
<td>360 (235-436)</td>
<td>439 (308-570)</td>
<td>124</td>
<td>550</td>
<td>511</td>
<td>240</td>
<td>438</td>
</tr>
<tr>
<td>Zr</td>
<td>71.3 (8-127)</td>
<td>46 (21.6-70.3)</td>
<td>85</td>
<td>40</td>
<td>82</td>
<td>90.7</td>
<td>81</td>
</tr>
<tr>
<td>Nb</td>
<td>5.8 (5-6)</td>
<td>1.4 (0.8-1.9)</td>
<td>3.1</td>
<td>1.4</td>
<td>[8.0]</td>
<td>2.3</td>
<td>2</td>
</tr>
<tr>
<td>Y</td>
<td>23.3 (19-26)</td>
<td>19.6 (17.2-21.4)</td>
<td>29</td>
<td>15</td>
<td>25</td>
<td>23.4</td>
<td>23</td>
</tr>
<tr>
<td>La</td>
<td>13.4 (1.9-21.7)</td>
<td>5.1 (1.7-8.5)</td>
<td>3.0</td>
<td>10</td>
<td>12.5</td>
<td>9.5</td>
<td>10</td>
</tr>
<tr>
<td>Nd</td>
<td>12.3 (9.8-16.6)</td>
<td>10.4 (8.4-12.4)</td>
<td>7.7</td>
<td>13</td>
<td>[13.1]</td>
<td>16.4</td>
<td>15</td>
</tr>
<tr>
<td>Zr/Nb</td>
<td>12.3 (1.3-21.2)</td>
<td>32 (27-37)</td>
<td>24.3</td>
<td>28.6</td>
<td>[16.3]</td>
<td>39.4</td>
<td>41</td>
</tr>
<tr>
<td>Zr/Y</td>
<td>2.9 (0.4-5.1)</td>
<td>2.2 (1.3-3.2)</td>
<td>2.9</td>
<td>2.7</td>
<td>3.3</td>
<td>3.9</td>
<td>3.5</td>
</tr>
<tr>
<td>Y/Nb</td>
<td>4.1 (3.2-4.6)</td>
<td>16.5 (11.6-21.5)</td>
<td>9.4</td>
<td>10.7</td>
<td>[3.8]</td>
<td>10.2</td>
<td>11.5</td>
</tr>
<tr>
<td>La/Nb</td>
<td>2.3 (1.8-3.6)</td>
<td>3.3 (2.1-4.5)</td>
<td>1.0</td>
<td>7.1</td>
<td>[1.0]</td>
<td>4.1</td>
<td>5</td>
</tr>
<tr>
<td>Ba/La</td>
<td>40.7 (2.6-86.6)</td>
<td>39.5 (36.4-42.6)</td>
<td>4.0</td>
<td>30</td>
<td>11.8</td>
<td>33.2</td>
<td>17.9</td>
</tr>
</tbody>
</table>

The LIL elements (Sr, Rb and La) vary widely, and cause variations in the La/Nb and Ba/La ratios. These are attributed to mobility of LIL during alteration. The Y, Nd and Nb contents and the Y/Nb ratio of XRF are similar those for the Saleh Complex and the Oha Formation. The Nb and Y contents are comparable to IACA, whereas Nd contents are similar to both IACA or BABB. Zr contents, and ratios Zr/Nb and Zr/Y are similar to the Saleh Metamorphic rocks, the Oha Formation and BABB.

4.6.3 Tectonic Discrimination Diagrams

Figs. 4.8a-c are ternary tectonic discrimination diagrams utilising immobile trace elements and major elements. On the Ti-Zr-Y diagram (Pearce & Cann, 1973), the Bacan Formation rocks (except BM533 which has a low Zr content) plot in the field which includes OFB and CAB; whereas on the Nb-Zr-Y diagram (Meschede, 1986) they are distributed in the fields of N-MORB & VAB and VAB & WPT. On the TiO₂-MnO-P₂O₅ diagram of Mullen (1983) they plot in the CAB field.
Figures 4.8a-c. Ternary tectonic discrimination diagrams applied to the Bacan Formation. See text for details.
On the MORB normalised diagram of Pearce (1982) the Bacan Formation rocks show characteristics of arc rocks: humped LIL; Nb depletion; flat Ce-P; Zr depletion, particularly BR73 and BM533; Ti-Sc depletions, all below MORB values; and strong Cr depletion (Fig.4.9a).

Fig.4.9b is a chondrite normalised diagram (Sun, 1980) showing a spiky distribution of the trace elements with a general decrease of abundance with increasing incompatibility. The Nb, Zr and Ti depletions, and the Ba, K and Sr enrichments are typical of island arc volcanic rocks. Fig.4.9c is a chondrite normalised (Wakita et al., 1971) diagram for the REE elements. On this diagram there is a clear enrichment of Heavy REE (HREE) relative to Light REE (LREE), with a slight Eu, Ho and Yb depletions. This pattern is typical of orogenic andesites showing plagioclase fractionation (Gill, 1981; Cullers & Graf, 1984). The REE concentrations of the Bacan Formation rocks are similar to the lower limit of the more REE-enriched continental andesites and the upper limit of the less REE-enriched island arc andesites. An important conclusion indicated by this diagram is that despite the wide variations of some of the trace elements, the REE are grouped together indicating that these samples are cogenetic and were derived from the same source.

Figs.4.10a-e are covariation diagrams of immobile trace elements (Cr against Y and Ce/Sr; Ti/Y against Nb/Y; and Zr against Ti and Zr/Y) of Pearce (1982) and Pearce & Norry (1979). These diagrams show the Bacan Formation rocks plot in the VAB and MORB fields. On the FeO*/MgO versus TiO₂ diagram of Hawkins & Evans (1983) they plot along the IAT trend (Fig.4.10f).

**4.6.4 Implications**

The major, trace elements, their ratios and tectonic discrimination diagrams clearly indicate that the Bacan Formation rocks are arc-related volcanics. Some trace element concentrations and ratios suggest formation in a back-arc basin. This is consistent with the presence of interbedded turbiditic deposits with the volcanic rocks.

There are many similarities between the Bacan Formation, the Saleh Metamorphic and the Oha Formation which could indicate they are the products of the same arc.

**4.7. AGE DETERMINATION**

Planktonic foraminifera were found in a small number of samples but only a single sample (BM419) proved dateable. This sample indicated that the Bacan Formation includes rocks of early Late Eocene age (F.T.Banner, pers.comm. 1992). Foraminifera present include: *Globigerinatethka tropicalis, Turborotalia centralis* and *globigerinids*. 

Figure 4.10f. Covaritions of TiO₂ and FeO*/MgO to distinguish back-arc from arc setting for the Bacan Formation (after Hawkins & Evans, 1983).
K-Ar analyses were performed on two samples. A sample from the north coast of Bacan (BR73) yielded a 14.9 ± 2.3 Ma age. A second sample from S. Kusu-Bibi (BM433) gave an age of 14.8 ± 1.2 Ma. Although consistent with each other, these ages are much younger than the biostratigraphic age. Fig.4.11 is an isochron diagram plotted for the Bacan Formation, showing a thermal event at ~14 Ma probably related to the intrusion of Saleh Diorite (see Chapter Nine), resetting the Bacan Formation. The thermal activity is interpreted to be associated with initiation of the Neogene arc.

Table 4.4 Summary of the K-Ar results from the Bacan Formation.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Material</th>
<th>Grain size (μ)</th>
<th>%K (1σ error)</th>
<th>Wt for Ar (g)</th>
<th>#Ar* (nl/g, 1σ error)</th>
<th>#Ar* (σ)</th>
<th>Age (Ma, 2σ error)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BR73</td>
<td>wr</td>
<td>125-250</td>
<td>0.213 ± 1.00%</td>
<td>1.0016</td>
<td>0.0966 ± 35.68%</td>
<td>97.01</td>
<td>11.6 ± 8.3</td>
</tr>
<tr>
<td>BR73dp</td>
<td>wr</td>
<td>250-425</td>
<td>0.212 ± 4.30%</td>
<td>1.0198</td>
<td>0.0409 ± 25.66% (1)</td>
<td>96.13</td>
<td>5.0 ± 2.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.0832 ± 8.35% (2)</td>
<td>88.99</td>
<td>10.0 ± 1.9</td>
</tr>
<tr>
<td>BM433</td>
<td>wr</td>
<td>250-500</td>
<td>2.100 ± 2.68%</td>
<td>1.0262</td>
<td>1.2124 ± 2.93% (1)</td>
<td>73.13</td>
<td>14.9 ± 2.3</td>
</tr>
</tbody>
</table>

(1) denotes lower temperature heating, (2) higher temperature heating, (∑) sum of both step heatings; see Appendix D for error calculation on the sum of step heatings.

4.8 STRUCTURE

Exposures along the S. Amasing have two principal bedding attitudes, with the volcanic rocks dipping ~120/45 and the sedimentary rocks ~180/40. Either there is an unconformity between the sedimentary and volcanic rocks or folding/faulting affects the formation. The inconsistency in bedding attitudes throughout the region, is supportive of the latter explanation.

Pervasive jointing occurs in most exposures. A plot of all the joint directions cutting through this formation is provided in Fig.4.12. There is a bidirectional distribution (N-S and ENE-WSW). These directions are similar to the regional structure in Bacan, and are also observed in the younger formations with different bedding attitudes, indicating post-Neogene joint development.

Pudjowalujo & Surjono (1982) measured the density of joints/metre² in S. Kaputusan (1-50, mostly 5-10) and observed that it is related to the % volume of sulphide. They concluded that sulphide mineralisation is connected to the joint development. In contrast, Cu-mineralisation is attributed by them to stock intrusion, supported by the presence of a sulphide and quartz-sericite halo.
Figure 4.11. Isochron diagram showing that the Bacan Formation rocks have been affected by a thermal event at ~ 15 Ma.

Figure 4.12. Summary diagram of joint directions affecting the Bacan Formation.
surrounding the Cu-rich areas. Joint directions (330-350 and 280-310), measured mostly in the feldspar-phyric andesite (n=194) with additional measurements of the direction of quartz veins (n=25), differ from those in this study.

4.9 SYNTHESIS AND TECTONIC SIGNIFICANCE
The lower Upper Eocene Bacan Formation is interpreted to be younger than the Sibela Metamorphic Complex, and older than all other formations of Bacan. This interpretation is based on its micropalaeontological age, stratigraphic position and the grade of metamorphism. The formation consists of volcanic and volcaniclastic rocks. The petrography, mineral and whole rock chemistry suggest that the basic-intermediate volcanic rocks are arc-related, and possibly formed in a back-arc basin. The volcaniclastic rocks are interpreted to be turbiditic, deposited in an arc-related basin.

The formation has been affected by prehnite-pumpellyite to pumpellyite-actinolite facies metamorphism. The lack of fabric development suggests a regional, static metamorphism, due to burial, with facies variations due to different levels of burial (~2 kb upwards, ~250-330°C). The "microconglomerate" weathering pattern is thought to be a result of alteration of lahars. The presence of quartz mineralisation is interpreted to be related to Neogene plutonic and volcanic activity. A thermal event at ~14-15 Ma affected this formation. The relationship of the Bacan Formation with the younger Tawali and South Bacan Formations and their tectonic significance will be discussed in more detail in Chapter Eleven.
CHAPTER FIVE
THE SOUTH BACAN FORMATION

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CHAPTER FIVE
THE SOUTH BACAN FORMATION

5.1 INTRODUCTION
Although the South Bacan Formation is apparently younger than the Tawali Formation (Chapter Six), it is discussed here because of its similarities with the Bacan Formation (Chapter Four). The distinction between the South Bacan and Bacan Formations is based entirely on age difference. The South Bacan Formation consists of basaltic-andesites and hornblende andesites interbedded with volcaniclastic conglomerates, sandstones, siltstones and mudstones, most of which are metamorphosed to a prehnite-pumpellyite facies. In this chapter, the lithology and petrography of the sedimentary rocks, the petrology and mineral chemistry of the volcanic rocks, and age determination and structural evidence from the formation are discussed and their implications considered.

5.2 SYNONYMY
The name of this formation is deliberately similar to the Bacan Formation to emphasise their similarities. Previous to this study, this formation has been called the Bacan Formation (Yasin, 1980) and the Old Volcanic Unit (Silitonga et al., 1981).

5.3 AERIAL PHOTOGRAPHY
As with the Bacan Formation, this formation is characterised by rugged, pointed, high hills forming continuous ridges with multiradial-rectangular drainage patterns.

5.4 LITHOFACIES
5.4.1 Type Area
The type area for this formation is S. Wayatim, south Bacan, where there is a sequence of bedded basaltic-andesites with indurated volcaniclastic conglomerates, sandstones, siltstones and mudstones. The basaltic andesites contain plagioclase laths and pyroxene phenocrysts up to 10 mm across, with a dark glass matrix. In places the volcanic rocks contain zeolite-filled amygdales, typically about 5-10 mm in diameter but otherwise lack internal structures (Plate 5.1a). A peculiar feature of the basaltic-andesite unit is the microconglomeratic appearance, similar to that of the Bacan Formation (see Section 4.4.1; Plate 5.1b).

Volcaniclastic sandstones at S. Wayatim have a consistent bedding thickness of ~0.2 m. They are parallel laminated on a mm scale and in some beds normal grading and convolute lamination are seen, indicating Bouma Tbc sequences (facies C2.2; Plate 5.1c). Red and white mudstones are
The South Bacan Formation

interbedded, on a 0.05-0.2 m scale, with finer grained volcanioclastic rocks (Bouma Tde, facies D2.2; Plate 5.1d). At this locality, the volcanioclastic unit is bounded at the base by a volcanic breccia and at the top by massive volcanic rocks. Exposures are typically pervasively jointed with feldspars veins up to 10 mm thick.

Monzodiorites, similar to the Nusa Babi Monzodiorite, intrude this formation in several localities (e.g. S. Mau, S. Rain and S. Bibinoi). Disseminated limonite, haematite, pyrite, galena and sulphide mineralisation are present near or along the contact with the monzodiorites. There are also outcrops of melanocratic microdiorites, mineralogically similar to the Saleh Diorite, intruding the basaltic-andesite rocks (e.g. along S. Wayakuba), locally forming dykes. Details of both intrusive body are given in Chapter Nine.

5.4.2 Lower and Upper Boundaries
The lower contact of the South Bacan Formation is not observed, and this formation is the oldest unit in the South Bacan block. An unconformable upper contact with the Ruta Formation is indicated by the presence of subrounded South Bacan Formation clasts in the Ruta Formation.

5.4.3 Thickness
Based on the height of the hills around S. Wayatim (413 m), composed entirely of this formation with a 40-50° dip, the minimum thickness of this formation is estimated to be ~350 m.

5.4.4 Lateral Extent and Variations
This formation is found throughout the southern block of Bacan (Fig.5.1), although the similarity with the Bacan Formation does not preclude the presence of this formation elsewhere. There are very little variations throughout the studied area.

Subrounded float of amphibole-bearing andesite, ~0.2 m across, were collected at S. Bibinoi. These differ from the Neogene volcanic rocks (Kaputusan Formation) by the presence of zeolite and prehnite veins and amygdales.

North of Silang Village at Tg. Boso-boso there are boulders of monomict, clast supported breccio-conglomerates with a siltstone matrix, up to 2 m across. These are poorly sorted with angular clasts up to 0.12 m across (Plate 5.1e). The exposures are deeply weathered and are cut by abundant calcite veins. These breccio-conglomerates may represent lahars or debris flows (facies Fl.1).
Plate 5.1a. Typical volcanic rock outcrop of the South Bacan Formation, along S. Wayatim.

Plate 5.1b. 'Microconglomerate' alteration pattern along S. Bibinoi. See text for discussion.

Plate 5.1c. Typical volcaniclastic outcrop of the South Bacan Formation, along the upper parts of S. Wayatim. Note conjugate normal faulting forming horst and graben structures next to the hammer head.

Plate 5.1d. South Bacan Formation mudstone along S. Wayatim. Note convolute lamination and slight grading. Pen and hand are towards the top of the outcrop.

Plate 5.1e. Typical breccio-conglomerate rocks within the South Bacan Formation, along Tg. Boso-boso coast.
Figure 5.1 Distribution of the South Bacan Formation.
5.4.5 Depositional Environment and Mechanism
Interbedded with the volcanic rocks are volcanioclastic and hemipelagic rocks. The presence of Bouma Tbc and possibly Ta in the sedimentary rocks argues for a 'turbidity current' deposit. Pickering et al., 1989 proposed that facies C2.2 indicates deposition from turbidity current of intermediate character by bed load tractional transport. Facies D2.2 demonstrates deposition from low concentration turbidity current with grain-by-grain deposition from suspension, followed by traction transport of the silt load. Facies F1.1 signifies debris flow deposit with deposited clasts due to cessation of movement from basal friction. The facies analyses indicate deposition by intermediate turbidity currents which dissipated their energy as they travel down slope.

5.5 PETROGRAPHY AND MINERAL CHEMISTRY
5.5.1 Volcanic Rocks
5.5.1.1 Petrography
In the volcanic rocks plagioclase laths are calcitised and albitised and show polysynthetic twinning and zoning. Euhedral clinopyroxenes are mostly fresh but locally replaced by calcite, zeolites, prehnite, pumpellyite, epidote and amphiboles. Opaque minerals occur as needles and microphenocrysts. The matrix consists of glass with a quenched texture. Pumpellyite and chlorite replace the matrix. Epidote has a drusy texture. Sphene is locally present. Amygdales are filled by feldspars, silica, prehnite, zeolites and epidote, and outlined by pumpellyite and/or chlorite. The mineral assemblages are typical of basic-intermediate volcanic rocks that have been affected by prehnite-pumpellyite facies metamorphism (Plate 5.2a,b).

5.5.1.2 Mineral Chemistry
Three samples (BM188, BM250 and BM282) have been chosen for mineral chemistry study. The aim of this study were: [1] to determine the metamorphic conditions, [2] to gain an insight into the tectonic setting and [3] to compare these rocks to the Saleh Complex and the Bacan Formation. Fig.5.2 shows the locations of samples used for analytical work.

Feldspars
Plagioclase (labradorite-albite) occurring as phenocryst, inclusions in amphibole and groundmass microlites. Phenocrysts consist of euhedral-subhedral laths showing polysynthetic twinning and complex zoning. The plagioclase phenocryst compositions are An_{66-69}Ab_{31-34}Or_{10-13} (mean An_{64,5}Ab_{37,25}Or_{11,75}; n=7). There are up to 57 mol% difference in the composition of these phenocrysts within one thin section (e.g. BM188). Plagioclase inclusions are present in BM250 and their composition (An_{33, 31}, Ab_{67,77}, Or_{10,2}; n=2) are very different from the phenocryst phase. Orthoclase inclusions also in
The South Bacan Formation

BM250 is An$_{2}$Ab$_{1}$Or$_{65}$ ($n=1$). Plagioclase groundmass crystals are An$_{5,71}$Ab$_{29,95}$Or$_{9.3}$ (mean An$_{50}$Ab$_{60}$Or$_{1}$; $n=10$), whereas the alkali feldspar groundmass crystals are An$_{1.5}$Ab$_{15.99}$Or$_{55.83}$ ($n=2$). As with the phenocrysts, up to 46 mol% variation in the groundmass can be detected within a thin section (e.g. BM250). The wide compositional variations within individual sample (Fig.5.3a) may be a reflection of the varying degree of albitisation.

Amphiboles

Pale green, euhedral-subhedral magnesiohornblende occur as primary phenocryst in BM250 and as granular magmatic crystals in BM282. Geothermometry estimation using the methods of Blundy & Holland (1990), amphiboles crystallises in the range of 770-834°C at 1 kb. Gill (1981) suggested that orogenic amphiboles with Si contents >6.5 are typical of the ones erupted above a continental crust. All amphiboles in the South Bacan Formation have Si contents >6.5.

Clinopyroxene

Euhedral-subhedral clinopyroxene phenocrysts in BM188 and BM282 are simply twinned and unzoned. Locally these are replaced by amphibole. The compositions of these are En$_{36.44}$Fs$_{59.29}$Wo$_{12.49}$ (mean En$_{4}$,Fs$_{1.1}$Wo$_{4}$; $n=6$). They are diopside, augite and salite (Fig.5.3b) and when plotted on the Nisbet & Pearce (1977) tectonic discrimination diagram, they mostly fall in the VAB field, with three analyses in the OFB & WPB field (Fig.5.3c).

Fe-Ti Minerals

There are three types of Fe-Ti minerals recognised: [1] low-Ti titanomagnetite in BM188, BM250 and BM282; [2] high-Ti titanomagnetite (>27 wt% TiO$_{2}$) in BM282; and [3] pyrite in BM282. In all cases these minerals form euhedral-subhedral phenocrysts and in BM250 they also form inclusions in amphibole. Titanomagnetite in BM250 are similar to the Bacan Formation oxides, plotting on the magnetite-ulvöspinel tie-line. Fig.5.3d shows the composition of titanomagnetites, plotting mostly between magnetite and pseudobrookite, near the magnetite end.

Apatite

Apatite is present in BM250 as inclusions in amphibole and is the only P-bearing mineral.

Chlorite

Chlorite occurs in BM188 replacing amphibole or groundmass and occasionally forming veins. All analyses are comparable to those from low-grade metavolcanic rocks reported by Hakim (1989) and they contain Na, K and Ca, indicating subgreenschist metamorphism (Offler & Aguirre, 1984).
Figure 5.2 Location of South Bacan Formation samples used in analytical work.

Key:

- Volcaniclastic Rocks
- Volcanic Rocks

Italics = Not In Situ
Figure 5.3a. Ternary diagram showing all feldspar compositions from the South Bacan Formation.

Figure 5.3b. Pyroxene quadrilateral showing compositions of clinopyroxene in the South Bacan Formation.

Figure 5.3c. Clinopyroxene tectonic discrimination diagram applied to the South Bacan Formation.

Figure 5.3d. Ternary diagram showing the compositions of Fe-Ti oxides in the South Bacan Formation.
Prehnite and Pumpellyite

Radial prehnite occurs in BM250, growing in veins, closely associated with quartz. These have high FeO contents compared to those reported by Deer et al. (1992).

Pumpellyite is observed in BM188 as a fibrous mineral, replacing pyroxene. These are similar to those from low-grade metavolcanic rocks reported by Coombs et al. (1976), Nakajima et al. (1977), Arvin (1982) and Hakim (1989).

Quartz, Carbonate and Zeolite

Quartz is a common secondary mineral replacing glass. Carbonate is present in some samples, forming veins. Heulandite occurs in BM250, replacing plagioclase.

5.5.2 Volcaniclastic Rocks

Conglomerate

Clasts in the conglomerates are similar to the metavolcanic rocks described above.

Sandstone

The sandstone is poorly-moderately sorted with subrounded-subangular lithic fragments, of similar composition to the volcanic unit, and more angular mineral fragments (Plate 5.2c,d). Grains are in long contact with each other. Plagioclase laths show complex twinning and zoning and are calcitised or albitised. Pyroxene and opaques are broken euhedral crystals. Epidote has a drusy texture and appears to replace pyroxene and plagioclase. Sphene is locally present. Chlorite replaces the matrix and the glass in the lithic clasts. Pumpellyite replaces pyroxene and matrix.

Mudstone and Chert

The micritic mudstones contain rare rounded planktonic foraminifera and some layers of poorly sorted, fine angular grains of plagioclase, pyroxene, opaques and minor volcanic lithic fragments. The chert includes angular grains of plagioclase, pyroxene and volcanic lithic fragments. It has a brecciated appearance with quartz and zeolite filling the voids between grains (Plate 5.2e,f).

5.5.3 Implications

In the volcanic rocks, primary mineral assemblage of plagioclase + clinopyroxene + oxides + apatite and the textural character suggest formation in an arc-related setting. This is supported by plagioclase petrography and the pyroxene tectonic discrimination diagram. Secondary mineral assemblages of albite + calcite + quartz + zeolite + chlorite + prehnite + pumpellyite are similar.
Plate 5.2a-b. XPL and PPL of the South Bacan Formation volcanic rocks (BM188).

Plate 5.2c-d. XPL and PPL of the South Bacan Formation sandstone (BM227).

Plate 5.2e. XPL of the South Bacan Formation mudstone (BM218).

Plate 5.2f. XPL of the South Bacan Formation chert (BM208).

Scale for all photomicrographs = 0.5 mm.
to those of the Bacan Formation which has suffered prehnite-pumpellyite facies metamorphism (~240-330°C and ~2 kb). The lack of fabric development indicates a regional, static metamorphism due to burial.

5.6 WHOLE ROCK CHEMISTRY
Three samples (BM188, BM250 and BM282) have been analysed for bulk rock chemistry. All samples are in situ, except BM250. The purpose of this study is to decipher the tectonic setting and to compare the rocks from this formation with those from the Saleh Complex, Bacan Formation and Nusa Babi Monzodiorite.

5.6.1 Major Elements
Sample BM188 has the highest LOI and lowest total weight percent and may be indicative of the degree of alteration. Following Le Bas & Streckeisen (1991) and Basaltic Volcanism Study Project (1981), these samples are calc-alkaline trachy-basalt, basaltic andesites and andesite (Fig.5.4a,b).

Figs.5.5a-e show covariations of SiO₂ with CaO, MgO, Na₂O, TiO₂ and FeO*/MgO, with only MgO correlates well with SiO₂, suggesting that the rocks may be related by fractionation. The other covariations, however, show two trends, with the basaltic samples demonstrating increase of CaO, Na₂O, TiO₂ and FeO*/MgO with increasing SiO₂, followed by decreasing CaO, Na₂O, TiO₂ and FeO*/MgO towards the andesites, which may be a reflection of source heterogeneity or crustal contamination. These trends, however, should be taken cautiously due to the limited amount of samples.

Table 5.1 compares the average composition of the South Bacan Formation metavolcanic rocks with MORB, HKCA, IACA and BABB. Data are taken from Wilson (1989), except BABB which is from the Mariana Trough (DSDP Leg 60, site 453; Wood et al., 1981). Within the South Bacan Formation, BM250 has the highest SiO₂ and lowest TiO₂, FeO*, MgO contents, suggesting derivation from fractionated magma. The other rocks are most similar to the IACA and HKCA, particularly in TiO₂, MnO and CaO concentrations. All South Bacan Formation rocks have high Na₂O contents which may be an indication of alteration.

5.6.2 Trace Elements
To clarify the tectonic setting of the metavolcanic rocks, a comparison of the trace element and selected ratios with MORB, HKCA, IACA and BABB is provided in Table 5.2. The average values of rocks from the Saleh Complex (except SM10, which has characteristics of MORB; see
section 3.5.3), the Bacan Formation (XRF results only) and the Jojok Member are also compared. Due to similarities, in both major and trace elements between BM188 and BM282, the trace element values for these two samples are averaged.

Table 5.1 Comparison of the major elements of the metavolcanic rocks of the South Bacan Formation with rocks from known tectonic settings.

<table>
<thead>
<tr>
<th>Wt%</th>
<th>BM188</th>
<th>BM282</th>
<th>BM250</th>
<th>MORB</th>
<th>HKCA</th>
<th>IACA</th>
<th>BABB</th>
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<tr>
<td>SiO₂</td>
<td>50.08</td>
<td>52.69</td>
<td>61.38</td>
<td>50.67</td>
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<td>TiO₂</td>
<td>0.694</td>
<td>0.721</td>
<td>0.429</td>
<td>1.28</td>
<td>0.93</td>
<td>0.70</td>
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<tr>
<td>Al₂O₃</td>
<td>19.31</td>
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<td>FeO</td>
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<td>MnO</td>
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<td>MgO</td>
<td>5.18</td>
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<td>CaO</td>
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<td>2.16</td>
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<tr>
<td>K₂O</td>
<td>1.595</td>
<td>0.627</td>
<td>1.300</td>
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<td>P₂O₅</td>
<td>0.158</td>
<td>0.251</td>
<td>0.138</td>
<td>0.20</td>
<td>0.59</td>
<td>0.20</td>
<td>0.31</td>
</tr>
</tbody>
</table>

Although there are slight variations within the South Bacan Formation rocks, particularly the Zr and La contents, trace elements from all of them are clearly not comparable to MORB, specifically the Zr/Rb, Zr/Nb, La/Nb and Ba/La ratios, or to BABB, especially with the Zr/Rb, Zr/Nb, Y/Nb, La/Nb and Ba/La ratios. The closest resemblance is with IACA. The high La/Nb and Ba/La ratios are typical of island arc volcanics (Saunders et al., 1980; Arculus & Powell, 1986).

Despite the South Bacan Formation’s slightly lower trace element (except Sr) concentrations compared to those of the Saleh Metamorphic Complex, their ratios are comparable. The South Bacan Formation resembles closely the Bacan Formation, particularly BM188 and BM282, despite major element results suggesting that the South Bacan Formation is more acidic than the Bacan Formation and this is attributed to similarities in chemistry (IACA). Most of the trace element contents and their ratios (especially Zr/Nb, Y/Nb and Ba/La) of this formation are different from the Tawali Formation.
The South Bacan Formation

Figure 5.4a. Classification of South Bacan Formation volcanic rocks based on total alkali against SiO₂ contents. After Le Bas & Streckeisen (1991).

Figure 5.4b. Classification of South Bacan Formation volcanic rocks based on K₂O against SiO₂ contents. After Basaltic Volcanism Study Project (1981).
Figure 5.5a-e. Covariations of bulk rock SiO₂ against CaO, MgO, Na₂O, TiO₂, and FeO*/MgO.
Table 5.2 Comparison of the South Bacan Formation volcanic rocks' trace elements and their ratios with rocks of known tectonic settings and those from the Saleh Complex, the Bacan Formation and the Jojok Member. Data for comparison are from Wilson (1989), except BABB which are from Wood et al. (1981). Results in squared parentheses are from BAT values in Wilson (1989). So.Bacan are mean values of BM188 and BM282.

<table>
<thead>
<tr>
<th></th>
<th>So.Bacan</th>
<th>BM250</th>
<th>N-MORB</th>
<th>IACA</th>
<th>BABB</th>
<th>Saleh</th>
<th>Bacan</th>
<th>Jojok</th>
</tr>
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<tr>
<td>Sr</td>
<td>490.5</td>
<td>577.9</td>
<td>124</td>
<td>550</td>
<td>511</td>
<td>240</td>
<td>439</td>
<td>370.8</td>
</tr>
<tr>
<td>Rb</td>
<td>19.3</td>
<td>27.1</td>
<td>1.0</td>
<td>14</td>
<td>9.0</td>
<td>27.2</td>
<td>20.8</td>
<td>13.1</td>
</tr>
<tr>
<td>Zr</td>
<td>47.7</td>
<td>62.0</td>
<td>85</td>
<td>40</td>
<td>82</td>
<td>90.7</td>
<td>46</td>
<td>55.9</td>
</tr>
<tr>
<td>Nb</td>
<td>1.4</td>
<td>1.4</td>
<td>3.1</td>
<td>1.4</td>
<td>[8.0]</td>
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</tr>
<tr>
<td>Y</td>
<td>19.3</td>
<td>12.3</td>
<td>29</td>
<td>15</td>
<td>25</td>
<td>23.4</td>
<td>19.6</td>
<td>23.7</td>
</tr>
<tr>
<td>La</td>
<td>5.0</td>
<td>8.7</td>
<td>3.0</td>
<td>10</td>
<td>12.50</td>
<td>9.5</td>
<td>5.1</td>
<td>4.6</td>
</tr>
<tr>
<td>Nd</td>
<td>10.8</td>
<td>10.7</td>
<td>7.7</td>
<td>13</td>
<td>[13.1]</td>
<td>16.4</td>
<td>10.4</td>
<td>11.7</td>
</tr>
<tr>
<td>Zr/Rb</td>
<td>5.6</td>
<td>2.3</td>
<td>85</td>
<td>2.9</td>
<td>9.1</td>
<td>3.3</td>
<td>3.3</td>
<td>4.3</td>
</tr>
<tr>
<td>Zr/Nb</td>
<td>35.5</td>
<td>44.3</td>
<td>24.3</td>
<td>28.6</td>
<td>[16.3]</td>
<td>39.4</td>
<td>32</td>
<td>50.8</td>
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<tr>
<td>Zr/Y</td>
<td>2.5</td>
<td>5.0</td>
<td>2.9</td>
<td>2.7</td>
<td>3.3</td>
<td>3.9</td>
<td>2.2</td>
<td>2.4</td>
</tr>
<tr>
<td>Y/Nb</td>
<td>14.5</td>
<td>8.8</td>
<td>9.4</td>
<td>10.7</td>
<td>[3.8]</td>
<td>10.2</td>
<td>16.5</td>
<td>21.5</td>
</tr>
<tr>
<td>La/Nb</td>
<td>3.8</td>
<td>6.2</td>
<td>1.0</td>
<td>7.1</td>
<td>[1.0]</td>
<td>4.1</td>
<td>3.3</td>
<td>4.2</td>
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<tr>
<td>Ba/La</td>
<td>44.1</td>
<td>33.8</td>
<td>4.0</td>
<td>30</td>
<td>11.8</td>
<td>33.2</td>
<td>39.5</td>
<td>19.1</td>
</tr>
</tbody>
</table>

5.6.3 Tectonic Discrimination Diagrams

Figs.5.6a-c are ternary tectonic discrimination diagrams utilising immobile trace elements and major elements. On the Ti-Zr-Y diagram (Pearce & Cann, 1973), the South Bacan Formation rocks plot in the overlap fields of OFB and CAB; whereas on the Nb-Zr-Y diagram (Meschede, 1986) they plot in the fields of N-MORB & VAB and VAB & WPT. The TiO₂-MnO-P₂O₅ diagram (after Mullen, 1983) shows them plotting in the CAB field.

Fig.5.6d shows the MORB normalised diagram of Pearce (1982), where the South Bacan Formation rocks display the characteristics of arc rocks (humped LIL; Nb depletion; slightly humped Nb-Zr; fairly flat Zr-Sc, all below MORB values; and strong Cr depletion). Fig.5.6e is a chondrite normalised diagram (after Sun, 1980), showing the South Bacan Formation rocks having a pronounced Nb depletion; flat La-P distribution except K and Sr; flat Zr-Y with a slight Ti depletion; and a general decrease of normalised elemental abundance with increasing...
Figure 5.6a-c. Ternary tectonic discrimination diagrams applied to the South Bacan Formation.

Figure 5.6d-e. Spider diagrams for the South Bacan Formation volcanic rocks.
incompatibility, typical of arc-related volcanic rocks.

Figs. 5.7a-e are covariation diagrams (Cr against Y and Ce/Sr; Zr against Zr/Y and Ti; Ti/Y against Nb/Y) used to decipher tectonic setting. In all diagrams, the South Bacan Formation rocks plot mostly in the VAB field.

### 5.6.4 Implications

Trace element contents and ratios and tectonic discrimination diagrams all point to a volcanic arc origin for the South Bacan Formation. Comparison with older units reveals that the South Bacan Formation is similar to both the Saleh Complex and the Bacan Formation. The resemblance to the Bacan Formation is particularly striking. Despite this resemblance, K-Ar and micropalaeontological ages from both formations are different. The relationship with the Saleh Complex is less clear. Comparison with the Nusa Babi Monzodiorite will be discussed in Chapter Nine.

### 5.7 AGE DETERMINATION

Two mudstone samples (BM207 and BM218) have been dated using foraminifera (F.T. Banner, pers.comm, 1992) and indicate an Early Miocene (N4, Aquitanian-Langhian) age. Characteristic faunas are: *Globigerinoides* sp(p), *Globigerina* sp(p), *Dentoglobigerina* cf *altispira*, *?Paragloborotalia* and *Tenuitellids*.

Two samples have been dated using the K-Ar technique and they yield: 20.8 ± 2.0 Ma (BM250: float) and 7.5 ± 1.1 Ma (BM188) ages. The ~21 Ma age is consistent with the biostratigraphic age, while the 7.5 Ma age is interpreted to be a reset ages related to the activities of the Halmahera arc. The older age is not affected by thermal events because of the fresher nature of the sample and the fact that the material analysed is hornblende separate, with a higher closure temperature than the whole rock. Figs. 5.8a,b are isochron diagrams showing the age relationship between the South Bacan Formation and the similar aged Nusa Babi Monzodiorite (*Chapter Nine*).

<table>
<thead>
<tr>
<th>Sample</th>
<th>Material</th>
<th>Grain size (µ)</th>
<th>%K (1σ error)</th>
<th>Wt for Ar (g)</th>
<th><em>40Ar</em> (nl/g, 1σ error)</th>
<th><em>40Ar</em> (‰)</th>
<th>Age (Ma, 2σ error)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BM250</td>
<td>hb</td>
<td>125-250</td>
<td>0.218 ± 1.00%</td>
<td>1.0275</td>
<td>0.1768 ± 3.39%</td>
<td>73.34</td>
<td>20.8 ± 2.0</td>
</tr>
<tr>
<td>BM188</td>
<td>wr</td>
<td>250-500</td>
<td>1.289 ± 2.27%</td>
<td>1.0236</td>
<td>0.3767 ± 7.13%</td>
<td>87.38</td>
<td>7.5 ± 1.1</td>
</tr>
</tbody>
</table>

Table 5.3 Summary of the K-Ar results from the South Bacan Formation.
Figure 5.7a-e. Trace element covariations used to decipher the tectonic setting of the South Bacan Formation.
5.8 STRUCTURE
Small scale normal faults forming horst and graben structures were observed within this formation (Plate 5.1c). Differences in the six bedding attitudes from S. Wayatim (~160/40) with the four from S. Bibinoi (~310/45) suggest that the formation has been deformed. There are pervasive joints with directions, measured at different areas, of 000 & 065; 030 & 100; 078 & 162; and 105 & 190. The interjoint angles are 60-90°, indicating conjugate sets. The presence of large intrusions in S. Bibinoi area may be a cause of localised deformation, causing differences in bedding attitudes and joint directions in different areas.

5.9 SYNTHESIS AND TECTONIC SIGNIFICANCE
The South Bacan Formation consists of interbedded volcanic and volcaniclastic rocks erupted in an arc-related setting. This is supported by the whole rock geochemistry which clearly indicates a volcanic arc origin for the South Bacan Formation. Facies analyses of sedimentary rocks suggest deposition by intermediate turbidity currents which dissipated their energies as they travelled down slope. All the samples examined have been affected by low temperature prehnite-pumpellyite facies metamorphism. The lack of fabric development in the formation suggests a regional, static metamorphism, due to burial (~240-330°C, ~2 kb). Locally this formation shows the effects of hydrothermal alteration.

Isotopic and micropalaeontological age analyses yielded Late Oligocene-Early Miocene ages, with evidence of a thermal overprint related to the activities of the Halmahera arc. These ages are similar to the age of the Nusa Babi Monzodiorite (Chapter Nine).

The most striking feature of this formation is its lithological and chemical similarities to the Bacan Formation and before dating they were assumed to be the same age. There are two possibilities to account for the age differences between the Bacan and South Bacan Formations: [1] the two formations are part of a single formation spanning from Late Eocene to Early Miocene; the age differences may reflect either an older sequence at north Bacan or be a function of incomplete sampling and [2] these two are genuinely different formations, hence the separate age and spatial distribution. This resemblance probably reflects the fact that two formations with similar lithological character and tectonic setting have been metamorphosed and deformed under similar conditions at the same time.

The continuous arc scenario is favoured here because of the following reasons: [1] both formations post-date the Middle Eocene regional unconformity; [2] both formations are covered by the Middle
Figure 5.8a-b. Isochron diagrams for the South Bacan Formation. Data includes the similar-aged Nusa Babi Monzodiorite.
Miocene Limestone (Ruta Formation), which throughout the southern Philippine Sea Plate region signifies post collisional formation, and [3] metamorphism and deformation on both formations must have taken place before the Middle Miocene, as the Ruta Formation is not affected by similar style metamorphism and deformation. It is, therefore, more logical to group the Bacan and South Bacan Formations together between the Middle Eocene (creation of Philippine Sea Plate and rapid northward subduction of Australia) and Early Miocene (collision of Australia with Philippine Sea Plate) regional unconformities. This suggests that these formations are related to subduction of Australia under Philippine Sea Plate. The missing Oligocene sequence in Bacan may be explained by movement of active eruptive centres and the Oligocene Tawali Formation of Kasiruta may be part of this arc sequence (cf. Section 8.5.4).

Although this explanation is plausible, there are several arguments against it, particularly the difference in the lithology, metamorphic character (e.g. the lower grade of metamorphism suffered by the Tawali Formation relative to the underlying and overlying formations) and trace element differences between the Tawali and the Bacan/South Bacan Formations. The implication of these on tectonic setting will be discussed in more detail in Chapter Eleven.
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CHAPTER SIX
THE TAWALI FORMATION

6.1 INTRODUCTION
There are two members of the Tawali Formation: the Jojok Member and the Marikapal Member. The term Jojok Member is used for pillow basalts of arc-related character with associated hemipelagic rocks. The term Marikapal Member is applied to thick volcaniclastic turbidites associated with debris flows and red mudstones which typically overlie the lavas. These two members are easily distinguished in the field, but cannot be separated on aerial photographs. Since the mode of mapping in this region is principally by aerial photographs, the two members are considered as part of one formation. The regional distribution of this formation is shown in Fig.6.1.

This chapter is concerned with lithological and petrographical description of both the Jojok and Marikapal Members, mineral and whole rock chemistry of the Jojok Member, age determination and structural evidence from the formation.

6.2 SYNONYMY
The Tawali Formation has not been previously described. Yasin (1980) mapped it as part of his Oligocene Bacan Formation, and correlated it with the Bacan Formation in Bacan, Halmahera and Morotai. His Bacan Formation, however, includes rocks of different lithologies and ages (see Section 4.2). Hakim & Hall (1991) and Hall et al. (1991) referred to this formation as the Kasiruta Formation and correlate it with the Kahatola Formation of Halmahera and the Mayalabit Formation of Waigeo.

6.3 AERIAL PHOTOGRAPHY
Aerial photographic examination reveals that this formation is characterised by rugged, continuous ridges of similar height, which appear to be part of dissected plateaux. When cut by rivers, the valleys formed are very steep. The drainage pattern is similar to those of the Kaputusan, South Bacan and Bacan Formations.

6.4 TYPE AREAS
The Tawali Formation is named after Kasiruta (known locally as Tawali Besar). The type area is the northwestern coast of Kasiruta, across Dokoh, along which are exposed pillow basalts of the Jojok Member in contact with the overlying volcaniclastic sandstone, siltstone and mudstone of the Marikapal Member (Plate 6.1a). The type locality of the Jojok Member is Jojok off the
The Tawali Formation

northwestern coast of Kasiruta. The Marikapal Member is named after Tg. Marikapal on the southwestern coast of Kasiruta.

6.5 THE JOJOK MEMBER

The Jojok Member includes the oldest rocks seen on Kasiruta. Lava flows and pillows are dominant, with hemipelagic mudstone, siltstone and sandstone occurring between lavas.

6.5.1 Lithofacies

*Type Locality*

At the type locality of the Jojok Member is a sequence of lavas and sedimentary rocks about 30 m in thickness, divided into upper and lower units. The lower unit, >5 m, is composed of pillow lavas overlain by discontinuous layers of well bedded sandstone and siltstone ~2-4 m in thickness. The upper unit is composed of pillowed and massive lava with columnar jointing.

The older pillows have perfect tear-drop shapes and are up to 2 m across (Plate 6.1b). Lavas are porphyritic basalts with euhedral plagioclase laths, pyroxene ± altered olivine phenocrysts in a microcrystalline or glassy matrix. Some basalts are amygdaloidal, with either silica or zeolite fillings. Locally the pillows have glassy outer edges. All pillows indicate these rocks are the right way up and subhorizontal. Between the pillows there are common friable, slightly calcareous volaniclastic and hyaloclastite siltstones of various colours (white, yellow, red and green).

Above these lower pillow lavas is a discontinuous layer of interbedded red-brown volcaniclastic hemipelagic siltstone and sandstone (sand:silt ~20:80) which blankets the irregular upper pillow surface (Plate 6.1c). The maximum thickness of these are 4 m, although typically they are ~2 m thick. They have prominent thin bedding and scours and load structures may be developed at the base of some sandstone beds, which are normally graded. Within the sedimentary layer there is a ~0.15 m thick discontinuous glassy basaltic sill.

Above the sedimentary rocks are more pillow basalts, which change up section, over a distance of ~2 m, to massive lava flows characterised by vertical columnar jointing. Individual flows are up to ~6 m thick and locally have well developed radial joints. The lavas are of the same composition and mineralogy as the underlying pillows. At Jojok, this member is cut by a series of normal faults, some of which appear to be listric.
Figure 6.1. The distribution of the Tawali Formation in the Bacan Region.
Plate 6.1a. Exposure of the Tawali Formation across P. Dokoh. The Jojok Member is overlain by the Marikapal Member.

Plate 6.1b. Typical exposure of Jojok Member at near P. Dokoh. Pillows are right way up.

Plate 6.1c. Pillow lavas blanketed by pelagic sedimentary rocks exposed at P. Jojok.
The Tawali Formation

Lower and Upper Boundaries
The base of the Jojok Member was not observed. The upper contact with the Marikapal Member is depositional, and this is seen near Dokoh and on the southeastern coast of Muari (Plate 6.1a), where the Marikapal Member blankets the pillows. Micropalaeontological evidence suggests that there is a period of non-deposition between the two members (see Sections 6.5.4 and 6.6.3).

Thickness
Incomplete exposure and faulting hamper accurate assessment of the total thickness of the Jojok Member. The observable thickness of this member at Jojok is ~30 m, although the total thickness at Jojok is at least 70 m, deduced from the height of the island. Along the S. Leleojaya in northeastern Kasiruta this member, with ~26° dip, is seen for a total height of ~145 m. It is estimated that the thickness of this member is at least 200 m.

Lateral Extent and Variations
The Jojok Member is extensively exposed along the western coast of Kasiruta and eastern coast of Muari in the studied area. Outside the study area this member is exposed along the western coast of northern Halmahera (Hall et al., 1992) forming the westernmost exposure of the present-day Halmahera arc.

Along the S. Kasiruta, on southern Kasiruta, pillows appear to be coated with manganese. Associated with the pillows are the semi-precious "Bacan Stones", which are agate and chalcedony with limonite and haematite alteration. These silica-rich minerals occur mostly in veins, up to ~50 mm thick, and less commonly as amygdales. Fig.6.2 represents a sketch log of the exposure of this member at S. Tabadiku, south of Leleojaya Village, on northwestern Kasiruta.

6.5.2 Petrography and Mineral Chemistry

6.5.2.1 Petrography
Unless stated, all modal percentages are based on visual estimation. Petrographic analysis of the lavas of the Jojok Member reveals that the phenocrysts are clinopyroxene (~10%), plagioclase (~17%), opaque (~3%) and occasional altered olivine (~3%). The basalts from northern Kasiruta contain more plagioclase and clinopyroxene phenocrysts in a microcrystalline matrix, thus appearing to be more acidic in composition (Plate 6.2a,b). Most rocks are vesicular with either porphyritic or occasionally seriate (Plate 6.2c) or intersertal (hyalophitic) textures (Plate 6.2d). Zeolite, smectite and silica minerals fill the vesicles. Silica-rich veins cutting through the basalts
The Tawali Formation

Figure 6.2 Simplified log of the Tawali Formation, Jojok Member at Tabadiku River, North Kasiruta.
The Tawali Formation

Plate 6.2a. XPL of the Jojok Member from north Kasruta (BR94). Plate 6.2b. XPL of the Jojok Member from south Kasruta (BR165). Scale for all photomicrographs = 0.5 mm.

Plate 6.2c. XPL of the Jojok Member showing seriate texture (BR122).

Plate 6.2d. XPL of the Jojok Member showing hyalophitic texture (BR148).
The Tawali Formation

are composed of pure quartz with minor haematite, limonite, epidote, sulphides and green smectite. Locally they form conchoidal filling texture (Plate 6.3a).

The silty inter-pillow material contains angular, poorly sorted grains (Plate 6.3b) composed of pyroxene, planktonic foraminifera, lithic fragments, plagioclase (locally replaced by calcite), opaque and devitrified glass (locally altered to clay, forming ~75% of the rock).

Intra-lava lensoidal deposits consist of fine-grained sandstone, siltstone and mudstone. All grains are angular, with moderate to poor sorting and are matrix supported (Plate 6.3c). They consist of fragmented pyroxene (~2%), plagioclase laths (~5%), opaques (~2%), ~5% lithic fragments (basaltic, some of which are zeolitized) and rarely broken fragments of green hornblende. Planktonic foraminifera are also present (~5%). The matrix is a mixture of clay minerals and micrite (~75%). Secondary minerals are calcite (~5%) and zeolites (~3%) both occurring in veins. The glassy basalt sill consists of cracked glass (~85%), which is locally devitrified (Plate 6.3d). Subhedral plagioclase (~10%) with simple twinning and subhedral opaques (~5%) are filling the interstitial spaces.

6.5.2.2 Mineral Chemistry

Eight samples (BR94, BR97, BR122, BR125, BR148, BR153, BR165, BM92) were analysed for their mineral chemistry. Fig.6.3 is a location map for samples used in the analytical work.

**Plagioclase**

Plagioclase occurs as phenocrysts and as groundmass microlites. Phenocrysts are euhedral laths, up to 0.03 mm across, with polysynthetic twinning and occasional zoning. Glass inclusions are present locally, forming at the centre of the crystals (Plate 6.4a), and more rarely along the crystal rims (Plate 6.4b) or randomly distributed, forming a sieve texture.

Phenocryst compositions (n=40) are An_{52.95} (average An_{55}: bytownite), with <4% Or component, and plot along a continuous line on the An-Ab-Or ternary diagram (Fig.6.4), suggesting various degrees of albitisation. In general, the larger phenocrysts appear to be more calcic than the microphenocrysts and groundmass, which could reflect the resistance of primary phases to the albitisation associated with very low grade metamorphism (Liou et al., 1987) or it could be a crystallization sequence. Variation of An content within a sample is common, reflecting variation in local bulk compositions and metastable equilibria (Liou et al., 1987).
Figure 6.3 Location of samples from the Tawali Formation used in analytical work

Key:
- Marikapal Member
- Jojok Member
Plate 6.3a. PPL of green silica-rich vein showing conchoidal filling texture (BM484).
Plate 6.3b. XPL of silty inter-pillow material with planktonic foraminifera (BR158).

Scale for all photomicrographs = 0.5 mm.

Plate 6.3c. XPL of interlava sandstone (BR167).
Plate 6.3d. PPL of glassy basaltic sill (BT53).
Zoning is gradational and only rarely abrupt. A grain in BR153 shows sharp oscillatory zoning with an NRM pattern and a general core-rim normal zoning (BR153P7-9). Occasional glass inclusions in the phenocrysts, re-calculated as plagioclase, vary between An_{40-80} (average An_{64}, n=7). They define the boundary of zonation (Plate 6.4c), reflecting rapid crystal growth and enhanced supercooling (Morrice & Gill, 1986). The high An content, complex pattern of zoning and numerous glass inclusions are all characteristic of arc-related rocks (Gill, 1981). Inclusions of clinopyroxene and opaque minerals occur in plagioclase phenocrysts and are discussed in sections below.

Plagioclase microlites form the microcrystalline matrix together with anhedral clinopyroxene and anhedral opaque grains (Plate 6.5a). They are compositionally similar to co-existing microphenocryst phase (An_{46-97}; average An_{82}, n=7).

**Clinopyroxene**

Pale olive green clinopyroxene is euhedral-subhedral in form and occurs both as phenocrysts (maximum size: 1.5 mm, mode: ~0.5 mm), often forming glomerocrysts, and as inclusions in plagioclase. Twinning is common, whereas zoning was not observed.

Clinopyroxene phenocrysts range from En_{37.5}Fs_{13.25}Wo_{31.44} (average En_{40}Fs_{18}Wo_{40}; n=31) while inclusions vary from En_{30-46}Fs_{18-31}Wo_{34-40} (average En_{37}Fs_{35}Wo_{38}; n=5). The Mg# of the phenocrysts vary from 0.63-0.85 (average 0.75), whereas that of the inclusions range from 0.56-0.76 (average 0.64). Most analyses are augite, with only minor salite (Fig.6.5a). The phenocrysts are enriched in Mg, depleted in FeO*+Mn and only slightly higher in Ca relative to the inclusion phase, indicating a tholeiitic liquid fractionation trend towards Fe-enrichment. Fig.6.5b is modified from the tectonic discrimination diagram of Nisbet & Pearce (1977) showing that the pyroxene phenocrysts are of volcanic arc origin.

**Olivine**

Euhedral-subhedral olivine occurs as phenocrysts up to ~1.3 mm across (mode: ~0.8 mm). It is altered to brown chlorite, dark red-brown 'iddingsite' and yellow-green smectite (bowlingite). Those rocks containing chlorite (e.g. BR125) have anomalously high whole rock K_{2}O contents, whereas the iddingsite-containing rock (BR153) has a low K_{2}O content. The occurrence of chlorite or smectite, is therefore, predicted to be related either to the bulk rock composition or due to K_{2}O enrichment from interaction with sea water (Alt et al., 1986).
The Tawali Formation

Plate 6.4a. XPL of plagioclase with glass inclusions in the crystal centre (BR153).

Plate 6.4b. XPL of plagioclase with glass inclusions in the rim of the crystal (BR153).

Plate 6.4c. XPL of zoned plagioclase with glass inclusions at the zonation boundary (BR153).

Plate 6.5a. Plagioclase microlites forming the matrix of BR122.

Plate 6.5b. Spinel with simplectic texture (BR148).
Scale for all photomicrographs = 0.5 mm.
Fe-Ti Oxides

Fe-Ti oxides occur as euhedral-subhedral phenocrysts with square sections and occasional simplectic texture (Plate 6.5b). Fig.6.6a demonstrates that most phenocrysts plot on the ulvöspinel-magnetite tie-line. Phenocrysts in the Jojok Member have Cr# between 0-0.07 (mean=0.03) and Fe3# of 0.84-0.92 (mean=0.88; n=22), and so are low in Cr, high in Ti (0.61-3.55 stoichiometric Ti, excluding BR9703-5; mean=1.20; n=25) and Fe3+ indicating a relatively differentiated source.

Oxide inclusions occur in both plagioclase and pyroxene. These are more Fe2+ and Ti rich than the phenocrysts, indicating either ionic exchange between the host mineral and oxide or an inherited difference in chemistry between the phenocryst and inclusion phases. The latter interpretation is preferred due to the relatively Fe- and Ti-poor nature of the host minerals, and is supported by the difference in Cr# (0-0.04, mean 0.01) and Fe3# (0.87-0.91, mean 0.90; n=4) between the phenocrysts and inclusion phases. The lower Cr and higher Fe2+ and Fe3+ content reflect a more evolved magma composition.

Glass

Groundmass glass is mostly replaced by brown-yellow or green smectite, or devitrified to quartz. Fresh, unaltered glass is found rarely and is either brown or black in colour, although compositional difference between these appears to be gradational. These are distinguished from plagioclase by their higher MgO (>0.33 wt%) and FeO (>0.4 wt%) contents and lower total (<94 wt%). Dark brown glass is enriched in FeO and depleted in Al2O3 relative to the black glass. Fig.6.6b exhibits this compositional difference in a Harker-type diagram. The whole rock chemistry reveals that the rocks containing brown glass have the lowest LOI within the Jojok Member. Samples containing fresh glass are not zeolite-bearing, indicating a lower degree of metamorphism. Analyses of glass inclusions in phenocrysts show that chemically they are similar to the groundmass.

Silica and Calcite

Secondary silica occurs as a product of devitrified glass and in amygdales and veins. Calcite is observed in BR94, BR125 and BR153, occurring in veins, replacing plagioclase or, in close association with chlorite, replacing olivine.

Smectites

Water-containing phyllosilicates are generally the first to form during metamorphic recrystallization
of igneous rocks; and in the Tawali Formation these include the smectite group and chlorite. Smectite is an ubiquitous secondary mineral, replacing the glassy matrix, plagioclase and olivine, filling veins and vesicles. The very fine grained smectite has high birefringence and is bright green or pleochroic yellow-brown-green. There is no difference in the habit of the different coloured forms of smectite; all form concentric linings to the vesicles, normally surrounding zeolite minerals. In BR125 green smectite is rimmed by brown variety. Glass inclusions are common in the middle of the yellow variety, resulting in a dark-brown appearance.

Chemically three types of smectite are distinguished, summarised in Table 6.1. These smectites resemble saponite, celadonite (Andrew, 1980) and nontronite (Deer et al., 1992). Type I and II are comparable to the type I and II smectites of Hakim (1989). The saponite is of Fe-rich, Mg-poor variety which is a product of non-oxidative diagenesis (Andrew, 1980). Celadonite often occurs as a result of interaction of basalt with sea water (Andrew, 1980), which may explain the high Al₂O₃ and MgO content. Fig.6.6c shows the smectite compositions on a ternary diagram.

<table>
<thead>
<tr>
<th>Wt %</th>
<th>Type I</th>
<th>Type II</th>
<th>Type III</th>
</tr>
</thead>
<tbody>
<tr>
<td>Al₂O₃</td>
<td>7-10</td>
<td>7-11</td>
<td>&lt;6</td>
</tr>
<tr>
<td>FeO*</td>
<td>14-18</td>
<td>12-15</td>
<td>18-23</td>
</tr>
<tr>
<td>MgO</td>
<td>4-9</td>
<td>11-17</td>
<td>3-5</td>
</tr>
<tr>
<td>CaO</td>
<td>0.2-1.5</td>
<td>0.7-2</td>
<td>0.1</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.2-1</td>
<td>0.1</td>
<td>0.2</td>
</tr>
<tr>
<td>K₂O</td>
<td>1.8</td>
<td>0.3-3</td>
<td>3.8-3</td>
</tr>
<tr>
<td>XPL Colour</td>
<td>Bright Green</td>
<td>Yellow &amp; Brown-Green</td>
<td>Green-Brown</td>
</tr>
<tr>
<td>Possible equivalent</td>
<td>Saponite</td>
<td>Celadonite</td>
<td>Nontronite</td>
</tr>
</tbody>
</table>

Chlorite
At low pressures, chlorite becomes the dominant phyllosilicate at ~230°C (Tómasson & Kristmannsdóttir, 1972). Red-brown, poorly crystalline chlorite is detected only in BR125 as a replacement of olivine, and is distinguished from smectite by low SiO₂ (<25 wt%). Its association with calcite is a common product of olivine alteration (Hakim, 1989). There are two types of chlorites detected: low Si, high Fe and high Si, low Fe, similar to the findings of Hakim (1989).
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The presence of K, Ca and Na in these chlorites suggests the presence of illite and smectite within chlorite layers, typical of sub-greenschist metamorphism (Offler & Aguirre, 1984).

**Zeolites**

Heulandite and stilbite are similar zeolite minerals, structurally, chemically, optically and by mineralogical association (Gottardi & Galli, 1985). Both are present in samples from northern Kasiruta (more acidic, higher whole rock SiO₂ contents), co-existing in veins and amygdales and replacing plagioclase. In BR97, both zeolites form tabular and radial aggregates, locally twinned; colourless under PPL, and 1° grey to isotropic in XPL. Heulandite has a higher SiO₂ (57-61 wt%), and lower Al₂O₃ content (13-19 wt%), compared to stilbite (SiO₂ = 50-55 wt% and Al₂O₃ = 15-21 wt%). There is no obvious whole rock chemical differences between rocks containing heulandite and stilbite.

Thomsonite and mesolite are members of the natrolite group of zeolites (Gottardi & Galli, 1985) and are found in northern Kasiruta only. They both occur in veins, replacing plagioclase and as amygdales. Thomsonite occurs with analcite, chlorite, smectite and calcite (BR125). It is fibrous and forms radial fan-shaped aggregates. In PPL it is colourless, while in XPL it shows 1° grey-yellow colour. The SiO₂ content ranges from 36-43 wt%, Al₂O₃ from 25-31 wt%, and the atomic Ca:Na ratio ranges from 2.4-3.3. Mesolite occurs with heulandite, stilbite, smectite and calcite (BR94). In PPL it is white, forming acicular crystals, while in XPL it shows 1° grey-yellow colours. The SiO₂ content ranges from 44-49 wt%, whereas Al₂O₃ ranges from 24-26 wt%, and the atomic Ca:Na ratio ranges from 6.3-9.5. Rocks bearing thomsonite and mesolite have the highest contents of K₂O, Na₂O and LOI.

Colourless, isotropic analcite is found associated with thomsonite, chlorite, smectite and calcite (BR125, BM92). It occurs in veins and amygdales and replacing plagioclase. It is mostly anhedral, locally showing twin laminae. Chemically it is characterised by SiO₂ between 54-58 wt%, Al₂O₃ from 20-23 wt% and particularly high Na₂O (9-12 wt%). Rocks containing analcite are high in K₂O and LOI.

6.5.2.3 Implications

The primary mineralogy of the Jojok Member rocks points to an arc-related volcanic sequence which has been affected by zeolite facies metamorphism. Based on the presence of inclusions, the order of crystallization in the Jojok Member is oxides, olivine, plagioclase ± pyroxene.
Figure 6.4. Ternary diagram showing plagioclase compositions from the Jojok Member.

Figure 6.5a. Quadrilateral diagram of pyroxenes in the Jojok Member. Figure 6.5b. Clinopyroxene discrimination diagram applied to the Jojok Member.

Figure 6.6a. Ternary diagram showing the compositions of Fe-Ti oxides in the Jojok Member. Figure 6.6b. Compositional differences between the black and brown glass in a Harker-type diagram. Figure 6.6c. Ternary diagram showing the compositions of smectite in the Jojok Member.
Secondary mineral assemblages of bytownite + smectite + (calcite or heulandite + stilbite or quartz) ± (mesolite or analcite + thomsonite + chlorite) are present. Liou et al. (1987) recognised the assemblage of chlorite/smectite + heulandite (or stilbite) + quartz + analcite as the low-T zeolite facies (<180°C), with the appearance of Ab defining the onset of the high-T zeolite facies (180°C). Based on the assemblages, the Jojok Member is interpreted to be affected by transitional low-T to high-T zeolite facies metamorphism, at 180°C and <2 kb. The Jojok assemblage is plotted on the PT diagram of Liou et al. (1987) in Fig.6.7.

The very low-grade metamorphism of the Jojok Member can be attributed either to ocean-floor alteration, very shallow level burial or both. The burial metamorphic Wakatipu Belt of New Zealand contains the assemblage smectite + heulandite + quartz + analcite + illite ± calcite in the so-called Heulandite-Analcite facies (Coombs et al., 1959; Kawachi, 1975). The presence of similar secondary mineral assemblages in the Marikapal Member above the Jojok Member and the similarity with the Wakatipu Metamorphic Belt argue for the burial mechanism. Metamorphism in the Jojok Member is therefore interpreted to be a product of shallow level burial.

Zeolite-bearing rocks occur in northern Kasiruta only. This might be a product of sampling bias or a function of more acidic bulk compositions in northern Kasiruta, but the detailed distribution of metamorphic facies is beyond the scope of present study.

6.5.3 Whole Rock Chemistry

The main purpose of the whole rock chemistry study was to characterise volcanic rock types and utilise published tectonic discrimination diagrams to identify the tectonic setting of the Jojok Member. Both major and trace elements were used to establish element mobility on alteration and to assess the reliability of tectonic discrimination diagrams. Detailed discrimination of magmatic sources and the processes of magmatic modification (e.g. assimilation, fractional crystallization, partial melting, contamination) are beyond the scope of this study.

Seven samples (BR94, BR97, BR122, BR125, BR148, BR153 and BR165) were chosen for the whole rock analyses, based on sample freshness and location. All samples analysed were found in situ. Samples BR148, BR153 and BR165 were collected from the type locality.
Figure 6.7. Metamorphic conditions of the Jojok Member plotted in a petrogenetic grid of Liou et al. (1987).
6.5.3.1 Major Elements
To assess the degree of alteration, a plot of alkalis and LILE against HFS was used (Fig. 6.8.; see section 4.6.1), from which it is apparent that K, CaO, and Na₂O all correlate well with Zr, except the two samples with the lowest Zr content (BR94, BR125), suggesting significant alteration. This is consistent with the fact that these samples have the highest LOI. These two samples are therefore interpreted with caution in the following discussion due to their possible element mobility. The strong correlation of Y and Zr suggests a co-genetic suite, with the possible exception of BR94.

Fig. 6.8 also indicates that the LILE are not particularly mobile, and thus volcanic classification using LILE variations can be utilised. Following Le Bas & Streckeisen (1991) and Basaltic Volcanism Study Project (1981), the Jojok Member rocks are mostly calc-alkaline basalt and andesite except sample BR148 which is a tholeiitic basalt (Fig. 6.9a,b). Rocks from northern Kasiruta have a higher SiO₂ content than the rest, but due to the highly porphyritic nature of the samples, the whole rock chemistry may not reflect the liquid composition.

The average major element composition of the Jojok Member is compared to the average Atlantic MORB, IAT, IACA and island arc High-K Calc-Alkaline basalt (HKCA), in Table 6.2. The high Al₂O₃ content of the mean Jojok Member composition reflects the highly feldspar-phyric nature of the rocks. It is also noticeable that the Jojok rocks have very low MgO and high FeO* contents, suggesting a liquid evolved through extensive fractional crystallization of olivine from primary magma (Roeder & Emslie, 1970), which is characteristic of island arc rocks or indicative of an enriched magma source (Wilson, 1989). The first interpretation is preferred due to similarities of major elements with the IACA and HKCA. This is further illustrated in Fig. 6.9c, showing the fractional crystallization trend of the Jojok Member. The low TiO₂ contents of the Jojok rocks is typical of island arc volcanism (e.g. Morrice & Gill, 1986; Gill, 1981).

6.5.3.2 Trace Elements
To decipher the tectonic setting of the Jojok Member, a comparison of the mean Jojok Member trace element composition with N-MORB, E-MORB, IACA and Back-Arc Tholeiite (BAT) basalts is provided in Table 6.3.
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Table 6.2 Mean bulk composition of non-altered Jojok Member rocks compared to basalts from different tectonic settings.

<table>
<thead>
<tr>
<th>Wt%</th>
<th>JOJOK (Mean)</th>
<th>JOJOK (Range)</th>
<th>MORB (Atlantic)</th>
<th>IAT</th>
<th>IACA</th>
<th>HKCA</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>51.16</td>
<td>48.41-56.78</td>
<td>50.67</td>
<td>49.20</td>
<td>49.40</td>
<td>51.00</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.92</td>
<td>0.75-1.10</td>
<td>1.28</td>
<td>0.52</td>
<td>0.70</td>
<td>0.93</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>16.82</td>
<td>14.08-18.92</td>
<td>15.45</td>
<td>15.30</td>
<td>13.29</td>
<td>13.60</td>
</tr>
<tr>
<td>FeO*</td>
<td>11.13</td>
<td>9.35-13.60</td>
<td>9.67</td>
<td>9.00</td>
<td>10.15</td>
<td>8.11</td>
</tr>
<tr>
<td>MnO</td>
<td>0.17</td>
<td>0.13-0.23</td>
<td>[0.15]</td>
<td>0.18</td>
<td>0.20</td>
<td>0.14</td>
</tr>
<tr>
<td>MgO</td>
<td>4.74</td>
<td>3.13-5.79</td>
<td>9.05</td>
<td>10.1</td>
<td>10.44</td>
<td>12.50</td>
</tr>
<tr>
<td>CaO</td>
<td>9.45</td>
<td>7.63-10.82</td>
<td>11.72</td>
<td>13.00</td>
<td>12.22</td>
<td>7.92</td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.25</td>
<td>2.6-3.61</td>
<td>2.51</td>
<td>1.51</td>
<td>2.16</td>
<td>2.67</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.96</td>
<td>0.39-1.87</td>
<td>0.15</td>
<td>0.17</td>
<td>1.06</td>
<td>2.37</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.20</td>
<td>0.09-0.24</td>
<td>0.20</td>
<td>0.06</td>
<td>0.20</td>
<td>0.59</td>
</tr>
</tbody>
</table>

Table 6.3 A comparison of average selected trace element concentrations in N-MORB, E-MORB, Back-Arc Tholeiite and Island Arc Calc-Alkaline (Wilson, 1989) with Jojok Member. LOD = Limit Of Detection.

<table>
<thead>
<tr>
<th>Element (ppm)</th>
<th>JOJOK (Fresh)</th>
<th>JOJOK (Range)</th>
<th>JOJOK (Altered)</th>
<th>N-MORB</th>
<th>E-MORB</th>
<th>BAT</th>
<th>IACA</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rb</td>
<td>13.1</td>
<td>4.1-33</td>
<td>16.9</td>
<td>1.0</td>
<td>3.9</td>
<td>6</td>
<td>14</td>
</tr>
<tr>
<td>Ba</td>
<td>87.7</td>
<td>34-186.3</td>
<td>224.4</td>
<td>12</td>
<td>68</td>
<td>77</td>
<td>300</td>
</tr>
<tr>
<td>Nb</td>
<td>1.1</td>
<td>0.8-1.5</td>
<td>0.6</td>
<td>3.1</td>
<td>8.1</td>
<td>8</td>
<td>1.4</td>
</tr>
<tr>
<td>La</td>
<td>4.6</td>
<td>0.9-8.7</td>
<td>4.1</td>
<td>3.0</td>
<td>6.3</td>
<td>7.83</td>
<td>10</td>
</tr>
<tr>
<td>Ce</td>
<td>17.3</td>
<td>10.2-28.8</td>
<td>15.3</td>
<td>9.0</td>
<td>15.0</td>
<td>19.0</td>
<td>23</td>
</tr>
<tr>
<td>Sr</td>
<td>370.8</td>
<td>160.9-771.4</td>
<td>1218.4</td>
<td>124</td>
<td>180</td>
<td>212</td>
<td>550</td>
</tr>
<tr>
<td>Nd</td>
<td>11.7</td>
<td>6.5-17.3</td>
<td>11.1</td>
<td>7.7</td>
<td>9.0</td>
<td>13.1</td>
<td>13</td>
</tr>
<tr>
<td>Zr</td>
<td>55.9</td>
<td>38.5-80.1</td>
<td>34.7</td>
<td>85</td>
<td>75</td>
<td>130</td>
<td>40</td>
</tr>
<tr>
<td>Y</td>
<td>23.7</td>
<td>17.7-27.7</td>
<td>26.4</td>
<td>29</td>
<td>22</td>
<td>30</td>
<td>15</td>
</tr>
<tr>
<td>Th</td>
<td>&lt;LOD-1.1</td>
<td>-</td>
<td>0.20</td>
<td>0.55</td>
<td>-</td>
<td>1.1</td>
<td></td>
</tr>
</tbody>
</table>

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Figure 6.8. Covariations of bulk alkali and LIL contents with HFS (Zr) in the Jojok Member to determine the degree of alteration. Covariation of Y and Zr is to establish whether the samples are co-genetic.

Figure 6.9a. Classification of Jojok Member volcanic rocks based on total alkali against SiO₂ contents. After Le Bas & Streckeisen (1991).

Figure 6.9b. Classification of Jojok Member volcanic rocks based on K₂O against SiO₂ contents. After Basaltic Volcanism Study Project (1981).

Figure 6.9c. Covariations of Zr with MgO and FeO* contents.
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It is apparent that for most elements, the trace elements have not been significantly affected by alteration. Sr, Rb and Ba are the exceptions; Sr and Ba are markedly enriched in the altered rocks and hence are considered unrepresentative of primary trace element geochemistry. The low Th and Nb concentrations, as well as the values for Rb, Nd and Zr suggest an arc-related origin, but La, Ce and Y compare with E-MORB values. To clarify this, a comparison of trace element ratios is provided in Table 6.4.

Table 6.4. Comparison of trace element ratios of Jojok Member rocks and the average IACA, BAT and E-MORB.

<table>
<thead>
<tr>
<th></th>
<th>BR94*</th>
<th>BR97</th>
<th>BR122</th>
<th>BR125*</th>
<th>BR148</th>
<th>BR153</th>
<th>BR165</th>
<th>IACA</th>
<th>BAT</th>
<th>E-MORB</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zr/Nb</td>
<td>66.20</td>
<td>47.70</td>
<td>61.62</td>
<td>45.37</td>
<td>48.13</td>
<td>44.27</td>
<td>43.13</td>
<td>28.57</td>
<td>16.25</td>
<td>9.25</td>
</tr>
<tr>
<td>Zr/Y</td>
<td>0.90</td>
<td>1.96</td>
<td>3.05</td>
<td>2.26</td>
<td>2.18</td>
<td>2.15</td>
<td>2.34</td>
<td>2.67</td>
<td>4.33</td>
<td>3.41</td>
</tr>
<tr>
<td>Y/Nb</td>
<td>73.60</td>
<td>24.30</td>
<td>20.23</td>
<td>20.13</td>
<td>22.12</td>
<td>20.64</td>
<td>18.47</td>
<td>10.71</td>
<td>3.75</td>
<td>2.72</td>
</tr>
<tr>
<td>Nb/Ce</td>
<td>0.02</td>
<td>0.03</td>
<td>0.06</td>
<td>0.11</td>
<td>0.08</td>
<td>0.10</td>
<td>0.12</td>
<td>0.06</td>
<td>0.42</td>
<td>0.54</td>
</tr>
<tr>
<td>Nd/Ce</td>
<td>0.68</td>
<td>0.60</td>
<td>0.64</td>
<td>0.87</td>
<td>0.64</td>
<td>0.88</td>
<td>0.77</td>
<td>0.57</td>
<td>0.69</td>
<td>0.60</td>
</tr>
<tr>
<td>La/Nb</td>
<td>16.0</td>
<td>8.7</td>
<td>5.54</td>
<td>0.38</td>
<td>1.13</td>
<td>2.91</td>
<td>2.07</td>
<td>7.14</td>
<td>0.98</td>
<td>0.78</td>
</tr>
<tr>
<td>Ba/La</td>
<td>48.40</td>
<td>21.41</td>
<td>16.97</td>
<td>205.67</td>
<td>54.89</td>
<td>14.56</td>
<td>10.97</td>
<td>30.0</td>
<td>9.83</td>
<td>10.79</td>
</tr>
</tbody>
</table>


Table 6.4 shows that trace element ratios in the Jojok rocks are closer to the IACA than to the BAT and E-MORB. The high La/Nb ratio (1.12-8.7) is typical of island arc volcanic rocks (Saunders et al., 1980), as is the high Ba/La (>3) (Arculus & Powell, 1986), although the latter ratio should be treated with caution due to the mobility of Ba.

6.5.3.3 Tectonic Discrimination Diagrams

On the Ti-Zr-Y diagram of Pearce & Cann (1973), the Jojok Member rocks plot in the field which includes Calc-Alkaline Basalt or Low K Tholeiites or Ocean Floor Basalt (Fig.6.10a). On the Nb-Zr-Y diagram (Meschede, 1986), the rocks plot in the Volcanic Arc Basalt - N-MORB overlap field (Fig.6.10b). The TiO2-MnO-P2O5 diagram (Mullen, 1983) shows the Jojok rocks plotting in the field of Island Arc Tholeiite and Calc-Alkaline Basalt (Fig.6.10c). From these three diagrams, the most likely tectonic setting for the Jojok Member is an island arc.

Using the MORB normalised "spider" diagram of Pearce (1982), the Jojok Member rocks show a "hump" (enrichment) on the LIL elements, a pronounced Nb depletion, a fairly flat Ce-Sc distribution and strongly depleted Cr (Fig.6.11a). All these are characteristic of arc-related rocks. When normalised against chondrite (after Sun, 1980), the rocks display a "spiky" pattern, with Rb,
Figures 6.10a-c. Ternary tectonic discrimination diagrams applied to the Jojok Member.
Figures 6.11a-b. Spider diagrams, normalised to MORB and chondrite, of the Jojok Member.
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K and Sr enrichment and Nb and La depletion. The elemental distribution from Sr to Y is relatively flat with a slight convex upward pattern (Fig.6.11b). The enrichment and depletion spikes and the concave upward pattern is again typical of a volcanic arc origin.

Figs.6.12a-d are covariation diagrams of immobile elements (Pearce, 1982). On the Cr against Y and Ce/Sr, Zr against Ti, and Ti/Y against Nb/Y diagrams, the Jojok rocks plot mostly in the Volcanic Arc Basalt field. Fig.6.12e is a similar diagram using Zr against Zr/Y (Pearce & Norry, 1979) in which the Jojok Member rocks are identified as Island Arc Tholeiites.

Fig.6.12f is the FeO*/MgO versus TiO2 diagram (Hawkins & Evans, 1983) which includes the Western Pacific arc rocks of Palau-Kyushu (Mattey et al., 1980) and the back-arc rocks of West Philippine Basin and North Mariana (Stern et al., 1990). The Jojok Member rocks clearly plot in the Island Arc Tholeiite field, which is consistent with the results from Table 6.4.

6.5.3.4 Implications

Major and trace elements suggest that the Jojok Member volcanic rocks are highly differentiated Island Arc Calc-Alkaline Basalts. Although there is chemical evidence that the Jojok Member most likely erupted in the arc, recent studies show that chemical evidence alone cannot be used to determine the tectonic setting of an arc-related volcanic rocks (e.g. Evans et al., 1991 and references therein). The deep marine pillow basalt of the Jojok Member overlain by the thick turbidite sequence of the Marikapal Member makes the most likely setting for the Tawali Formation is a back-arc basin.

6.5.4 Age Determination

At the type locality, nannofossils (E.Finch & P.T.Robertson Utama, pers.comm) from the inter-pillow material and the inter-lava deposits indicate an age range for the Jojok Member of Early-Late Oligocene (NP23). Foraminifera from the inter-pillow material indicate an Early Oligocene age (P18) (S.J.Roberts, pers.comm, 1990). At some localities (Muari and Jojok), these two determinations overlap at ~33.3 ± 0.3 Ma (Fig.6.13). Micropalaeontological evidence suggests an open marine depositional environment. The following fossils are the index fossils; nannoflora: *Cyclicargolithus abisectus*, *C. floridanus*, *Discoaster deflandrei*, *Helicosphaera ampliapertera*, *Sphenolithus distentus*, *S. predistentus*, *S. moriformis* and *Zygrahablithus bijugatus*; planktonic foraminifera: *Globoquadrina tripartita*, *Globegerina praebulloides leroy*, *Gg. yeguaensis*, *Gg. tauriensis*, *Gg. praeapsis*, *Gg. ampliapertura* and *Gg. euapertura*. 

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Fig 6.12f

FeO/MgO vs TiO2 (wt%) graph with data points for different regions and markers for regions: Japik, W. Philippine, North Mariana, Palau-Kyushu.
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Figures 6.12a-e. Covariations of immobile trace elements to determine the tectonic setting of the Jojok Member. See text for details.
The Tawali Formation

Figure 6.13 Summary of the Tawali Formation Biostratigraphic Dates.
6.6 MARIKAPAL MEMBER

6.6.1 Lithofacies

The Marikapal Member is exposed discontinuously along the ~7.5 km coastal section from Tg. Marikapal to Kakupang Village on the southern coast of Kasiruta. This member is named after Tg. Marikapal in southwestern Kasiruta, just south of which there is an approximately 30 m sequence of volcaniclastic breccia, sandstone and siltstone (Plate 6.6a). The sequence around this area contains more lithofacies variation than elsewhere, hence its designation as the type locality. The description below is taken from P. Ngaimadodera, southwest of Tg. Marikapal, where a ~12.5 m sequence is exposed. The rocks are well lithified in most outcrop areas.

Type Locality

The breccias (facies A2.3) are very thickly bedded (>2 m) and show normal grading. Clasts are angular to subangular, moderately sorted and are predominantly of basaltic composition, similar to the Jojok Member rocks. Clast sizes range from 0.01-0.15 m across and no imbrication is observed (Plate 6.6b). The matrix is coarse sandstone, similar to the unit above, as described below.

The volcaniclastic sandstones (facies B1.1) are massive and thickly bedded (>1 m) and are the dominant lithology (~70-95% on the southern coast of Kasiruta). Sorting is poor-medium, with angular-subangular clasts of probable basic composition and subordinate limestone. Sole marks occur at the contact with mudstone.

The fine-grained sandstones (C2.2) are thinner bedded (<1 m) and are intercalated with, or grade into, the siltstone and mudstone units. Sorting is poor-medium, while the clasts are angular-subangular. Normal grading is present in thicker basal unit (Ta). Parallel and some minor cross stratification are observed (Tb, Tc). Laminations are up to 0.3 m thick. The base of the sandstone beds show load structures.

The siltstones are very thinly bedded, forming composite medium sized beds. Bioturbation (parallel to bedding, ~50 mm long; Plate 6.6c), convolute bedding, load cast and flame structures are present (Tc, Td). Siltstones locally grade into mudstones, both of which are calcareous (facies D2.2). Fig.6.14 is a sketch log of this member at Ngaimadodera, showing the different facies and the classical Bouma Ta-d.
Plate 6.6a. Typical exposure of the Marikapal Member at P. Ngaimadodera.

Plate 6.6b-c. Details of the Marikapal Member conglomerate and mudstone at P. Ngaimadodera.

See text for discussions.
Figure 6.14. Simplified Log of the Marikapal Member, Tawali Formation, P.Ngaimadodera, Southwest Kasiruta.

Massive coarse sandstone/microconglomerate. Contain clasts of limestone and Jojok type volcanics. No apparent grading. Shows spheroidal weathering in places.

Bouma Ta

Convolute lamination, loading structures at the base.

Bouma Td

Fine sandstone, shows parallel, cross and convolute lamination on mm scale. Burrows are present on an angularblock, 50 cm across, composed of the same material.

Bouma Tbc

Monomict matrix supported breccia: clasts are angular, up to 4 cm across, Jojok type volcanic, some of which shows chloritization. Matrix is composed of coarse sandstone. Outcrop is cut by calcite veins.

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Lower and Upper Boundaries
The lower boundary between the Marikapal Member with the Jojok Member is depositional (Plate 6.1a). The upper boundary with the Ruta Formation is not observed. An unconformable upper contact is inferred from the distribution of micropalaeontological ages.

Thickness
As with the Jojok Member, the thickness of this member is difficult to ascertain. The hills inland from the southern coast of Kasiruta, composed mostly of this member having ~20° dip, rise up to 350 m. Assuming that the area is not severely faulted and/or folded, as justified by the aerial photographic study, the thickness of the member is estimated to be at least 500 m.

Lateral Extent and Variations
At the type area for the Tawali Formation, the Marikapal Member is in depositional contact with the Jojok Member. Buff-tan-grey mudstone and sandstone (ss:ms ~80:20) forming medium beds blanket the pillows. Clasts are angular-subrounded and moderately sorted, consisting of plagioclase, pyroxene, opaques and basaltic and limestone lithic fragments. Grading, convolute bedding, parallel and cross lamination (mm scale) and load structures are present in some beds (Plate 6.7a). The proportion of sandstone decreases up section to about 11 m above the pillows, above which the sandstone percentage is roughly equal to the mudstone. Locally, coarser grey sandstone layers, with mm scale lamination, form lenses within the finer sandstone. A subangular, ~50 mm recrystallized limestone clast is present in the coarser sandstone which otherwise contains the same clasts as the fine sandstone. This coarse sandstone shows flame structures, rip-up clasts and convolute bedding (Plate 6.7b). The sequence is interpreted to be facies B2.

Along S. Leleojaya, there are sandstones overlain by mudstone and breccia units (possibly facies B, C and A). Clasts are angular, poorly-moderately sorted and include pyroxene and green amphibole, suggestive of a volcanic origin. The sandstone unit shows parallel bedding (~15 mm thick) with a composite thickness of ~1 m. The mudstone has a sharp contact with sandstone and is ~0.5 m thick with mm-cm scale lamination. The breccia unit is ~1.5 m thick and contains Jojok-type basalt, mudstone and limestone clasts of unknown origin.

This member is also exposed on the northern coast of Kasiruta and Muari. At these localities the percentage of mudstone is higher (ss:ms ~ 25:75). Clasts are subangular and moderately sorted. Mudstones are more calcareous than sandstones. Sandstone beds show normal grading, while
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mudstone shows load structure when in contact with sandstone (facies C).

Slump deposits on a metre scale can be seen on northern Kasiruta (west of Tawa Village) within the sandstone unit. More resistant units are disrupted and formed irregular, overturned, convoluted patterns. Slumping appears to occur within specific beds only, bounded by coherent beds (Plate 6.7c). The presence of non-rotational basal shear planes suggests a slumping mechanism (Stow, 1986). Current ripple marks are present in the finer grain sandstone at Tawa village.

6.6.2 Petrography

Petrographically, the majority of breccia clasts resemble the basalts of the Jojok Member. The lesser limestone clasts in the breccia are composed of coral fragments, pyroxene, lithic fragments, opaques and zeolites in a micritic matrix. Some limestone clasts are completely recrystallized.

The sandstone unit shows point and long contacts between grains. Sorting is poor-moderate and grains are angular-subrounded. The grains consist of basaltic lithic fragments (partly replaced by calcite, clays and zeolite, ~20%), plagioclase (broken albitised laths, showing complex twinning, ~15%), clinopyroxene (some twinned, ~10%), opaques (broken crystals, ~5%), minor light green hornblende, rare altered olivine, bioclastic limestone fragments and planktonic foraminifera. The matrix consists of green, brown or yellow clay (?smectite, ~40%). Secondary minerals present include calcite, prehnite and zeolite, all in the form of veins. Smectite occurs surrounding some grains (Plate 6.8a,b).

The siltstone unit is matrix-supported or has point contact between the moderately sorted, angular grains. Grains are opaque (~5%), plagioclase (~5%), pyroxene (~5%), occasional green hornblende, bioclasts (~5%) and volcanic lithic fragments (~5%), with similar characteristics to the sandstone unit. The matrix is composed of both clay minerals and micrite (~75%). Carbonate replaces some plagioclase, and volcanic clasts are altered to green, brown and yellow clays (Plate 6.8c,d).

The secondary minerals (zeolites, smectites) found in the clasts, the matrix and the veins indicate low temperature zeolite facies metamorphism similar to that of the Jojok Member.

6.6.3 Age Determination

Foraminiferal analyses from the type locality (S.J. Roberts & P.T. Robertson Utama, pers.comm., 1990) indicate that this member was deposited during the Late Oligocene (P22).
Plate 6.7a. Sandstone bed within the Marikapal Member across P. Dokoh. Note the presence of normal grading, convolute bedding, parallel and cross lamination (mm scale) and load structures.

Plate 6.7b. Coarse sandstone bed within the Marikapal Member across P. Dokoh. Note the presence of flame structures, rip-up clasts and convolute bedding.

Plate 6.7c. Slumped unit within the Marikapal Member, along the west coast of Kasiruta.
Plate 6.8.a,b. PPL and XPL of Marikapai Member sandstone (BR170).
Scale for all photomicrographs = 0.5 mm.

Plate 6.8.c,d. PPL and XPL of Marikapal Member coarse siltstone with planktonic foraminifera (BM541).
Micropalaeontological evidence suggests that there may have been a period of non-deposition between the Jojok and Marikapai Members (Fig. 6.13). The following planktonic foraminifera are the marker species: *Globoquadrina tripartita*, *Gq. binaensis*, *Gq. praedehiscens*, *Gq. venezuelana*, *Globigerina praebulloides*, *Gg. yeguaensis*, *Gg. tauriensis*, *Gg. praeapsis*, *Gg. ampliapertura* and *Gg. euapertura*.

### 6.6.4 Depositional Environment and Mechanism

From the sedimentary structures and the immature texture, the Marikapai Member is interpreted as a "proximal" turbidite with associated slump deposit. The narrow age distribution reflects a high sedimentation rate.

Pickering *et al.* (1989) suggested that facies A2.3 of the breccia unit represents a high concentration turbidity current with grain by grain deposition. Facies B1.1 of the sandstone represents high concentration turbidity currents with rapid deposition due to intergranular friction. Facies C2.2 reflects an intermediate concentration current with grain deposit from suspension followed by bed load traction. Facies D2.2 was deposited by a low concentration turbidity current or a weak bottom current with a rapid clasts deposit from suspension, followed by traction transport. The latter two facies contain the Bouma Ta-d sequence and suggest deposition by high density turbidity currents which lost energy as they travelled down slope. Repeated deposition is deduced from the repeated Bouma sequence, the presence of amalgamated beds and the thickness of the member.

Both the turbidites and slump deposit require the existence of a slope, and coralline limestone clasts suggest reefs were present near the upper part of the slope. This is consistent with the micropalaeontological evidence, where foraminifera reflect middle-outer neritic environments. The sediment source was a mixture of the Jojok Member and limestone. The poor quality of exposure and absence of palaeocurrent evidence prevent detailed palaeogeographic reconstruction.

### 6.7 STRUCTURE

For the most part, Jojok Member rocks are flat lying, except near Ruta Village, in northwestern Kasiruta, where the dips are ~10°, some up to 30°. The bedding direction of the Marikapal Member varies with the prominent dip directions to the NNW and SSE (Fig. 6.15a), indicating folding along an E-W axis and therefore a dominant N-S stress orientation.
Figure 6.15a. Rose diagram showing the dip directions of the Marikapal Member.

Figure 6.15b. Summary diagram of joint direction affecting the Tawali Formation.
The Tawali Formation

The most significant structure affecting the Tawali Formation is normal faulting, some of which appears to be listric in nature. The widespread minor deformation observed is attributable to this syndepositional extensional faulting causing rotation of small blocks.

It is noticeable that the Tawali Formation is affected by the same joint patterns as the Bacan Formation (Fig.6.15b), indicating post-Tawali Formation joint development. These joints are interpreted as related to more recent tectonic processes, due to its consistency with present-day fault and joint directions.

Aerial photographic study suggests the presence of large-scale, open folds in the central Kasiruta. The fold axes are running NE-SW, suggesting a NW-SE stress direction, consistent with the dip directions. Discontinuous exposure prevents a more detailed structural study.

6.8 SYNTHESIS AND TECTONIC SIGNIFICANCE
The Tawali Formation is the oldest formation observed on Kasiruta, with an upper unconformable contact with the Ruta Formation. The Marikapai Member is the younger of the two members.

The Jojok Member pillow basalt is a clearly a submarine lava, erupted in the Early Oligocene, in an deep, open marine environment, in an arc-related setting. The presence of calcareous planktonic foraminifera and nannofossils implies deposition above the CCD. Multiple flows are inferred from the presence of sedimentary units between pillow lavas and the distinct layering between the basaltic flows. The thick sequence and narrow range of ages is a function of a rapid deposition. Geochemical analyses shows that this member is composed of highly differentiated, volcanic arc rocks.

The Marikapai Member filled an arc-related basin and is interpreted as the product of repeated, high density, proximal turbidity currents with associated debris flows during the Late Oligocene. The thickness of the member and short range of ages testify to the high deposition rate. The source of the Marikapai Member is interpreted as dominantly from the Jojok Member.

Both the Jojok and the Marikapai Members are affected by zeolite facies, shallow level burial metamorphism. The present day exposure of this formation along the westernmost margin of the Halmahera Arc is attributed to the fore-arc collision, uplifting this formation subaerially and creating a gravity high belt (Hall et al., 1992).
CHAPTER SEVEN
THE RUTA AND AMASING FORMATIONS:
Middle Miocene Carbonate Platform

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7.1 INTRODUCTION
Although the Ruta and Amasing Formations are distinct lithologically and aerial photographically, in this chapter both these formations are discussed together because of their significance in terms of geological evolution. The Ruta Formation consists of a shallow marine limestone deposited on a carbonate platform, interrupted by influxes of volcaniclastic material, which is the Amasing Formation.

Lithological descriptions of both formations are provided. Microfacies study of the Ruta Limestone was conducted and the results compared to Flügel’s (1982) standard microfacies. The depositional environment and facies distribution of the Ruta Formation is reviewed. The petrography of the Amasing Formation is addressed, and the depositional environment is examined. The age and the relationship of the Ruta and Amasing Formations are discussed. Structural evidence is considered and finally section 7.7 interprets the depositional environment of these two formations and assesses their tectonic significance.

7.2 SYNONYMY
The Ruta Formation is defined by Yasin (1980) as massive, crystalline limestone with local sandy limestone horizons. Although his stratigraphic nomenclature is adopted here, field and aerial photographic studies suggest the distribution of the Ruta Formation is more widespread than that of Yasin’s study (Fig.7.1 and Fig.2.1). The differences mainly arise from the similarities between the Ruta Limestone, the Mandioli Member of the Kaputusan Formation and the Quaternary Limestone. In order to differentiate these formations, biostratigraphic dating of numerous samples has been performed.

Yasin (1980) defined the Amasing Formation as tuffaceous sandstone, interbedded with claystone and marl, with limestone intercalations, and suggested that the Ruta Formation interfingers with the Amasing Formation. The distribution of his Amasing Formation includes the Bacan, Amasing and Ruta Formations of this study.

A large area of the Amasing Formation occurs on P. Obit, which Yasin (1980) had previously mapped as the Obit Formation. The term Obit Formation is now abandoned to avoid confusion with the Marikapal Member of the Tawali Formation (see Chapter Six) and the Pacitak Member
of Kaputusan Formation (*see Chapter Eight*), both previously mapped as the Obit Formation by Yasin (1980).

### 7.3 AERIAL PHOTOGRAPHY

A pronounced karstic topography characterises the Ruta Formation on aerial photographs which distinguish it from the Mandioli and Quaternary Limestones. The Amasing Formation is characterised by rugged hills, with radial to dendritic drainage patterns. Although geomorphologically similar to the Kaputusan Formation, its lower hills with slightly rounder tops, particularly evident in the finer grained rocks, distinguish the Amasing Formation.

### 7.4 THE RUTA FORMATION

The Ruta Formation is dominated by well bedded bioclastic limestones with a subordinate volcaniclastic component.

#### 7.4.1 Lithofacies

*Type Locality*

Although the type locality for this formation is at the Ruta Village, on northeast Kasiruta (Yasin, 1980), the best exposure studied is along the coast to the north of the village. At this locality, the rock consists of well lithified, well bedded (dm-m scale), white-buff, fine grained wackestone-packstone (Plate 7.1a). Faunal constituents include fragments of large corals, red algae, benthonic and planktonic foraminifera, echinoderm, bryozoan, bivalve (shells) and gastropoda. The matrix is micritic, although locally it is recrystallized to pseudospar. Calcite veins, up to 10 cm thick, are locally observed (Plate 7.1b).

At the headland between Ruta and Leleojaya Villages, there is a thinly bedded bioturbated wackestone, containing burrowing structures ~5 cm long and ~2 cm wide, parallel to bedding. Limestones analysed from the type area fall into two microfacies (*see Section 7.4.2 Microfacies A & B*).

*Lower and Upper Boundaries*

No contact with other formations was observed at the type locality. In S. Amasing, the Ruta Formation rests with an angular unconformity on the Bacan Formation. An unconformable contact is indicated by the presence of reworked Bacan Formation detritus in the Ruta Limestone at other localities. For example the conglomerates found at S. Nyonyifi contain clasts of the Bacan Formation and quartz-monzodiorite in a micritic matrix; the bioclastic limestones at S. Goro-goro
The Ruta and Amasing Formations

Plate 7.1a. Well bedded Ruta Formation outcrop, north of Ruta Village. Rock drill in the foreground was used for palaeomagnetic sampling.

Plate 7.1b. Calcite veining in the Ruta Formation, north of Ruta Village.

Plate 7.1c. Outcrop of Ruta Formation Microfacies C (Foraminiferal Packstone) at Tg. Supai, south of Ruta Village. Note cross lamination near Dr. Ali's knees.
The Ruta and Amasing Formations contain clasts of the Bacan Formation. An unconformable basal contact with the Tawali Formation in S. Kasiruta is deduced from the change in bedding attitudes and biostratigraphic information. An unconformable lower contact with the South Bacan Formation is inferred from biostratigraphy and the presence of subangular-subrounded clasts of the South Bacan Formation in Ruta Limestone float at S. Wayatim and S. Wayamoa.

The upper contact, in the S. Amasing, is a conformable transition to the Amasing Formation. Although the direct relationship with younger formations was not observed, micropalaeontological data suggests a hiatus between the Ruta Formation and the overlying Kaputusan Formation.

**Thickness**

The thickness of this formation is difficult to assess because of structural complexities and poor exposure. The maximum thickness exposed at Bukit Gabijapan in southern Kasiruta is ~30 m. The hill itself stands at 252 m high and is assumed, from aerial photographic study, to be composed solely of the Ruta Formation. The thickness of this formation is, therefore, estimated to be at least 250 m.

**Lateral Extent and Variations**

The Ruta Formation crops out at eastern Kasiruta, northern coast of Kaputusan, northern tip of western Bacan, southern Bacan and in patches in northern Bacan (Fig.7.1). There are also large angular boulders of the Ruta Formation, up to 6 m across, at S. Nyonyifi and S. Goro-goro, resting on the Bacan Formation. Similar boulders are found at S. Wayamoa, resting on the South Bacan Formation. For the most part these packstones are similar in character to those at the type locality, with a notable exception of the presence of impure packstone. Rock fragments of the Nusa Babi quartz-monzodiorite, Bacan and South Bacan Formations are observed, especially near basal contacts. Detrital clasts are subangular-subrounded, up to 1 cm across, occasionally forming floatstone and rudstone. Parallel lamination on mm-cm scale is present in the more clastic-rich packstone. At S. Goro-goro and western coast of northern Bacan, the Ruta Limestone also contains plant debris (see Section 7.4.2 microfacies D).

At Tg. Supai, on the eastern coast of Kasiruta, there is a well bedded foraminiferal packstone. The bioclasts are mostly planktonic foraminifera with subordinate volcanic lithics, all of which are well rounded and well sorted. Parallel and cross laminations on a cm-dm scale are present (Plate 7.1c). S.J. Roberts & J.Ali (pers.comm., 1990) reported a dark, massive, foraminiferal packstone forming a steep sided gorge with 30 m cliffs at S. Singga (see Section 7.4.2 microfacies C).
Figure 7.1. The distribution of the Ruta and Amasing Formations in the Bacan Region.
The Ruta and Amasing Formations

7.4.2 Microfacies

Microfacies study has enabled the grouping of limestones into major types which reflect their depositional environments. In areas where sedimentary structures are lacking, such as in the Bacan region, microfacies study provides the only tool of facies recognition. Originally devised by Wilson (1975), this method has been refined by Flügel (1982), whose standard microfacies are used throughout this study. In the Bacan region, five microfacies have been distinguished, and in this section they are discussed in decreasing order of abundance. Modal percentages are based on visual estimation. Identification of fauna is used to determine palaeosalinity and water depth, while textural information is used for facies and palaeoenergy interpretation. All of the data obtained are from disparate outcrops and boulders, with no clear relationships between microfacies. Interpretation for the different microfacies are therefore dealt with separately.

**Microfacies A**

This microfacies is a skeletal wackestone-packstone (Dunham, 1962). It contains a diverse bioclastic assemblage, dominated by large coral, red algae and benthonic foraminifera fragments. Red algae (coralline algae) occur in two forms: rounded allochems with honey comb structures and elongated allochems with delicate branching structures. Other bioclastic constituents are all fragmented and include echinoid, planktonic foraminifera, bivalves, bryozoans and rare gastropoda. The larger bioclasts are abraded, whereas the smaller ones are better preserved. Rare subrounded micritic intraclasts and volcanic lithic fragments are observed and there are occasional angular quartz (monocrystalline and polycrystalline) and plagioclase clasts. A micritic matrix forms ~35% of the rock, although locally it is recrystallized to pseudospar. This rock shows textural inversion (Plate 7.2a,b).

All allochems are poorly sorted, disarticulated, fragmented; and are mostly angular-subangular. There is no evidence for grading or preferential alignment of bioclasts.

Brown organic-rich patches occur as low amplitude stylolites. Severely deformed benthonic foraminifera suggest a high degree of compaction. This microfacies is similar to standard microfacies 10.

**Interpretation**

The abundance of coral and coralline algae, coexisting with benthonic and planktonic foraminifera, brachiopoda and echinoderms implies normal salinity, shallow marine (10-100 m) environments (Flügel, 1982). Abraded clasts observed in a micritic matrix (textural inversion) represent original
deposition in a high energy environment, subsequent transportation and final deposition in a low energy environment. A mixture of high and low energy in a shallow marine environment can be achieved on an open shelf or open platform. The presence of angular volcaniclastic detritus indicates a near-shore depositional basin which argues for the latter interpretation. This is further supported by the lack of interbedded pelagic rocks.

**Microfacies B**

This microfacies is an algal boundstone (Dunham, 1962). It is dominated by abraded, organically bound coralline algae with minor brachiopoda, echinoids, benthonic foraminifera and bivalves. Coralline algae again occur in two forms: rounded allochems with honey comb structures and elongated allochems with delicate branching structures developing an organic framework. Minor volcaniclastic lithic fragments are locally observed. A micritic matrix constitutes ~15% of the rock, and as microfacies A, it is occasionally recrystallized to pseudospar (Plate 7.2c,d).

Allochems are poorly sorted. Except the coralline algae, other bioclasts are disarticulated, fragmented and are angular-subangular. Volcaniclastic fragments are subrounded, and are of coarse sand-gravel grade size. This microfacies is similar to standard microfacies 7.

**Interpretation**

The domination of coralline algae and the presence of benthonic foraminifera, brachiopoda and echinoids indicate a normal salinity, shallow marine (10-100 m depth) palaeoenvironment (Flügel, 1982). The presence of organically bound coralline algae indicates *in situ* growth of reef, whereas the other allochems are transported. This type of organic building may represent a minor organic framework on the platform margin. The well preserved structures of the coral and the micritic matrix may be due to a low energy environment or infiltration of carbonate mud on the reef.

**Microfacies C**

This microfacies is a well sorted foraminiferal packstone (Dunham, 1962). The rock is composed of ~70% benthonic and planktonic foraminifera, in a micritic matrix. Minor constituents include echinoid fragments, red algae, corals, brachiopoda, bivalves, ?glaucony and volcanic lithic fragments. Rocks are often packed with allochems in a micritic matrix (~10%; Plate 7.3a).

All coarse sand size allochems are subrounded-well rounded with point contacts between them. This is similar to standard microfacies 12 or 18.
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Interpretation
The coexistence of planktonic and benthonic foraminifera, coralline algae and echinoids indicates 10-100 m water depth, with a normal marine salinity environment (Flügel, 1982). This foraminiferal packstone is interpreted to be equivalent to either a coquina packstone with abundant foraminifera (SMF 12) or a foraminifera grainstone (SMF 18). SMF 12 is commonly found on open slopes and shelf edges (winnowed platform). SMF 18 is characteristic of deposition in tidal bars and channel lagoons, developed on an open or restricted platform. The presence of cross lamination (see Section 7.4.1) favours the tidal bar interpretation. This is consistent with the high sphericity of allochems suggesting a high degree of reworking. The abundance of planktonic foraminifera indicates an open platform environment.

Microfacies D
This is a bioclastic-lithoclastic packstone (Dunham, 1962). Allochems are corals, red algae, benthonic foraminifera, micritic intraclasts and lithic fragments. Bioclasts are fragmented. Recognisable lithic fragments are of the Bacan and South Bacan Formations and Nusa Babi quartz-monzodiorite. Micritic matrix constitute less than 10% of the rock (Plate 7.3b,c).

Allochems are poorly sorted. The bioclasts are angular, while lithic fragments are subrounded-rounded. The size of allochems varies from ~0.5 cm - fine sand size. This is comparable to standard microfacies 4.

Interpretation
The angularity of the bioclastic fragments suggests a low level of reworking, whereas the sphericity of the volcaniclastic component implies the reverse. A fore slope talus environment can account for both levels of reworking; with previously reworked volcaniclastics and unreeved bioclastics derived from the erosion of the fore slope. The angularity of bioclasts and near absence of matrix demonstrates a high energy environment. The presence of corals, red algae and benthonic foraminifera indicates 10-100 m water depth with normal salinity (Flügel, 1982).

Recrystallized Limestone
Crystalline limestones are found, interbedded with all microfacies. These have calcite dominated spars, although locally, rhombic dolomite occurs. Spars, in general, exhibit irregular boundaries except in veins where the boundaries are more regular. Veins also have larger crystals (Plate 7.3d). The presence of areas of relict bioclasts indicates that the spars are neospars.
The Ruta and Amasing Formations

Plate 7.2a-b. PPL and XPL of Ruta Formation Microfacies A (BM529). Note the presence of red algae and large benthonic foraminifera in a micritic matrix.

Plate 7.2c-d. PPL and XPL of organically bound coral head (Ruta Formation Microfacies B; BT5).

Scale for all photomicrographs = 0.5 mm.
Plate 7.3a. PPL of Ruta Formation Microfacies C (BR199), dominated by rounded planktonic foraminifera.

Plate 7.3b. PPL of Ruta Formation Microfacies D (BM212). Note similarity with Microfacies A, except presence of volcaniclastic fragments and micritic intraclasts.

Plate 7.3c. XPL of Ruta Formation Microfacies D (BM212).

Plate 7.3d. PPL of recrystallized limestone (BR235). Note presence of relict benthonic foraminifera and algae.

Scale for all photomicrographs = 0.5 mm.
Interpretation

Interpretation of the depositional environment of the recrystallized limestone is not possible, due to lack of original fabric. However, neospars are still important indicators of the presence of a high thermal gradient attributed either to burial or hydrothermal fluid. The occurrence of both recrystallized and well preserved limestone at most localities favours the latter cause.

7.4.3 Age Determination

Micropalaeontological determination indicates an Early-Middle Miocene age, with shallow-middle neritic (shelf) depositional environment. Fig.7.2 shows the distribution of the ages and the location of samples. The following index fossils were used to determine the age of the samples; planktonic foraminifera: *Globorotalia (F.) foshi* Group, *Gr. (G.) panda*, *Globoquadrina venezuelana*, *Dentoglobigerina altispira*, *Orbulina universa*, *O. Surtralis*, *Globigerinoides subquadratus*, *Sphaeroidinellopsis seminulina seminulina*; large benthonic foraminifera: *Cycloclypeus* sp., *Nephrolepidina* sp., *Miogypsinoides dehaartii*, *Myogypsina borneensis*, *Lepidocyclina* sp., *Amphistegina* spp., *Spiroclypeus* spp.; nannoflora: *Helicosphaera kamptneri*, *Cyclocargolithus* sp., *Pontosphaera* sp..

The deposition of Ruta Limestone Formation began in Late Aquitanian times, when it was dominated by the open platform, fore slope talus and platform margin facies. During Langhian times, deposition was interrupted by influxes of volcaniclastic material, forming the Amasing Formation. Although none of the samples dated give the same age as the Amasing Formation, field relationships demonstrate that the two formations are interfingering, consistent with the suggestion of Yasin (1980). Samples collected from S. Amasing, where this relationship is clearest are unfortunately recrystallized and cannot be dated. Deposition of the Ruta Formation continued until Serravallian times, and was dominated by the tidal bar facies. There is a hiatus between the Ruta Formation and the overlying Kaputusan Formation.

7.4.4 Depositional Environment and Facies Distribution

The presence of burrows, coral, algae, shells and plant fragments and micropalaeontological environmental analysis, indicate that this formation represents a shallow water deposit. This is further supported by the fact that the limestones are well bedded, which is suggestive of shelfal origin.
Figure 7.2. Biostratigraphic dates for the Ruta and Amasing Formations.
Figure 7.3. Facies distribution of the Ruta Formation in the Bacan Region.
The Ruta and Amasing Formations

Microfacies analysis demonstrates that all limestones were deposited in a shallow marine (10-100 m), normal salinity environment. Facies A represents deposition on an open platform. Facies B was part of a platform margin build up, possibly representing growth of patch reefs, indicated by the limited extent of this facies. Facies C is interpreted to be the deposit of tidal bars on the down slope margin of a patch reef (open platform). Facies D is interpreted to be a fore slope (talus) deposit. The distribution of the facies and a cartoon diagram of the different facies applicable to the Ruta Formation is shown in Fig.7.3.

7.5 THE AMASING FORMATION

The Amasing Formation consists of interbedded volcaniclastic conglomerate, sandstone, siltstone and mudstone. There are three facies recognised in the Amasing Formation: interbedded fossiliferous sandstone and mudstone (Facies A); coarse sandstone with large scale cross bedding (Facies B); and well sorted conglomerate (Facies C).

7.5.1 Lithofacies

Type Locality

Originally named by Yasin (1980), this formation is named after S. Amasing where interbedded conglomerates, sandstones, siltstones and mudstones are present. The type locality is divided lithologically into facies A and B.

**Facies A**

The lowest observable unit is a friable, thinly bedded greenish grey conglomerate, sandstone, siltstone and mudstone. The conglomerate consists of mudstone clasts supported in a coarse sandstone matrix. Sandstone, siltstone and mudstone units are texturally immature, forming parallel laminations (~1.5 cm thick). Sandstone beds are typically 25 cm thick, whereas the siltstone and mudstone beds are ~40 cm thick. Minor cross lamination occurs within the sandstone beds, which are locally graded. Convolute laminations, load and flame structures and rip-up clasts are present in the finer grained beds and in layers containing disarticulated bioclasts (bivalves and gastropods) and plant remains (Plate 7.4a). Some bioturbation structures is observed in mudstone floats. The sandstone beds may contain calcite veins, perpendicular to bedding. The siltstone and mudstone units are blackish in colour and occasionally contain pyrite. Fig.7.4 is a schematic sedimentary log of this facies.

**Facies B**

Above facies A is a moderately lithified, texturally immature tuffaceous sandstone with carbonate
The Ruta and Amasing Formations


Graded sandstone, contains parallel lamination ~ 0.5 cm thick. Some shell fragments are present. Mudstone with occasional discontinuous sandstone beds ~ 10 cm thick. Some are covered by recent material.

Graded sandstone with minor cross lamination and broken shell fragments. Black claystone with convolute lamination, load and flame structures.

NO EXPOSURE

Sandstone with ~ 3 cm parallel lamination. Mudstone with ~ 1 cm parallel lamination. Sandstone with ~ 5cm lamination with mudstone rip up clasts. Mostly eroded away. Mudstone interbedded with sandstone.

Mudstone: Black, contains some pyrite and is carbonaceous. Sandstone: Parallel lamination and some minor cross lamination, all < 1 cm thick. Some calcite veining.
The Ruta and Amasing Formations

cement showing a grey-blue colour when fresh and brown tan when weathered. The clasts are coarse sand size, angular and moderately sorted. This unit forms 2-3 m parallel bedding with internal parallel stratification thickness of 3-10 cm and low angle trough cross lamination (~5 cm thick), forming sets ~1.5 m high, ~3 m long (Plate 7.4b). Current ripple marks are present in some bedding surfaces, and plant debris are found in some layers. Measured current directions are towards 310 and 320. The sequence is right way up.

Stratigraphically above it is a fine grained sandstone, exposed along S. Lemo-lemo, a tributary of S. Amasing. Although lithologically similar with the coarser sandstone, this unit is thinner bedded (mm-cm scale) with thin lenses of coal, up to 1 cm thick, occurring locally. Brouwer (1923) reported the presence of coal lumps in S. Amasing and Yasin (1980) found 10-20 cm cobbles of coal at S. Lemo-lemo. Locally this facies contains green, fine grained, well bedded, well lithified calcareous sandstones (~1 m thick) with convolute lamination and calcareous mudstone with large clasts of siltstone, interpreted to be a slumped unit.

Lower and Upper Boundaries

In S. Amasing and S. Singga, there is a depositional lower contact with the Ruta Formation. The relationship with the Ruta Formation is interpreted as an interfingering one, based on the nature of contact, consistency in bedding attitudes, the ages of the two formations (Fig.7.2) and their palaeoenvironmental interpretation.

The upper boundary of this formation is not seen. However, micropalaeontological and isotopic dating evidence suggest the presence of a hiatus between the Amasing/Ruta and the Kaputusan Formations.

Thickness

The minimum thickness of this formation is 180 m, deduced from the height of a hill climbed from S. Lemo-lemo, composed entirely of flat lying coarse grained tuffaceous sandstone.

Lateral Extent and Variations

Another place where the Amasing Formation is exposed is around the northern coast of Labuha, at Bacan, and at Obit. The exposure along this coast is dominated by Facies A, which is in fault contact with the Pacitak Member of the Kaputusan Formation. The similarity between the two units makes it difficult to distinguish them except micropalaeontologically. Fig.7.1 shows the distribution of the Amasing Formation in the Bacan region.
Plate 7.4a. The Amasing Formation Facies A, an interbedded sandstone (more resistant units) and mudstone, at S. Amasing.

Plate 7.4b. The Amasing Formation Facies B, a reworked tuffaceous sandstone deposit. Note large scale, low angle trough cross lamination. Exposed at S. Amasing.
The Ruta and Amasing Formations

At the NE end of Obit, well laminated, parallel bedded volcaniclastic conglomerate and sandstone crop out. The conglomerate is well sorted with rounded clasts of hornblende andesites in a mudstone matrix. Discrete units of green, shelly sandstones and siltstones are bioturbated, resembling coquina layers. Beds are up to 70 cm thick, though typically thinner (~ 30cm). Although spatially limited, this unit is different in being dominated by conglomerate, and is distinguished as Facies C.

At S. Singga, the Ruta Formation is conformably overlain by an ~10 m thick volcaniclastic conglomerate. Conformably above the conglomerate is a shelly calcarenite containing up to 10 cm thick parallel lamination, dewatering and loading structures, vertical and horizontal burrows (up to 2 cm across) and mudstone rip-up clasts (S.J.Roberts, pers.comm, 1990).

Brouwer (1923) reported the presence of a brownish grey quartz-sandstone, near S. Kubung at the northwestern foothill of Bukit Sibela, which is interpreted to be part of the Amasing Formation (?Facies B). This outcrop, however no longer exists (Hall, pers.comm, 1985).

7.5.2 Petrography

Facies A

Petrographic analysis of sandstone (Plate 7.5a) reveals that grains are in point and long contacts, with moderate porosity. The grains are angular-subangular lithoclasts and subrounded bioclasts (7%), notably planktonic foraminifera and corals. A micritic matrix (35%) contains plagioclase (20%), lithics (20%), with subordinate amounts of opaques (5%), pyroxenes (3%), amphiboles (10%) and biotite (<1%). Plagioclase grains are broken laths, showing complex twinning and zoning. Lithic fragments are chloritised volcanics probably derived from the Bacan Formation.

Facies B

The sandstone grains in this facies are in point and long contacts, with moderate porosity. They are angular-subrounded, with a size mode of 0.3 mm. Grains are plagioclase (30%), lithic fragments (25%), opaques (10%), pyroxene (15%), green hornblende (1%), quartz and epidote (1%), and bioclasts (5%; planktonic foraminifera and coral fragments). Plagioclase grains consist of broken laths, showing complex zoning, twinning, intergrowth and contain numerous glass inclusions. Lithic fragments are dominated by altered volcanic glass (replaced by chlorite and smectite). Opaques are finer and euhedral. Broken euhedral orthopyroxene and clinopyroxene may show simple twinning and slight green pleochroism. Quartz and epidote are detrital secondary minerals, after plagioclase and pyroxene respectively. Locally micritic limestone intraclasts are also
The Ruta and Aniasing Formations

Plate 7.5a. XPL of the Amasing Formation sandstone Facies A (BR228). Note the presence of planktonic foraminifera.

Plate 7.5b. XPL of the Amasing Formation sandstone Facies B (BM58).

Plate 7.5c. XPL of the Amasing Formation sandstone Facies B, slump sequence (BR225).

Plate 7.5d. XPL of the conglomerate clast (porphyritic andesite) from the Amasing Formation Facies C (BR216).

Scale for all photomicrographs = 0.5 mm.
The Ruta and Amasing Formations

present. The matrix is composed of micrite and brown smectite (Plate 7.5b).

This facies sometimes contains slumped sequences, e.g. at S. Amasing. These are finer grained, more calcareous, better lithified and more compact. The grains are angular, with poor-moderate sorting, and are in point and long contacts. They consist of plagioclase (10%), lithic fragments (5%), limestone intraclasts (10%), opaque (5%), minor pyroxene, epidote and green amphibole in a micritic matrix. Plagioclase shows complex twinning and zoning. Lithic fragments are all phytic volcanics, with plagioclase laths, amphibole, pyroxene (locally replaced by epidote) phenocrysts. Intraclasts consist of micrite with occasional planktonic foraminifera (Plate 7.5c).

**Facies C**

Conglomerate clasts in this facies are all andesite, composed of plagioclase, pyroxene, amphibole and opaque phenocrysts in a glassy matrix, showing porphyritic texture (Plate 7.5d). Plagioclase laths show simple twinning. Pyroxene are euhedral-subhedral, with occasional simple twinning. Strongly pleochroic euhedral amphiboles are red brown-orange in colour; some are replaced by opacite. Opaques are euhedral, with a squared outline. Smectite locally replaces pyroxene.

The matrix of the conglomerate is calcareous mudstone. Interbedded sandstone and mudstone units in this facies have the same petrographical characteristics as sandstone and siltstone of Facies A.

**7.5.3 Age Determination**

This formation has been dated using foraminifera and nannofossils. Poor foraminifera assemblages have been extracted and indicate an age range of Early-late Middle Miocene (N5-N15), with a near shore marine palaeoenvironment. Reworking of Cretaceous, Eocene and Oligocene faunas is common. Good nannofloral assemblages indicate a latest Early Miocene (NN4-NN5) age. Fig.7.2 shows the distribution of ages in the Amasing Formation. Analysis includes samples from facies A, B and C (matrix). Characteristic faunas include: planktonic foraminifera: *Globoquadrina baroemoenensis, Dentoglobigerina altispira altispira, Globigerinids, Globigerinoides immaturus, Gs.triloba* and *Neogloboquadrina continuosa*; nannoflora: *Cyclocargolithus abisectus, Cyclocargolithus floridanus, Discoaster exilis, Discoaster variabilis, Sphenolithus heteromorphus, Sphenolithus belemnos, Sphenolithus moriformis, Coccolithus miopelagicus* and *Coccolithus pelagicus*.
7.5.4 Depositional Environment

The presence of layers of disarticulated bioclasts (bivalves and gastropods) and plant remains resembling coquinas, in Facies A, are interpreted to be storm related horizons \( \text{cf.} \) Johnson & Baldwin, 1986 and references therein). This argues strongly for a subnormal wave base marine environment, indicating a near shore, shallow marine environment. Sedimentary structures, i.e. interbedded sandstone, siltstone and mudstone, parallel and cross lamination, grading and bioturbation structure are all consistent with this interpretation (Johnson & Baldwin, 1986). Textural immaturity of the sediments suggests a low level of reworking, supportive of the shallow marine origin. Finally, nannofossil analysis also indicates a near shore, shallow water environment. Convolute laminations, load and flame structures and rip-up clasts indicate significant variation in the thixotropy of the sediments at the time of deposition. The presence of black mudstone with associated pyrite, suggests that the environment within the sediment is reducing, either due to restricted environment or rapid sedimentation. The high biota content, the thickness of the facies and the restricted age range favours the latter cause. This is consistent with sudden high influxes of volcanic material, indicated by the amount and immaturity of volcaniclastic detritus in this facies.

The presence of coal, plant-rich layers and nannofossil analysis in Facies B, indicate a near shore, shallow water environment. The sedimentary structures, i.e. parallel bedding, large scale cross lamination and ripple marks, suggest a current dominated environment; an estuarine or shallow marine bar (shoal). The immaturity of the volcaniclastic detritus demonstrates a low level of reworking, in contrast to either interpretation. This is attributed to the sudden and abundant volcaniclastic and tuffaceous input. The restricted occurrence of this member represents local development of the estuarine or bar. Slumped units within this member suggests the presence of a slope. The presence of limestone intraclasts containing planktonic foraminifera and coral fragments is supportive of shallow marine sediment source.

Lack of exposure hampers palaeoenvironmental interpretation of Facies C, which is tentatively interpreted as a beach deposit. This would explain the maturity of the conglomerate clasts and the mudstone matrix. The presence of coquina layers is suggestive of storm affected horizons, indicating a shallow marine environment. Further interpretation is not likely to be possible due to lack of exposure.

Micropalaeontological evidence and field relationships suggest an interfingering relationship with the laterally continuous carbonate platform of the Ruta Formation. The sequence at S. Amasing,
The Ruta and Amasing Formations

therefore, represents a coarsening upward, shallowing-upward deposit, from a shallow water carbonate shelf (Ruta Formation) to a subnormal wave base, storm dominated, shallow marine sequence (Amasing Formation, Facies A) to a current dominated, shallow marine estuarine or bar, deposited above the storm wave and below fair weather base (Amasing Formation, Facies B). Facies C (beach deposit) may be located above Facies B, representing a continuous shallowing-up sequence.

The source of the Amasing Formation is interpreted to be either localised volcanic activity or uplifted older volcanic formations (e.g. South Bacan, Tawali or Bacan Formation).

7.6 STRUCTURE

Aerial photographic study of northern Kasiruta, shows that the Ruta Formation has been folded on a km scale. The folds are gentle, open folds with the axes trending northeast - southwest. The bedding attitudes of the Ruta Formation are variable, with most dips under 20°. At the type locality it dips towards the southwest and east, while at southern Kasiruta it dips towards the north. In Bacan the bedding attitudes of the Ruta Formation are also variable; dipping towards the east (30°-60°) in western Bacan, and towards the southwest (~12°) in eastern Bacan.

Similarly, the Amasing Formation is affected by large scale, open folds, readily observed in Facies A sequence. The fold axis is 070, similar to the Ruta Formation's in Kasiruta. Facies A is also affected by normal faulting with a 230/40 fault plane attitude. Both folding and faulting are consistent with a northwest-southeast compressional stress. In contrast, Facies B is nearly flat lying and appears to be hardly affected by folding. This is attributed to the better lithified, coarser grain nature of this facies. The bedding attitudes of the Amasing Formation are similar to the Ruta Formation. Similarity in deformational style is further evidence of the interfingering relationship between the Ruta and Amasing Formations.

7.7 SYNTHESIS AND TECTONIC SIGNIFICANCE

After the Early Miocene uplift and erosion period (N4-N5 or NN1-NN2 inclusive), the Middle Miocene Ruta Formation was unconformably deposited on the Bacan Formation in northern Bacan, the South Bacan Formation in southern Bacan and on the Tawali Formation in Kasiruta. This formation is deposited above the regional unconformity and represents the first common formation deposited above rocks of Australian and Philippine Sea Plate affinities. The oldest date obtained from this formation (~19 Ma) therefore signifies the latest possible time of collision between the Australian and Philippine Sea Plates (see Chapters Four, Five and Six) in the region. The
The Ruta and Amasing Formations

consistency in deformatonal style, above different formations, is supportive of this interpretation.

There are four microfacies recognised in the Ruta Limestone: wackestone-packstone, algal boundstone, foraminifera packstone and bioclastic-lithoclastic packstone, representing open platform, platform build up (platform margin or patch reef), tidal bar facies (down slope of patch reef) and fore slope talus deposit, respectively. These form a shallow marine carbonate platform. The relative uniformity of facies over such a widespread geographical area can be partly attributed to resistance of facies to weathering and selective sampling due to inaccessibility of outcrops. Alternatively the Ruta Formation may represent a laterally extensive shallow marine carbonate platform.

Carbonate rocks of similar age and lithology to the Ruta Formation can be recognised throughout Eastern Indonesia, such as the Subaim Limestone of Halmahera (Hall et al., 1988b), the Fluk Limestone of Obi (Hall et al., 1991), the Waigeo Formation of Waigeo (Charlton et al., 1991) and the limestones of Misool (Froidevaux, 1974) and Irian Jaya (e.g. Pigram & Davies, 1987), although it is not established whether these form part of the same platform.

Pigram et al. (1990) suggested that the Middle Tertiary carbonate platform of Papua New Guinea developed as a result of rapid subsidence of the foreland basin and northward movement of the plate which carried the basin into temperate climatic regime. In the Bacan region, development of a thick, laterally extensive carbonate platform is envisaged to be related to the strike slip regime, where carbonate deposition kept up with subsidence. This supported by the palaeomagnetic results of Hall et al. (1993) which shows the Halmahera region, including Bacan, at the southern margin of the Philippine Sea Plate (~5°S latitude) moving along a proto Sorong Fault system.

The Amasing Formation is a shallow marine shoal deposited at the latest Early Miocene and represents localised, high influxes of volcanioclastic material onto the platform. Three facies which comprise the Amasing Formation have been recognised: shallow marine, shoal and beach deposits. Although clasts of the Bacan Formation are recognised in the Amasing Formation, the majority of the clasts are more andesitic and less metamorphosed in character, suggesting a younger formation as a source.
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8.1 INTRODUCTION
The Kaputusan Formation is divided into three members: the Goro-goro Volcanic and Pyroclastic Member, the Pacitak Volcaniclastic Member and the Mandioli Limestone Member. The Goro-goro Member consists of andesite, minor basalt and dacite, the Pacitak Member includes volcaniclastic conglomerate, sandstone, siltstone and mudstone, the Mandioli Member consists of limestone with occasional volcaniclastic input.

In this chapter, the lithologies and petrography of all members of the Kaputusan Formation are discussed. The mineral and whole rock chemistry of the Goro-goro Member is used to constrain the tectonic setting, to ascertain petrogenetic trends, and, in conjunction with isotopic age, to investigate temporal variations. The age and depositional environment of the Pacitak Member is discussed, microfacies analyses of the Mandioli Member are evaluated, and its depositional environment considered. Structural evidence is reviewed and finally a synthesis and a proposed tectonic setting for this formation is presented.

8.2 SYNONYMY
The Kaputusan Formation includes the Bacan, Kayasa, Obit and Ruta Formations as well as the Quaternary Limestone of Yasin (1980). The Goro-goro Member roughly corresponds to the younger volcanic unit of Silitonga et al. (1981).

8.3 AERIAL PHOTOGRAPHY
This formation is characterised on aerial photographs by rugged, high hills, forming ridges, with a radial-dendritic drainage pattern.

8.4 TYPE AREA
The type area for the Kaputusan Formation is the area around Kaputusan Village on the west coast of northern Bacan. The Goro-goro Member is exposed inland from Kaputusan Village, along S. Kaputusan, where grey HBAN1 crops out and locally forms breccio-conglomerate units. These andesites have a microcrystalline or glassy matrix with hornblende and plagioclase phenocrysts up to 20 mm in length. Hornblendes locally forms glomerocrysts, up to 0.2 m across. Parallel alignment of phenocrysts is present locally. The andesites otherwise lack any internal structures.

Exposure of the Pacitak Member can be seen around the coast from the Kaputusan Village. There
The Kaputusan Formation

is an interbedded conglomerate, sandstone and siltstone-mudstone of tuffaceous origin in the hill behind the village. The monomictic conglomerate is matrix supported with subrounded-rounded HBAN clasts, up to 120 mm across. It is poorly sorted and massive although normal grading is observed in some beds. The sandstone has a fine to coarse grain-size and contain poorly sorted angular-rounded HBAN clasts and plant fragments. Reverse grading is present in one bed, otherwise this unit shows very thin parallel bedding. A better lithified grey sandstone unit contains more carbonaceous debris, orientated parallel to bedding. The siltstone-mudstone units have mm-scale parallel and trough cross-laminations and show convolute lamination when they are overlain by the conglomerate. Flame structures, slumping and other dewatering structures are common and storm beds are also identified in the sequence. Fig.8.1 is a simplified log of the exposure, showing a fining upward tendency.

The Mandioli Member is exposed in Tg. Poan area, north of Kaputusan Village, where there is a grey, moderately dipping, laminated limestone containing large benthonic foraminifera, coral fragments and algae draped by the poorly sorted breccio-conglomerate unit of Goro-goro Member, with clasts up to 0.5 m across. A detailed description of the member is provided below.

8.5 THE GORO-GORO VOLCANIC AND PYROCLASTIC MEMBER

These widely exposed andesites are grey when fresh, but reddish brown when weathered. Phenocrysts are euhedral and composed of plagioclase laths, amphibole, pyroxene, biotite and spinel, forming up to 30% of the rocks. They are typically less than 10 mm across. The matrix is microcrystalline, cryptocrystalline or glassy. Petrographically they can be grouped into four types: Two Pyroxene Andesite (TPAN), Hornblende Pyroxene Andesite (HPAN), Hornblende Andesite (HBAN) (nomenclature follows Hakim, 1989) and Hornblende Biotite ± Pyroxene Andesite (HBIAN). Due to lack of characteristic structures, field recognition of the different units is often difficult.

8.5.1 Lithofacies

Type Locality

Along S. Goro-goro, there is extensive exposure of poorly sorted, massive, tuffaceous sandstone with very angular TPAN clasts, up to 0.5 m in diameter, forming hills up to 60 m high (Plate 8.1a). At the base of these hills, there are boulders of poorly sorted, clast-supported volcanic breccia (Plate 8.1b) and sandstone, showing low-angle cross laminations (Plate 8.1c).
The Kaputusan Formation

**Fig. 8.1 Simplified Log of the Kaputusan Formation, Pacitak Member, Kaputusan Village.**

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>15 M</td>
<td>Breccia, subangular andesite clasts (up to 5 cm across), poorly sorted. Grey mudstone, contains plant fragments.</td>
</tr>
<tr>
<td>14 M</td>
<td>More resistant grey mudstone unit, contains plant fragments. Plant fragments appear to be perpendicular to bedding.</td>
</tr>
<tr>
<td>13 M</td>
<td>Massive, poorly sorted tuffaceous sandstone. Angular andesitic clasts, contains plant fragments.</td>
</tr>
<tr>
<td>12 M</td>
<td>Convolute bedding.</td>
</tr>
<tr>
<td>11 M</td>
<td>Massive, poorly sorted tuffaceous sandstone. Angular andesitic clasts, contains plant fragments.</td>
</tr>
<tr>
<td>10 M</td>
<td>Mudstone with mm scale lamination.</td>
</tr>
<tr>
<td>9 M</td>
<td>No Exposure.</td>
</tr>
<tr>
<td>8 M</td>
<td>Tuffaceous sandstone, graded. Clasts are hornblende andesite up to 6 cm across, some shows glomerocryst texture, poorly sorted. Contains plant fragments.</td>
</tr>
<tr>
<td>7 M</td>
<td>Poorly sorted breccia unit, massive. Clasts are hornblende andesite, up to 6 cm across.</td>
</tr>
<tr>
<td>6 M</td>
<td>Reverse grading, clasts are subrounded.</td>
</tr>
<tr>
<td>5 M</td>
<td>Parallel laminated tuffaceous sandstone. ~ 1 cm in thickness.</td>
</tr>
<tr>
<td>4 M</td>
<td>Poorly sorted subrounded hornblende andesite conglomerate, up to 12 cm across, no grading observed.</td>
</tr>
<tr>
<td>3 M</td>
<td>Poorly sorted rounded conglomerate, graded.</td>
</tr>
<tr>
<td>2 M</td>
<td>Tuffaceous sandstone with occasional rounded conglomeratic clasts.</td>
</tr>
<tr>
<td>1 M</td>
<td>Poorly sorted rounded conglomerate, clasts are hornblende andesite up to 10 cm across. Towards the bottom there are lenses of sandstone interbedded with mudstone.</td>
</tr>
</tbody>
</table>

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Plate 8.1a. Typical exposure of massive sandstone within the Goro-goro Member, S. Goro-goro.

Plate 8.1b. Boulders of volcanic breccia along S. Goro-goro.

Plate 8.1c. Boulder of sandstone with low-angle cross laminations at S. Goro-goro.

Plate 8.1d. Poorly sorted volcanic breccia along the coast between Tg. Sepi and P. Gitalang.
The Kaputusan Formation

Along the east coast of Bacan, from Tg. Sepi to P. Gilalang, there are outcrops of very thickly bedded TPAN interbedded with poorly lithified, poorly sorted and highly porous tuffaceous sandstones with angular-subangular TPAN clasts. Poorly sorted volcanic breccias crop out locally, which have fine tuffaceous matrix and angular TPAN clasts, up to 1 m across, although most clasts are typically <0.4 m across (Plate 8.1d). In places, the clasts are glassy and/or vesicular, and some exhibit stretched vesicles. The exposure at this locality is gently folded.

Lower and Upper Boundaries

The lower boundary was not observed. However, unconformable, fault bounded contacts with the Bacan and South Bacan Formations are inferred from aerial photographic study. An unconformable lower contact with the Ruta Formation is implied from the micropalaeontological and radiometric ages.

The upper contact with the Pacitak Member is depositional, indicated from the consistency in bedding attitudes and similarity in lithology between the two members. Upper contact with the Mandioli Member is depositional as observed in northern P. Mandioli, around Akidabu Village and Tg. Poan area.

Thickness

From the height of the hills in the S. Goro-goro area, having subhorizontal bedding, the thickness of this member is estimated to be at least 500 m, possibly up to 1000 m.

Lateral Extent and Variations

Grey-brown pyroclastic deposits, along the southern coast of Kaputusan Village, consist of interbedded breccio-conglomerate, microbreccia, sandstone and mudstone (Plate 8.2a). The clast-supported breccio-conglomerate unit is poorly sorted, with angular andesite clasts, up to 1 m, typically 0.2 m across. The matrix consists of fine-grained silty material (Plate 8.2b). The sandstone is tuffaceous and poorly sorted, with angular grains showing mm scale parallel lamination. Locally, poorly sorted microbreccia layers are present within the sandstone (Plate 8.2c). Similar outcrops can also be seen on the group of islands south of Kasiruta. Gentle, open folded sequences with parallel and convolute laminations typify these deposits. Both the fold axes and bedding directions are variable (000, 070, 150, 180 and 250) with dips consistently <20°.

At the northern end of Bacan, along the coast of Tg. Minyahi, there is an ~50 m sequence of
Plate 8.2a. Typical exposure of pyroclastic deposits, along the southern coast of Kaputusan Village, consisting of interbedded breccio-conglomerate, microbreccia, sandstone and mudstone.

Plate 8.2b. Poorly sorted breccio-conglomerate boulder along the southern coast of Kaputusan Village. Clasts are HBAN.

Plate 8.2c. Microbreccia layer within sandstone, along the southern coast of Kaputusan Village, ~1 m above Samsyadin and Mohtar's heads.
volcanic conglomerate with at least five fining upward sequences. Detailed description was prevented by rough sea conditions.

Along S. Rain, southern Bacan, there is a sequence of poorly sorted microbreccia, with angular HBAN clasts, which grades into ~150 mm thick parallel laminated sandstone. This fining upward sequence appears to be repeated. The outcrop suffers from pervasive weathering. Detailed description was not possible due to the steepness of the cliff. In southern Bacan, float samples were collected at S. Bibinoi and S. Mau, some affected by pyrite mineralisation. Fig.8.2 shows the distribution of the Goro-goro Member in the Bacan region.

**8.5.2 Petrography and Mineral Chemistry**

**8.5.2.1 Petrography**

Unless stated, all percentages are modal percentages based on visual estimation.

*TPAN* Fresh andesite composed of subhedral-euhedral plagioclase (25%), orthopyroxene (17%), clinopyroxene (7%) and oxide (5%) phenocrysts in a microcrystalline, cryptocrystalline and glassy matrix (Plate 8.3a,b). Plagioclase displays complex twinning and zoning, with glass inclusions in crystal centres and the rims. Both pyroxenes show embayed crystals and often form glomerocrysts. Twinning is present and occasionally the cores of the crystals are corroded. Locally trachytic texture is observed (e.g. BR21). Nine representative samples (BR3, BR4, BR16, BR17, BR58, BT36A, BM123, BM361 and BM380) were chosen for mineral chemistry analysis.

*HPAN* Generally fresh andesite consisting of plagioclase (20%), amphiboles (20%) and pyroxene (<5%) phenocrysts in a microcrystalline, cryptocrystalline or glassy matrix (Plate 8.3c,d). Plagioclase occurs mostly as euhedral laths, showing complex twinning and zoning, with glass inclusions present at the centres and the edges of crystals. Euhedral amphibole crystals are usually brown, although some are green, and they show twinning and zonation, with well developed cleavage. The cores are corroded and most of them have "opacite" rimming the crystals. Locally they form glomerocrysts. Pyroxene grains are subhedral and twinned, showing pleochroic haloes, and are mostly intergrowing with the hornblends. Microphenocryrst oxides (<5%) are also present. Eight samples were selected for microprobe study (BP30, BR49, BR203, BM26, BM117, BM131, BM407 and BM453).
Figure 8.2 Distribution of the Kaputusan Formation in the Bacan region.
The Kaputusan Formation

HBAN  Petrographically very similar to HPAN, with the notable absence of pyroxene (Plate 8.4a,b). Four samples were analysed for mineral chemistry (BR82, BR174, BR185 and BM135).

HBIAN  Similar in character to HPAN, with the addition of minor biotite phenocrysts (Plate 8.4c,d). This is the most affected by alteration. Seven samples (BR53, BR197, BM161, BM256, BM355, BM363 and BM538) were chosen for mineral chemistry analysis.

Pyroclastic Rocks  The pyroclastic sandstones from the Tg. Sepi area consist of HPAN/TPAN clasts (e.g. BR2 and BR27). They are characterised by point contacts and consist of both pyroxenes (15%), plagioclase (15%), both green and brown hornblende (some rimmed by opacite, 10%) and oxides. Microcrystalline plagioclase and glass form the matrix (Plate 8.5a,b).

8.5.2.2 Mineral Chemistry
Mineral chemistry results are discussed in descending order of abundance, starting with phenocrysts, followed by secondary minerals. Fig.8.3 is a map showing the localities of samples used in analytical work. This map shows that TPAN is concentrated in the Goro-goro area, with HPAN, HBAN and HBIAN distributed throughout the studied area.

Plagioclase
In all four andesite types it occurs as three types: [1] phenocrysts (>0.5 mm) and microphenocrysts (0.1-0.5 mm); [2] inclusions in other phenocrysts; and [3] microlites forming part of the groundmass. Compositional zoning in the phenocrysts was detected in nearly all samples including combinations of oscillatory, normal, reverse and patchy zoning. No compositional difference was detected between the phenocrysts and microphenocrysts. The small crystal sizes of the inclusions and microlites limited the number of analyses.

The composition of phenocrysts in TPAN is An_{54.87}, with an average of An_{66} (n=58). Oscillatory zoning is observed (e.g. BR4P11-14, BR16P5-12, BR58P13-16, all core-rim analyses), but the extent never exceeds 12 mol%, with all the cores more sodic than the rims. The inclusions, in pyroxene, are more calcic (An_{67.78}, with an average of An_{71}; n=7), whereas the ground mass microlites are An_{50.70}, with an average of An_{63} (n=4). The decrease in calcium from the inclusion to phenocryst to groundmass suggests a progressively differentiated magma.
Plate 8.3a,b. PPL and XPL of Two Pyroxene Andesite (TPAN; BR17).

Plate 8.3c,d. PPL and XPL of Hornblende Pyroxene Andesite (HPAN; BM26).

Scale for all photomicrographs = 0.5 mm.
Plate 8.4a,b. PPL and XPL of Hornblende Andesite (HBAN; BR82).

Plate 8.4c,d. PPL and XPL of Hornblende Biotite Andesite (HBIB; BR55).

Scale for all photomicrographs = 0.5 mm.
Fig. 8.3 Location of samples from the Kaputusan Formation used in analytical work.

Key:
- □ Mandioli Member
- △ Pacitak Member
- ○ Hornblende Biotite Andesite
- ☽ Hornblende Andesite
- ◦ Hornblende Pyroxene Andesite
- ▣ Two Pyroxene Andesite
- Italics = Not In Situ
The Kaputusan Formation

The phenocrysts in HPAN span An_{44.89}, with an average of An_{59} (n=44). Oscillatory zoning is also observed (e.g. BR49P9-13, BM26P7-9, BM117P7-11 and BM453P2-5, all core-rim analyses), with compositional variations of <33 mol%. The inclusions are An_{31.82}, with an average of An_{57} (n=6), whereas the microlites are An_{49.57} (n=2). The phenocrysts in HBAN are An_{43.77} (average An_{57}; n=26), again with oscillatory zoning (e.g. BR82P1-3, core-rim). Only one analysis each was obtained for the inclusion and microlite phases (An_{49} and An_{46} respectively).

HBAN phenocryst compositions range An_{41.79} (mean An_{57}, n=28), and inclusions are An_{32.64} (mean An_{57}; n=3). The groundmass microlites are An_{52.54} (n=2), except BM363 which has microlites with An_{31.12} (n=3) indicating groundmass albitisation. Normal, reverse and oscillatory zoning were detected, of extent <22 mol%, with the cores always more calcic than the rims (e.g. BR53P4-6, BR197P5-7, BM161P1-2 and BM538P5-9, core-rim analyses). Phenocrysts in BM355 are albitised to An_{46}. Sample BM256 is the only one in the group which lacks plagioclase phenocrysts. The presence of albite in the HBIAN indicates that they have suffered a low temperature metamorphism.

The average plagioclase phenocryst composition in all four andesite groups is labradorite, typical of an acid andesite (Gill, 1981). The variation in An composition and complex zoning patterns indicate a complicated polygenetic history. Plagioclase composition becomes progressively more sodic from TPAN to HPAN, HBAN and HBIAN, indicating that TPAN is of more basic composition (Fig.8.4a).

Clinopyroxene

A common phenocryst and inclusion phases in the TPAN, HPAN and HBIAN. In the TPAN the phenocrysts span En_{39.16}Fs_{4.16}Wo_{41.04} with an average composition of En_{43}Fs_{14}Wo_{40} (n=22). The inclusion phase, in plagioclase, is found only in BM361 (En_{43.45}Fs_{13.14}Wo_{40}; n=2). In HPAN the phenocrysts range En_{36.44}Fs_{7.14}Wo_{41.49} (mean En_{45}Fs_{12}Wo_{43}; n=23), whereas the inclusions are En_{46.45}Fs_{12.11}Wo_{43.47} (mean En_{44}Fs_{14}Wo_{43}; n=4). Zoning is observed, with a general calcium enrichment towards the rim (BR203PX4-6, core-rim). The HBIAN phenocrysts are En_{31.32}Fs_{5.18}Wo_{42.51} (mean En_{43}Fs_{11}Wo_{46}; n=21), with no apparent zoning detected.

There is a general increase in the Al^IV content from TPAN to HBIAN to HPAN (Fig.8.4b), similar to the findings of Morrice & Gill (1986) who attributed this to increasing differentiation. Fig.8.4c shows that TPAN and HPAN clinopyroxene analyses plot in the fields of augite and salite with
The Kaputusan Formation

minor diopside; HBIAN plots in the salite and endiopside fields with minor augite and diopside. No difference was detected between phenocrysts and microphenocrysts.

Amphibole

A prominent phenocryst phase in HBAN, HPAN and HBIAN, occurring with or without fine-grained "opacite". Zoning is absent or weak. There is no compositional difference between the phenocrysts and microphenocrysts. Two types of amphibole are observed: brown and green, of which the former is more common. Gill (1981) suggested that the brown coloration is derived from oxidation indicated by high Fe\textsuperscript{3+}/Fe\textsuperscript{2+} ratios. This is not the case in HPAN, as the green variety exhibit higher ratios (2-8 n=3) than the brown ones (0.1-1.7, n=33). Ballantyne (1990) suggested that amphibole colours are related to Na\textsubscript{2}O and TiO\textsubscript{2} contents. This is not the case in the Kaputusan Formation. Most amphiboles in HPAN are magnesio-hastingsite (Si ~5.8-6.7, Ca 1.7-1.9, total alkalis 2.2-3.0), with subordinate tschermakitic hornblende, magnesio-hornblende and pargasite. Sample BP30 contains magnesio-taramite and anthophyllite, however these appear to be a product of alteration and are heavily rimmed by opacite. Gill (1981) indicated that amphiboles from orogenic andesites contain 39-49 wt% SiO\textsubscript{2}, with the continental andesites having >6.5 Si content. The only sample in HPAN with most analyses >6.5 Si is BR49, located east of the Sibela Metamorphic Complex. Gill (1981) also suggested that amphiboles with <0.5 Al\textsuperscript{vi} content signify precipitation at P\textsubscript{H2O} <9 kb, typical of an orogenic andesite. The Al\textsuperscript{vi} content of amphiboles in HPAN are <0.5, with only BP30A1 having Al\textsuperscript{vi} content of 0.6. The opacites in HPAN (n=6) resemble fresh phenocrysts, with the exception of BR49A7, BM117A7 and BM453A7 which contain no Fe\textsuperscript{3+} and are non-stoichiometric. This may be a function of replacement of amphibole by magnetite.

Brown and green amphiboles are also present in the HBAN; the green varieties have high Fe\textsuperscript{3+}/Fe\textsuperscript{2+} ratios (0.8-1.3, n=3) compared to the brown ones (0-1.2, mostly <0.8, n=18). Most amphiboles are magnesio-hastingsite (Si 6.1-6.7, Ca 1.7-1.9, total alkalis 2.3-2.9), with minor magnesio-hornblende, tschermakitic hornblende, pargasite, actinolitic hornblende and edenitic hornblende. Amphiboles in HBIAN are tschermakitic hornblende, magnesio-hornblende and magnesio-hastingsite (Si 6.0-6.6, Ca 1.6-1.9, alkalis 2.0-2.8), with the green variety again exhibiting higher Fe\textsuperscript{3+}/Fe\textsuperscript{2+} ratios (0.7-2.4, mostly >1, n=11) compared to the brown examples (0.6-1.7, mostly <1, n=8). None of the samples in HBAN or HBIAN consistently contains Si>6.5. The Al\textsuperscript{vi} content in HBAN and HBIAN are <0.5. The opacites in HBIAN (n=3) have low totals (81.1-84.9 wt%), no Fe\textsuperscript{3+} content and are non-stoichiometric. This is attributed to replacement of amphibole by
magnetite.

Only two analyses of amphibole inclusions in plagioclase are available, and these have similar compositions to the phenocrysts. Likewise, plagioclase inclusions in hornblende have compositions comparable with the phenocrysts, indicating that plagioclase crystallises at the same time as amphibole.

There are several trends observed in the amphiboles (excluding the altered examples in BP30, described above). Fig.8.5a shows the variation of Ca and total alkalis (Na+K) contents with Si. There is decreasing Ca and alkalis, with increasing Si, from HPAN-HBAN-HBIAN, possibly indicating increasing fractionation.

Orthopyroxene

Orthopyroxene occurs in TPAN and rarely in HPAN and HBIAN. In TPAN it forms phenocrysts and inclusions. The phenocryst compositions are En\textsubscript{62.74}Fs\textsubscript{32.31}Wo\textsubscript{2.7}, with an average of En\textsubscript{69}Fs\textsubscript{32}Wo\textsubscript{3} (n=29). Rare, weak zoning is present (e.g. BR3PX5-6, core-rim). Orthopyroxene inclusions, in plagioclase, are observed in BR4 and BR58 of compositional range En\textsubscript{71.73}Fs\textsubscript{24.26}Wo\textsubscript{2.3} (n=2).

In HPAN orthopyroxene occurs only in BR49, both as phenocrysts and inclusions in hornblende. The phenocryst compositions are En\textsubscript{56.68}Fs\textsubscript{38.43}Wo\textsubscript{2} (n=2), and the inclusions are En\textsubscript{66.57}Fs\textsubscript{31.32}Wo\textsubscript{2} (n=2). Although there are only limited data available, it appears that the HPAN orthopyroxene inclusions are chemically similar to the phenocrysts in TPAN, suggesting a possible evolutionary trend from TPAN-HPAN. The orthopyroxenes in HBIAN are En\textsubscript{54.69}Fs\textsubscript{37.44}Wo\textsubscript{2.3} (mean En\textsubscript{66}Fs\textsubscript{39}En\textsubscript{2}; n=3). The small number of analyses limits further interpretation.

Fig.8.5b shows that most orthopyroxenes are hypersthene (after Morimoto, 1988). The orthopyroxene in TPAN are more Mg-rich and Fe-poor relative to that of HPAN and HBIAN. Following Lindsley (1982), two pyroxene thermometry estimated for TPAN is ~800-900°C, for HPAN is ~500-800°C and for HBIAN is ~500-700°C at 1 kb.

Fe-Ti Oxides

Oxides occur as phenocryst, microphenocryst and inclusion phases in virtually every sample. There are two types of Fe-Ti oxides: a low Ti (<20 wt%), high Fe\textsuperscript{3+}; and a high Ti (>30 wt%), low Fe\textsuperscript{3+}. 

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variety. No optical difference was detected between the two types of oxides. All analyses plot between ulvöspinel and magnetite (Fig.8.5c). Both varieties are present in TPAN (n=26), with the large phenocrysts consistently higher in Al and lower in Ti contents compared to the microphenocryst and groundmass oxides. Gill (1981) suggested that this Ti-Al variation occurs during eruption and can be attributed to reduction in temperature, $f_o$, or change in composition of coexisting plagioclase and pyroxene. The oxide inclusions in pyroxenes (n=11) have higher Fe$^{3+}$ and lower Fe$^{2+}$ content than the phenocrysts, indicating decreasing $f_o$, during crystallisation (Prévôt & Mergoil, 1973).

The oxides in HPAN (n=30) are the low-Ti type, except for a single analysis (BM131O3). The larger phenocrysts also show higher Al and lower Ti contents. Inclusions in both amphiboles and pyroxene (n=15) have higher Fe$^{3+}$ and lower Fe$^{2+}$ contents relative to the phenocrysts. Inclusions in pyroxene are generally higher in Ti and lower in Mg compared to those in the amphiboles. All analyses in HBAN are the low-Ti type, and the inclusions appear to have higher Fe$^{3+}$ and lower Fe$^{2+}$ values, relative to phenocrysts.

The oxides in HBIAN (n=24) are all low-Ti type. Most larger phenocrysts show higher Al contents, although with variable Ti. Most inclusions show higher Fe$^{3+}$ and lower Fe$^{2+}$ relative to the phenocrysts. Inclusions in biotite are higher in Ti, Al and Mg compared to the inclusions in amphiboles, whereas those in plagioclase are similar to those in amphiboles.

Although the oxide compositions all plotted along the ulvöspinel-magnetite tie-line, there appears to be increasing Ti and Fe$^{2+}$ and decreasing Fe$^{3+}$ contents from HPAN-HBAN-HBIAN-TPAN.

**Biotite**

Biotite is the characteristic mineral of the HBIAN group and always co-exists with amphibole. It normally has an "opacite" corona. No systematic variation was detected across the analyses (n=17). These biotite-bearing rocks have some of the highest whole rock K$_2$O values, with SiO$_2$ content >63%, and the presence of biotite indicates quenching temperatures of 800-950°C (Gill, 1981).

**Olivine**

Only one grain of olivine was observed (BM407), a euhedral crystal (Fo$_{86-91}$), locally altered to smectite. Gill (1981) suggested that the average olivine in orogenic andesite is Fo$_{85-85}$, with compositions >Fo$_{82}$ indicative of a xenocryst origin.
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Figure 8.4a. Ternary diagram showing all plagioclase compositions from the Goro-goro Member.

Figure 8.4b. Covariation of Si with Al" in clinopyroxene of the Goro-goro Member.

Figure 8.4c. Pyroxene quadrilateral showing all clinopyroxene compositions from the Goro-goro Member.

Figure 8.5a. Covariations of Ca and total alkalis (Na + K) contents with Si in amphiboles of the Goro-goro Member.

Figure 8.5b. Pyroxene quadrilateral showing all orthopyroxene compositions from the Goro-goro Member.

Figure 8.5c. Ternary diagram showing all Fe-Ti oxide compositions from the Goro-goro Member.
Accessory Phases
Apatite occurs as euhedral microphenocrysts or inclusions in amphibole, pyroxene and plagioclase. It is present in all petrographical groups (1 analysis in TPAN, 10 in HPAN, 7 in HBAN and 4 in HBIAN) with no detectable compositional difference. Only a single crystal of pyrrhotite was recognised (BT3603-4). This cubic microphenocryst is bronze in colour, which distinguishes it from pyrite. It is the only S-bearing mineral present.

Quartz and Calcite
There are two types of quartz: [1] embayed crystals (?xenocrysts) and [2] the products of devitrifying glass. The former occur in all HBIAN samples, while the latter occur in altered samples (BR49, BM117, BM135). Calcite occurs only in BM135, BM161 and BM363, associated with alteration of plagioclase, pyroxene or forming veins.

Smectite and Chlorite
These two phyllosilicates are fine-grained and occur as alteration products of pyroxene, amphibole, olivine or glass. BM123 (TPAN) contains a pleochroic, green-brown smectite mineral replacing pyroxene, similar in composition to nontronite (Deer et al., 1992), but with higher FeO*, Mg, K and lower Na. BP30 (HPAN) contains a red-orange smectite mineral replacing pyroxene, of similar composition to montmorillonite (Deer et al., 1992) with higher FeO*, K and lower Mg, Ca contents. Chlorite replaces amphibole in BM135 (HBAN).

Zeolite
There are two types of zeolites: thomsonite and heulandite/stilbite. Thomsonite occurs as a fibrous mineral with radial habit in BM363 (HBIAN). Heulandite and stilbite in BM131 (HPAN), BM135 (HBAN) and BM161 (HBIAN) are either stubby or fibrous, all replacing plagioclase. These samples were collected from the same area in the S. Kaputusan, indicating localised alteration. The analyses are comparable to values quoted by Gottardi & Galli (1985) and Hakim (1989).

Glass
The glasses analysed in all samples are very acidic, similar to those of the present day Halmahera arc (Keunen, 1935). In all analyses the abundance of SiO₂ and K₂O distinguishes glass compositions from plagioclase microlites. Table 8.1 compares the results from the different petrographical groups. In TPAN and HBIAN the host mineral is plagioclase, while in HPAN it is amphibole.
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Table 8.1 Summary of glass compositions from the different petrographical groups. Ic = Inclusions in phenocrysts.

<table>
<thead>
<tr>
<th></th>
<th>TPAN</th>
<th>Ic</th>
<th>HPAN</th>
<th>Ic</th>
<th>HBAN</th>
<th>Ic</th>
<th>HBIAN</th>
<th>Ic</th>
</tr>
</thead>
<tbody>
<tr>
<td>n</td>
<td>29</td>
<td>6</td>
<td>21</td>
<td>1</td>
<td>16</td>
<td>0</td>
<td>18</td>
<td>1</td>
</tr>
<tr>
<td>Si (10 O)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>wt%</td>
<td>3.32-4.45</td>
<td>3.32-4.31</td>
<td>3.03-4.60</td>
<td>4.37</td>
<td>3.37-4.27</td>
<td>-</td>
<td>3.19-4.52</td>
<td>4.29</td>
</tr>
<tr>
<td>TiO₂ wt%</td>
<td>&lt;1.4</td>
<td>&lt;1.4</td>
<td>&lt;0.7</td>
<td>0</td>
<td>&lt;0.3</td>
<td>-</td>
<td>&lt;0.6</td>
<td>0.4</td>
</tr>
<tr>
<td>K</td>
<td>0.04-0.51</td>
<td>0.13-0.27</td>
<td>0.03-0.46</td>
<td>0.03</td>
<td>0.01-0.28</td>
<td>-</td>
<td>0.03-1.08</td>
<td>0.26</td>
</tr>
</tbody>
</table>

It is apparent that there are hardly any variation in the groundmass glass composition. The K content of the HBIAN, however, extends to markedly higher values than the others, suggesting a more K-rich liquid composition.

8.5.2.3 Implications

Based on the presence of inclusions, the order of crystallisation of the Goro-goro Member is: Fe-Ti oxide, apatite, plagioclase ± orthopyroxene, clinopyroxene, amphibole, biotite. The phryic nature of the andesites implies crystallisation with magmatic pressure <8 kb. Ubiquitous plagioclase indicates that the pre-eruptive magma contains <2-5 wt% H₂O with P_H₂O<P_total (Gill, 1981). Resorption is indicated by abundant glassy inclusions in plagioclase phenocrysts, aligned along twin planes with diffuse internal boundaries (Plate 8.5c). In contrast, rapid crystal growth is inferred from the presence of glassy inclusions along the plagioclase twin planes, showing sharp internal boundaries (Plate 8.5d). Resorption may be initiated due to decreasing P, increasing P_H₂O, or mixing with hotter magma, while the growth mechanism requires rapid crystallisation during vapour saturation or mixing with colder magma (Morrice & Gill, 1986).

Crystallisation of amphiboles requires P <8 kb, with magmatic H₂O greater than 3 wt%, normally 6-10 wt% (Gill, 1981). The amphibole-bearing andesite is therefore interpreted as a product of the restricted, upper portion of Na and H₂O enriched magma. This is consistent with the normal occurrence of amphibole-bearing andesites in an active volcano; at a high topographic and stratigraphic level. The eruptive temperature of the TPAN is 800-900°C, while the HPAN, HBAN and HBIAN are 500-800°C.

Possible fractionation from TPAN to HPAN is indicated by plagioclase compositions, composition of pyroxene inclusions found in HPAN and lack of high-Ti oxides in groups other than TPAN.
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Plate 8.5a. PPL of pyroclastic rock from the Goro-goro Member (BR27).

Plate 8.5b. XPL of plagioclase phenocryst with glassy inclusions along twin planes with diffuse internal boundaries (BR197).

Plate 8.5c. as Plate 8.5b, except with sharp internal boundaries (BM453).

Scale for all photomicrographs = 0.5 mm.
The Kaputusan Formation

The presence of biotite phenocrysts in HBIAN suggests a more acid magma, possibly due to contamination with quartz-bearing rock, hence its coexistence with quartz xenocrysts. This is further supported by the occurrence of olivine xenocrysts and a bimodal plagioclase composition in some samples.

The mineral chemistry study therefore demonstrates that the Goro-goro Member is a product of volcanism with a complex magmatic history, including fractionation, possibly magma separation (immiscibility), replenishment and mixing, phenocryst resorption, contamination and multiple eruption.

The presence of zeolites, quartz, smectite and chlorite indicates zeolite facies metamorphism (Boles, 1984). The geographically restricted occurrence of this metamorphism suggests a local hydrothermal origin. The HBIAN group appears to be more affected by this metamorphism.

8.5.3 Whole Rock Chemistry

The aim of the whole rock chemistry study was: [1] to characterise the formation, [2] to decipher its tectonic setting and [3] to observe any temporal and petrological variations and identify the processes involved. Twenty samples were chosen for whole rock analysis: BR4, BR16, BR58, BT36A, BM361 and BM380 (TPAN); BR49, BR203, BM26, BM117, BM131, BM407, BM453 and BP30 (HPAN); BR82, BR174 and BM135 (HBAN); BR53, BR197 and BM256 (HBIAN). All samples are in situ except BR174, BM256 and BM407.

8.5.3.1 Major Elements

Petrographical and mineralogical studies indicate that all samples analysed are relatively fresh, indicated by the low LOI (mostly <2%). Fig. 8.6a is a plot of SiO₂ against K₂O (Basaltic Volcanism Study Project, 1981) showing that TPAN are calc-alkaline andesites, except for BT36A which is a basalt. The majority of the amphibole-bearing rocks plot in the high K field. Most HPAN are basaltic andesites, with one dacitic sample (BR49). HBAN and HBIAN plot mostly in the andesite field, with one dacitic sample (BR53). Samples BR203 and BT36A are shoshonitic. Using the IUGS chemical classification (Le Bas & Streckeisen, 1991), these samples plot in the fields of basalt, basaltic andesite, andesite and dacite (Fig. 8.6b).

Figs. 8.6c-i are Harker diagrams of SiO₂ against MgO, Na₂O, Al₂O₃, CaO, Fe₂O₃, TiO₂ and FeO*/MgO. In Fig. 8.6c, TPAN, with the exception of BT36A, has the highest MgO followed by
The Kaputusan Formation

HBAN and HBIAN, which may be related to the modal abundance of pyroxene. Fig. 8.6c shows that although there is no clear relationship between Na₂O and SiO₂, the Na₂O content decreases from HPAN to HBAN to HBIAN to TPAN, which may be related to the modal abundance of amphibole (e.g. Morrice & Gill, 1986). Figs. 8.6d-g show decreasing Al₂O₃, CaO, Fe₂O₃ and TiO₂ with increasing SiO₂. There is also a decrease in the amount of CaO, Fe₂O₃ and TiO₂ from TPAN and/or HPAN to HBAN and/or HBIAN. These trends suggest fractional crystallisation of plagioclase and oxide, whereby TPAN changes progressively to HPAN, HBAN and finally HBIAN. In contrast, the FeO*/MgO, which reflects the amount of Fe enrichment, does not show a simple relationship with increasing SiO₂ (Fig. 8.6i).

Table 8.2 compares the mean chemical compositions of samples from each petrographical groupings with IACA and HKCA. IACA is taken from the South Sandwich andesite, whereas HKCA is from Rinjani (Sunda Arc) basaltic-andesite. The K₂O concentration of the Goro-goro Member is the highest, whereas the MgO and Al₂O₃ contents are comparable to HKCA. The Kaputusan volcanic rocks also have the lowest TiO₂, FeO*, MnO and Na₂O, possibly reflecting the higher degree of fractional crystallisation compared to that of the South Sandwich or Rinjani rocks. Additionally, major trends can be recognised within the groupings; Fe₂O₃, MgO and CaO decrease from TPAN to HPAN to HBAN to HBIAN, while K₂O increases, which is consistent with the fractional crystallisation of plagioclase, pyroxene, Fe-Ti oxide and biotite. The major element chemistry of the Kaputusan Formation rocks is comparable with the modern Halmahera arc (Morris et al., 1983), the Weda Group of Halmahera (Hakim, 1989) and the Sangihe arc (Morrice & Gill, 1986).

8.5.3.2 Trace Elements

To investigate chemical groupings within the Goro-goro Member, covariation plots of incompatible trace element ratios were used (La and Y = REE; Nb, Th and Zr = HFS; Rb = LILE). Figs. 8.7a-e show the covariation of La/Nb, Th/Nb and Rb/Zr against Zr/Nb (indicator of partial melt), and Y/Zr and Th/Nb against Th/Zr. Four major groupings can be recognised in all diagrams: Group I (BR49, BR53 and BM256); Group II (BR4, BR16, BR58, BM361 and BM380); Group III (BR174, BM26 and BT36A); and Group IV (BR82, BR197, BM117, BM131, BM135, BM407 and BM453). Samples BR203 and BP30 cannot easily be classified (Figs. 8.7c,e), which may be attributed to low temperature metamorphism (e.g. BP30) or different degrees of LILE enrichment in the shoshonitic rocks (e.g. BR203), due to mantle source heterogeneity induced by slab derived LILE-rich fluid (Keller, 1983). Figs. 8.7c,e show that Rb/Zr varies systematically with Zr/Nb, and
Figure 8.6a. Volcanic rocks classification based on bulk rock K\textsubscript{2}O against SiO\textsubscript{2}, applied to the Goro-goro Member.

Figure 8.6b. Volcanic rocks classification based on total alkali against SiO\textsubscript{2}, applied to the Goro-goro Member.
Figures 8.6c-i. Harker-type diagram, showing covariations of SiO₂ with MgO, Na₂O, Al₂O₃, CaO, Fe₂O₃, TiO₂ and FeO*/MgO.
Th/Nb varies with Th/Zr, indicating that most of the samples are genetically related. Group III has unusually high Sr (>1700 ppm) and La (51-80 ppm) and consequently high La/Nb, which may be related to the fact that this group has some of the highest K$_2$O contents (2.1-5.7 wt%) and could therefore be anomalously LILE-enriched, including Rb, Sr, Ba, Zr (Keller, 1983).

Table 8.2 Comparing the mean whole rock chemistry of the Kaputusan Formation rocks with IACA and HKCA. Data taken from Wilson (1989).

<table>
<thead>
<tr>
<th>Wt%</th>
<th>TPAN</th>
<th>HPAN</th>
<th>HBAN</th>
<th>HBIAN</th>
<th>IACA</th>
<th>HKCA</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO$_2$</td>
<td>57.36</td>
<td>56.38</td>
<td>58.14</td>
<td>63.98</td>
<td>60.23</td>
<td>55.49</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>0.64</td>
<td>0.76</td>
<td>0.64</td>
<td>0.53</td>
<td>1.12</td>
<td>0.91</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>17.18</td>
<td>18.04</td>
<td>18.01</td>
<td>16.61</td>
<td>14.13</td>
<td>18.45</td>
</tr>
<tr>
<td>FeO*</td>
<td>6.60</td>
<td>6.27</td>
<td>5.52</td>
<td>4.23</td>
<td>10.53</td>
<td>8.32</td>
</tr>
<tr>
<td>MnO</td>
<td>0.13</td>
<td>0.12</td>
<td>0.13</td>
<td>0.08</td>
<td>0.22</td>
<td>0.16</td>
</tr>
<tr>
<td>MgO</td>
<td>4.26</td>
<td>3.51</td>
<td>3.30</td>
<td>2.51</td>
<td>2.13</td>
<td>3.10</td>
</tr>
<tr>
<td>CaO</td>
<td>8.38</td>
<td>7.63</td>
<td>6.63</td>
<td>4.71</td>
<td>6.12</td>
<td>7.47</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>2.64</td>
<td>3.68</td>
<td>3.69</td>
<td>3.38</td>
<td>3.78</td>
<td>4.09</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>1.50</td>
<td>2.04</td>
<td>2.19</td>
<td>3.50</td>
<td>1.18</td>
<td>1.60</td>
</tr>
<tr>
<td>P$_2$O$_5$</td>
<td>0.27</td>
<td>0.41</td>
<td>0.33</td>
<td>0.22</td>
<td>0.26</td>
<td>0.28</td>
</tr>
</tbody>
</table>

These groupings are also evident on covariation diagrams of incompatible trace elements (Ni, Cr and Y against Zr; Figs.8.8a-c). Fig.8.9 shows the geographical distribution of these samples according to their chemical grouping. Group I is concentrated in northern Mandioli and the islands west of Kaputusan, Group II in the Goro-goro area, Group III in the central and southern blocks of Bacan and Group IV in the Kaputusan coast and the northwest region of Bacan. These groupings are therefore interpreted as related to the spatial distribution of volcanic centres, indicating different volcanoes. It is worth noting that Group I is dominated by HPAN, Group II by TPAN and Groups III and IV by a mixture of HPAN, HBAN and HBIAN. Table 8.3 lists the incompatible trace element ratios of the Goro-goro Member samples (excluding BP30 and BR203), illustrating the chemical basis of these groupings, in particular by the Rb/Zr, Zr/Nb and Th/Zr ratios.
Figures 8.7a-e. Covariations of trace element ratios (Zr/Nb against La/Nb, Th/Nb and Rb/Zr, Th/Zr against Y/Zr and Th/Nb) to distinguish the different chemical groupings within the Goro-goro Member.
Figures 8.8a-c. Trace element covariations (Zr against Ni, Cr and Y) showing the different chemical groupings in the Goro-goro Member.
Table 8.3 Comparison of the mean trace element ratios within the Goro-goro Member. Range values are given in parenthesis.

<table>
<thead>
<tr>
<th>Ratios</th>
<th>Group I</th>
<th>Group II</th>
<th>Group III</th>
<th>Group IV</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ba/La</td>
<td>17.8 (16.4 - 20.5)</td>
<td>18.6 (15.3 - 25.4)</td>
<td>9.6 (8.8 - 10.4)</td>
<td>17.6 (12.4 - 26.3)</td>
</tr>
<tr>
<td>La/Nb</td>
<td>4.0 (2.1 - 8.0)</td>
<td>6.5 (5.7 - 7.2)</td>
<td>19.5 (17.0 - 21.6)</td>
<td>10.4 (7.5 - 11.7)</td>
</tr>
<tr>
<td>Rb/Zr</td>
<td>0.7 (0.6 - 0.7)</td>
<td>0.2 (0.2 - 0.3)</td>
<td>0.2 (0.1 - 0.2)</td>
<td>0.5 (0.3 - 0.6)</td>
</tr>
<tr>
<td>Ba/Y</td>
<td>29.9 (16.3 - 56.9)</td>
<td>9.1 (7.4 - 13.5)</td>
<td>29.4 (22.3 - 42.5)</td>
<td>29.1 (13.2 - 39.5)</td>
</tr>
<tr>
<td>Ce/Y</td>
<td>2.3 (2.0 - 2.9)</td>
<td>1.1 (1.0 - 1.4)</td>
<td>6.8 (5.0 - 9.8)</td>
<td>3.5 (1.4 - 4.4)</td>
</tr>
<tr>
<td>Y/Zr</td>
<td>0.1 (0.1 - 0.2)</td>
<td>0.3 (0.2 - 0.3)</td>
<td>0.1 (0.1 - 0.2)</td>
<td>0.2 (0.1 - 0.4)</td>
</tr>
<tr>
<td>Zr/Nb</td>
<td>17.6 (12.5 - 26.7)</td>
<td>52.9 (48.6 - 59.4)</td>
<td>61.1 (58.2 - 63.1)</td>
<td>37.8 (33.8 - 40.4)</td>
</tr>
<tr>
<td>Th/Zr</td>
<td>0.08 (0.07 - 0.08)</td>
<td>0.02 (0.02 - 0.03)</td>
<td>0.08 (0.06 - 0.10)</td>
<td>0.11 (0.08 - 0.15)</td>
</tr>
</tbody>
</table>

To explore fractional crystallisation trend further, compatible trace elements, such as Ti, Ni, Cr and Sc, and the ratios of compatible versus incompatible elements (Rb/Sc and Ti/Zr) are plotted against SiO₂ in Figs.8.10a-f. Most of the chemical groups are reasonably self-consistent, with the exception of Group IV, suggesting a more complicated petrogenetic history. Furthermore, in the Ti/Zr against SiO₂ diagram (Fig.8.10f), there appears to be two subgroupings of Group IV: those with higher Ti/Zr (BR82, BM131 and BM407) are all located on northern Bacan. Although there are clear differences in the way each group behaves, the limited data prevented further conclusions.

The Goro-goro Member samples have Ba/La ratios of 8.76-26.32 which is higher than usual in arc volcanic rocks (Arculus & Powell, 1986). The La/Nb ratio of 2-6 is typical of arc volcanic rocks (Saunders et al., 1980) and most of the Goro-goro Member samples fall into this category, except samples from Groups III and IV which have higher ratios.

8.5.3.3 Tectonic Discrimination Diagrams

On the Ti-Zr-Y tectonic discrimination diagram (Pearce & Cann, 1973), the Goro-goro Member plots mostly in the Calc-Alkaline Basalt (CAB) field (Fig.8.11a). Due to the high Zr contents of Groups I and III, they plot outside of CAB field; sample BM407 of Group IV plot outside of CAB field because of its high Y content.
Figure 8.9. Distribution of the different chemical groupings within the Goro-Goro Member.
Figures 8.10a-f. Covariations of SiO$_2$ with Ti, Ni, Cr, Sc, Rb/Sc and Ti/Zr showing fractional crystallisation trends.
On the Nb-Zr-Y diagram (Fig.8.1b) (Meschede, 1986) Group I plots in the Volcanic Arc Basalt (VAB) & Within Plate Tholeiite (WPT) field, except BM256 which plots in the Within Plate Basalt (WPB) field. Group II and IV plots in the VAB & WPT field and N-MORB & VAB field; Group III plots outside of the VAB & WPT field and WPB field, due to their high Zr content. Fig.8.1c is a similar diagram (Mullen, 1983), utilising the major elements TiO₂-P₂O₅-MnO and by this diagram Group I plots in the CAB field, except BM256 which plots in the Ocean Island Alkali (OIA) field. Group II and IV plot in the CAB field; Group III plots in the OIA field, with BT36A plotting in CAB field.

When normalised against MORB (Fig.8.12a-d; after Pearce, 1982), the Goro-goro Member rocks shows LIL-enrichment, pronounced Nb depletion, flat Ce-Sc distribution and strongly depleted Cr, all of which are characteristic of arc-related rocks. When normalised against chondrite (Sun, 1980), the rocks show Rb, K and Sr enrichment and Nb, La depletion, again typical of arc volcanism (Fig.8.12e-h). The enrichments are attributed to fluids from the dehydrated subducted slab, whereas depletions are due to fractionation processes (Pearce, 1982). Although not presented here, covariation diagrams of Cr against Y and Ce/Sr; Zr against Ti and Zr/Y (Pearce & Norry, 1979); and Ti/Y against Nb/Y all indicate an island arc volcanic origin for the Goro-goro Member.

8.5.3.4 Implications

Tectonic discrimination diagrams successfully show that the Goro-goro Member is a product of island arc volcanism. Petrographical, mineralogical and chemical similarities with the Weda Group (Hakim, 1989; Hakim & Hall, 1991) of western Halmahera are recognised. The present location of the Kaputusan Formation strongly suggest this is a product of the eastward subduction of Molucca Sea Plate.

Major element chemistry indicates a magma that has undergone a high degree of fractional crystallisation, of plagioclase, pyroxene, Fe-Ti oxide and biotite, resulting in the observed volcanic diversification. An overall fractionation trend of TPAN to HPAN to HBAN to HBIAN is recognised, which is analogous to the findings in the Weda Group of western Halmahera (Hakim, 1989).

The different chemical groupings suggest the presence of at least four different, but genetically related, volcanic centres each with a distinct chemistry and dominant rock type. These are the South Bacan Group (I), the Goro-goro Group (II), the North Mandioli / South Kasiruta Group (III)
Figures 8.11a-c. Ternary tectonic discrimination diagrams applied to the Goro-goro Member.
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Figures 8.12a-d. MORB normalised diagrams applied to the Goro-goro Member Groups I-IV.

Figures 8.12e-h. Chondrite normalised diagrams applied to the Goro-goro Member Groups I-IV.
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and the Kaputusan / North Bacan Group (IV). Compatible trace element variations with SiO₂ suggest fractionation within each chemical grouping, with Group IV possibly showing two subgroupings, on northern Bacan and on the Kaputusan Coast.

There have been several published opinions on the cause of the occurrence of high-K rocks (Group III): [1] source heterogeneity due to metasomatism by fluids (high LILE) derived from a subducted slab (Keller, 1983); [2] magma derived from a subduction modified OIB source (high HFS) component (Wilson, 1989); [3] sediment incorporation and crustal contamination (Whitford et al., 1977); and [4] a combination of any of these sources (e.g Ellam et al., 1988). In the Kaputusan Formation, sediment incorporation may be unlikely (Tera et al., 1986), whereas crustal contamination can be ruled out due to the low ⁸⁷Sr/⁸⁶Sr ratios of some of these high-K samples (E. Forde, pers.comm, 1993). Despite these differing opinions, it appears that most high-K calc-alkaline rocks were erupted in a locally extensional setting within a dominantly convergent zone (Nicholls & Whitford, 1983; and references therein). Edwards (1990) suggested that metasomatic fluid, frozen in the mantle wedge, can be easily remelted following an extensional phase. All of these imply a more complex history of magmatic evolution than that suggested for the Weda Group (Hakim, 1989).

8.5.4 Age Determination

Sixteen samples were analysed by the K-Ar technique. The main Neogene volcanism on Bacan started, judging from the oldest ages, at ~7 Ma and appears to have been relatively continuous up to the Late Pliocene-Pleistocene. There are four main concentrations of ages (values given in brackets are the average of the two isochron ages): the first at ~7.0-6.5 Ma (6.84 ± 0.06 Ma), the second at ~5.4-4.7 Ma (4.97 ± 0.14 Ma), succeeded by volcanicity at ~2.4-2.2 Ma (2.33 ± 0.13 Ma) and the last event at ~1.2-0.6 Ma, which may be reset ages. These concentrations of ages may be actualistic or a function of incomplete sampling. Fig.8.13a is a plot of the ages in a bar diagram, while Figs.8.13b-i are isochron plots, including samples from the Bacan and South Bacan Formations which have been affected by Kaputusan Formation volcanism. Details of the K-Ar technique, isochron plots and sample descriptions are given in Appendix D.

These four concentrations of ages largely correspond with the chemical groupings, although there are some overlap between groups I, II, and III, indicating that during each period of eruption (as defined by a concentration of ages), several volcanoes (as defined by chemical groupings) may be simultaneously active. Fig.8.14 shows the spatial distribution of samples analysed, with their
Figure 8.13a K-Ar Isotopic Ages from the Kaputusan Formation Rocks.
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Figures 8.13b-i. Isochron diagrams showing the four main concentration of ages in the Goro-goro Member (0.5, 2.2, 5.1 and 7 Ma).
Figure 8.14 Location of Samples from the Kaputusan Formation for Radiometric Dating
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ages, possibly suggesting that the volcanic centres migrated northward with time (Group I is oldest
and at south Bacan, Groups II and III are in central Bacan and Group IV which is the youngest,
located at north Bacan).

Table 8.4 Summary of the K-Ar results from the Kaputusan Formation. Data arranged by chemical groupings and age. WR
= whole rock, HB = hornblende, PX = pyroxene.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Material</th>
<th>Grain size (µ)</th>
<th>%K (1σ error)</th>
<th>W' for Ar (g)</th>
<th>4Ar (1σ error)</th>
<th>4Ar (1σ error)</th>
<th>Age (Ma. 2σ error)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BR49</td>
<td>wr</td>
<td>250-500</td>
<td>2.104 ± 1.00%</td>
<td>0.0590 ± 0.00</td>
<td>0.6119 ± 0.42%</td>
<td>78.66 ± 0.0007</td>
<td>7.47 ± 0.0700</td>
</tr>
<tr>
<td>BRM26</td>
<td>wr</td>
<td>250-500</td>
<td>1.632 ± 1.00%</td>
<td>0.3493 ± 0.00</td>
<td>0.3035 ± 2.98%</td>
<td>72.16 ± 0.00</td>
<td>4.78 ± 0.3000</td>
</tr>
<tr>
<td>BR53</td>
<td>wr</td>
<td>250-425</td>
<td>1.318 ± 2.3%</td>
<td>1.2230 ± 0.00</td>
<td>0.0234 ± 14.28%</td>
<td>93.13 ± 0.0003</td>
<td>0.457 ± 0.1390</td>
</tr>
<tr>
<td>BR58</td>
<td>wr</td>
<td>250-425</td>
<td>1.066 ± 1.00%</td>
<td>2.2149 ± 0.00</td>
<td>0.3044 ± 1.78%</td>
<td>58.28 ± 0.0009</td>
<td>6.70 ± 0.2800</td>
</tr>
<tr>
<td>BR16</td>
<td>wr</td>
<td>250-425</td>
<td>1.498 ± 5.31%</td>
<td>1.1052 ± 0.00</td>
<td>0.3414 ± 4.19%</td>
<td>80.11 ± 0.0009</td>
<td>5.52 ± 0.7900</td>
</tr>
<tr>
<td>BRM61</td>
<td>wr</td>
<td>125-250</td>
<td>0.966 ± 1.29%</td>
<td>0.3237 ± 0.00</td>
<td>0.2073 ± 6.75%</td>
<td>86.46 ± 0.0007</td>
<td>5.52 ± 0.7600</td>
</tr>
<tr>
<td>BR174</td>
<td>hb</td>
<td>125-250</td>
<td>2.258 ± 1.00%</td>
<td>1.0276 ± 0.00</td>
<td>0.448 ± 4.57%</td>
<td>81.35 ± 0.0007</td>
<td>10.0 ± 0.4700</td>
</tr>
<tr>
<td>BM26</td>
<td>hb px</td>
<td>125-250</td>
<td>1.998 ± 2.69%</td>
<td>0.6240 ± 0.00</td>
<td>0.2121 ± 3.73%</td>
<td>77.58 ± 0.0007</td>
<td>4.97 ± 0.4600</td>
</tr>
<tr>
<td>BT16A</td>
<td>wr</td>
<td>125-250</td>
<td>1.461 ± 1.57%</td>
<td>0.6465 ± 0.00</td>
<td>0.2792 ± 13.89%</td>
<td>93.06 ± 0.0003</td>
<td>4.91 ± 1.3700</td>
</tr>
<tr>
<td>BR197</td>
<td>wr</td>
<td>125-250</td>
<td>2.166 ± 1.00%</td>
<td>0.0544 ± 0.00</td>
<td>0.4664 ± 24.21%</td>
<td>95.84 ± 0.0007</td>
<td>5.53 ± 2.6800</td>
</tr>
<tr>
<td>BM453</td>
<td>wr</td>
<td>250-425</td>
<td>1.382 ± 1.56%</td>
<td>2.2700 ± 0.00</td>
<td>0.1366 ± 4.69%</td>
<td>81.91 ± 0.0007</td>
<td>2.71 ± 0.2100</td>
</tr>
<tr>
<td>BM131</td>
<td>wr</td>
<td>250-425</td>
<td>1.544 ± 2.75%</td>
<td>2.4686 ± 0.00</td>
<td>0.1526 ± 5.74%</td>
<td>84.74 ± 0.0007</td>
<td>2.50 ± 0.3200</td>
</tr>
<tr>
<td>BR82</td>
<td>wr</td>
<td>125-250</td>
<td>1.558 ± 1.56%</td>
<td>1.0238 ± 0.00</td>
<td>0.1178 ± 2.42%</td>
<td>86.66 ± 0.0007</td>
<td>2.23 ± 0.1300</td>
</tr>
<tr>
<td>BM135</td>
<td>wr</td>
<td>250-500</td>
<td>1.262 ± 1.00%</td>
<td>2.2399 ± 0.00</td>
<td>0.0443 ± 17.82%</td>
<td>94.52 ± 0.0007</td>
<td>0.60 ± 0.3200</td>
</tr>
<tr>
<td>BR203</td>
<td>hb px</td>
<td>125-250</td>
<td>1.890 ± 2.69%</td>
<td>1.0420 ± 0.00</td>
<td>0.3114 ± 13.13%</td>
<td>92.59 ± 0.0007</td>
<td>4.22 ± 1.1300</td>
</tr>
<tr>
<td>BR30</td>
<td>wr</td>
<td>250-500</td>
<td>1.088 ± 1.00%</td>
<td>1.2044 ± 0.00</td>
<td>0.1069 ± 7.23%</td>
<td>87.57 ± 0.0007</td>
<td>2.53 ± 0.5700</td>
</tr>
</tbody>
</table>

Out of the 13 mineral separate analyses, only 3 give an acceptable results. In contrast, out of 17
whole rock analyses, 13 were acceptable, indicating that the whole rock analyses are more reliable
than the mineral separates (see Appendix D for details). This is contrary to the accepted view (e.g.
Faure, 1986). Whole rock analyses from these fresh, young glassy volcanic rocks are suitable as
the argon loss is minimal, due to their rapid solidification effectively creating one blocking
temperature.

Temporal Variation

Gill (1981) recommended the whole rock ratio of FeO*/MgO to study temporal variation, as the
effect of differentiation is normalised. Fig.8.15 is a plot of whole rock FeO*/MgO against the age
of the Goro-goro Member. From this plot it is apparent that there are areas where the chemical
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groupings are dominant. Groups I, II and III are the oldest and it appears that Group I (excluding BM256 and BR53 which may be reset) is the oldest eruptive centre, proceeded by Group II and Group III. Group IV is clearly the youngest group. There is a possible enrichment of FeO*/MgO content with time within Groups II and III, implying fractional crystallisation. Group IV, however, show decreasing FeO*/MgO with time, possibly indicating source replenishment. Group I is not interpretable due to minimum data and possible resetting of ages. No petrographical consistency is recognised in this diagram.

The radiometric dating confirms that the Goro-goro Member includes several volcanoes, each with a history of multiple eruptions, lasting ~2 Ma, producing different rock types, with fractional crystallisation as the main mode of magma diversification.

8.6 THE PACITAK VOLCANICLASTIC MEMBER

This member is characterised by poorly lithified, greenish-grey, intercalated volcaniclastic conglomerate, tuffaceous sandstone, siltstone and mudstone.

8.6.1 Lithofacies

Type Locality

P. Pacitak, west of Obit, is the type locality where there are excellent coastal outcrops. On western Pacitak there is a 2 m sequence of fine grained calc-arenite with clay intercalations overlain by a 5 m exposure of carbonaceous siltstone and mudstone interbedded with coarser sandy material. The sandstone layers are greenish grey-brown and tuffaceous, with poorly sorted, angular clasts, often with bioturbations structures. Bed thicknesses are typically ~0.5 m with lamination on mm scale. The darker coloured siltstone and mudstone are poorly stratified due to bioturbation, with numerous burrowing and feeding structures, plant and shell fragments. Convolute bedding, flame and load structures and trough cross ripple lamination are present, indicating the sequence is right way up. Current directions are towards 140 and 150.

On the eastern coast of Pacitak an outcrop of low-dipping sandstone and mudstone is present. The sequence includes pebble beds with limestone and andesitic clasts 0.1-0.2 m in diameter. The clasts increase in size down dip, possibly indicating a palaeoslope. Several pebble beds change laterally to laminated mudstones. The greenish-brown, moderately lithified, porous, tuffaceous sandstone unit is poorly-moderately sorted, and contains angular to rounded, coarse to fine HPAN and HBAN clasts, with a tendency for the finer grains to be more rounded. They form very thin
Figure 8.15. Covariations of FeO*/MgO with K-Ar age of the Groen-Goorn Member.
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to thin beds. Dewatering structures and rip-up clasts are common features (Plate 8.6a). Certain mudstone horizons are better lithified, having a higher sand content with densely packed fining upwards layers rich in wood fragments, seeds and disarticulated biogenic material (bivalves, shells, gastropods, ostracods, corals), forming microcoquina layers, some of which are laterally discontinuous, and are interpreted as storm beds representing distinct, periodic high energy events which can be picked out throughout the sequence (Plate 8.6b). This entire sequence is well laminated with the laminations picked out by carbonaceous debris. There are also slump units within the sequence (Plate 8.6c). Fig.8.16 is a log of the section at eastern Pacitak.

Across the strait on western P. Parapotang there is a 30 m exposure of calcareous siltstone and mudstone, containing storm horizons, and immature sandstones. Siltstone layers may include concretions, up to 0.4 m across, of well lithified coarse, moderately sorted, subangular sandstone. The mudstones have an extremely high organic content and are rich in bivalves and gastropods. Fig.8.17 is a simplified logged section at Parapotang.

At P. Pinangkara, north of Obit, there is an interbedded layer of polymict microconglomerate, immature sandstone and mudstone. The microconglomerate consists of subangular, poorly sorted pumiceous lithic fragments, mudstones and clasts of HBAN up to 0.1 m across. The green-brown sandstone unit contains plant fragments and is slightly calcareous, with very thin parallel and low angle cross bedding. Mudstone rip-up clast and load structures at the base of the sandstone unit indicate right way up.

Lower and Upper Boundaries

There is a conformable lower contact with the Goro-goro Member (see Section 8.5.1). At the northwestern end of Pacitak, this member becomes increasingly calcareous up section, changing to a sandy limestone with parallel lamination, suggesting a gradational upper contact with the Mandioli Member.

Thickness

The thickest sequence observed is ~30 m. The thickness of this member is at least 123 m, calculated from the elevation of P. Pacitak which is composed solely of this member with a 20° dip. From the widespread occurrence of this member, it is estimated that the thickness of it is >200 m.
Plate 8.6a. Dewatering structure and rip-up clasts within the Pacitak Member tuffaceous sandstone.

Plate 8.6b. Microcoquina layer in the Pacitak Member mudstone.

Plate 8.6c. Slump unit within the Pacitak Member. Plates 8.6a-c are all exposed at P. Pacitak.
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Fig. 8.16 Simplified Log of The Kaputusan Formation, Pacitak Member, P. Pacitak, across Batang Lomang Village.

26 M
24 M
22 M
20 M
18 M
16 M
14 M
12 M
10 M
8 M
6 M
4 M
2 M
0 M

Grey mudstone, structureless. Contains 1 cm long organic debris.

Brown tan coarse sandstone, discontinuous.Calcite replaced. Changes laterally to mudstone. Pebbles are up to 5.0 cm long and are subrounded. Graded to claystone.

Grey mudstone, some organic debris, discontinuous. Lower unit is brown coarse sandstone 15-10 cm thick. contains numerous shell fragments. Brown coarse sandstone with numerous shell fragments.

Thin layer of grey mudstone.

Conglomeratic bottom, top is coarse sandstone. Porous xenoliths in angular clasts of the bottom of the unit. Matrix is fine to medium parallel to strike. Pebbles are up to 5.0 cm across and consist of limestone and hornblende. Most of the conglomerate units consist of hornblende andesite clasts.

Muddy sandstone with numerous plant fragments. Contain much magnetite sand.

Storm beds, contain broken fragments of shells, corals, bivalves, gastropods and ostracods.

Sandstone, locally laminated, interstratified with mudstone.

Interbedded sandstone and mudstone, forming slump sequence. Not mapped, but the position is accurate in the stratigraphic column.

Sandstone, interbedded with mudstone, forming parallel lamination.

ATTITUDE 221/02

ATTITUDE 201/03

PHOTO 11-25
PHOTO 11-26
PHOTO 11-27

PHOTO 9-28
PHOTO 9-29

PHOTO 368.39

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Fig.8.17 Sketch log of the Kaputusan Formation, Pacitak Member at P. Parapotang.

Interbedded unit of claystone and shelly sandstone. Laminations picked out by carbonaceous material.

Indurated bed of concretions.

Fining upwards calcarenite. Indurated calcarenite concretion bed

Sand horizon with storm events which are laterally discontinuous.

Indurated sandstone bed.

Coarse sandstone. Organic material picking out laminations.

Calcereous clay unit containing three separate storm events. Abundant fragments of biogenic material within these horizons.

Immature sandstone.

Calcereous claystone with storm events and laminated carbonaceous horizons. Discontinuous concretion bed.

Calcereous claystone with three storm events.

Claystone grading into sandier storm bed.

Laminated claystone. Laminae picked out by carbonaceous material.

KEY

Organics (predominantly wood)
Concretions
Parallel laminations
The Kaputusan Formation

_Lateral Extent and Variations_

A massive conglomerate with clasts of limestone and andesite up to boulder size (0.5 m in diameter) is present in the Tg. Poan region, north of Kaputusan Village (see Section 8.4). The outcrop is at least 1.5 km in length. The polymict conglomerate is matrix-supported by a coarse sand with lithic and hornblende grains. Beds of finer well laminated sands are intercalated throughout the unit and these contain dewatering structures and slumps.

On P. Tambelit a similar sequence to Tg. Poan is present with the predominant clasts in the conglomerate being andesite. The sandy beds are more discontinuous, forming lens-like structures, and include storm beds. Finer muddy sandstones occur in parts and a cleavage has developed. The tops of these muddier units are usually eroded by the base of the subsequent conglomerate.

A sequence in P. Tuduh is similar to that on Tambelit with noticeably finer-grained muddy sandstone interbedded with coarser sandstone. Laminations within these units are picked out by coarse sand and mudstone. The mudstone has a high carbonaceous content and the sandstone is normally graded and contains echinoid fragments. Conglomeratic units are less frequent, but the clast population remains dominated by andesite. The contacts between the finer sandstones and the conglomerates are erosive.

T. Charlton (pers.comm, 1989) reported the occurrence of blue-grey volcaniclastic sandstone interbedded with siltstone containing shell fragments, along S. Ruta in northern Kasiruta. This unit is interpreted to be the Pacitak Member. This is supported by similarities in the sandstone petrography. Samples from this locality are, unfortunately, barren in nannoflora. Fig.8.2 shows the distribution of the Pacitak Member in the Bacan region.

8.6.2 Petrography

The siltstones of the Pacitak Member are mud-supported, with angular grains of twinned plagioclase (10%), brown and green amphiboles (10%), pyroxenes (5%) and opaques (5%). The sandstones are composed of plagioclase, and characteristically rich in glassy and trachytic lithic fragments and Y-shaped glass shards (Plate 8.7a,b). Fragmented plagioclase grains exhibit complex twinning and zoning. Grains have long and point contacts. Pyroxene, amphibole, biotite, opaque and bioclastic fragments (benthonic and planktonic foraminifera and rare shell fragments) are also present in low percentages. Conglomeratic clasts have the same petrography as the rocks of the Goro-goro Member (see Section 8.5.2.1).
8.6.3 Age Determination

A rich foraminiferal fauna was recovered from 40 samples, and the resulting micropalaeontological age determinations span the entire Pliocene. The oldest dates were from samples collected on P. Tambelit (BR180) which yield a Miocene - Pliocene age with reworked Early Miocene species. The youngest sample (BS37) occurred in the Pacitak region where a latest Pliocene (N21) age was recorded (Fig.8.18). Characteristic planktonic foraminifera are *Pulleniata obliquuloculata*, *Globorotalia (Gr.) tumida* Group, *Gr. (M.) menardii*, *Gr. (M.) cultrata*, *Gr. (H.) margaritae* Group, *Gr. (T.) tosaensis tosaensis*, *Gr. (T.) tosaensis tenuithecata*, *Gr. (T.) crassaformis crassaformis*, *Gr. (T.) crassaformis oceana*, *Neogloboquadrina humerosa-dutertrei* Transition, *Nq. pseudopima*, *Nq. pachyderma*, *Globigerina decorperata*, *Gg. angustiumbilicata*, *Gg. apertura*, *Gg. bollii*, *Dentoglobigerina altispira altispira*, *Globoquadrina dehiscens*, *Sphaeroidinellopsis seminulia seminulia* and *Ss. paenedehiscens*. Characteristic nannoflora are *Calcidiscus macintyrei*, *Ceratolithus armatus*, *Discoaster pentaradiatus*, *D. variabilis*, *Helicosphaera kamptneri* and *Reticulofenestra pseudoumbilica*. S.J.Roberts (1993) has established six biozones, based on planktonic foraminifera, in the Pacitak Member. These zones are: *Gr. paralenguensis* (early N17), *Gr. lenguensis* (late N17), *Gr. tumida tumida* (N18), *Gr. margaritae* (N19 to end of Early Pliocene), *Gr. cultrata* (end of Early Pliocene to end of N19) and *Sphaeroidinellopsis* (early N21).

8.6.4 Depositional Environment

The presence of plants, seeds and shells debris, burrows and nannofossil palaeo-environment data suggest that the Pacitak Member was deposited in an oxygen-rich, near shore, shallow marine environment. Water depth must have been shallow enough to allow periodic storms to scour the sea bed, forming the coquina layers (cf. Johnson & Baldwin, 1986). Sedimentary structures such as parallel, trough and convolute bedding and the lack of welded tuffs support this interpretation. Sediment immaturity indicates that the depositional basin was proximal to a volcanic terrain. The conglomerates are interpreted to be the base of the shallow marine sequence. Dewatering structures indicate rapid deposition, whereas slumped units represent either locally steep or tectonically active slopes.

Petrographic analyses of the clasts in the Pacitak Member indicate that they are similar to the pyroclastic deposits of the Goro-goro Member. It is therefore inferred that the sedimentary member is the product of reworking of the Goro-goro Member. The older age of the Goro-goro Member supports this interpretation.
### Figure 8.18 Summary of the Kaputusan Formation Biostratigraphic Dates.

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<th>Letter Stage</th>
<th>Nanofossil Zonation</th>
<th>Foraminifera Zones</th>
<th>Foraminifera Letter Stages</th>
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**Legends:**
- **EMF**: Dr. E.M. Finch
- **SUR**: Dr. Spencer Roberts
- **FTB**: Prof. F.T. Banner

**Note:**
- Foraminifera zonations after Blow (1969, 1979)
- Nanofossil zonations after Martin (1970)
- Foraminifera letter stages after Adams (1984)
8.7 THE MANDIOLI LIMESTONE MEMBER

8.7.1 Lithofacies

Type Locality
The type locality for this member is the northern part of Mandioli, around Akidabu Village, where there are calc-arenites containing volcanioclastic debris, shell fragments and bioturbation structures, interbedded with mudstone. These change progressively up-section to wackestones and packstones, similar in character to the limestones of the Ruta Formation, containing large benthonic and planktonic foraminifera as well as coral, algae and echinoid fragments. At the southeastern end of Mandioli this member is exposed as a massive limestone in cliffs, at least 25 m high, containing large fragments of coral, benthonic foraminifera and bivalves, locally recrystallized.

Lower and Upper Boundaries
At the northern end of Mandioli, there is a lower depositional contact with the Goro-goro Member. The calc-arenite here contains euhedral hornblende crystals, which increase in abundance to the east, until the contact with the HBAN. The lower depositional contact with the Pacitak Member is inferred on northwestern Pacitak (see Section 8.6.1). The upper contact is not seen.

Thickness
Locally, a few tens of metres.

Lateral Extent and Variations
This member is exposed along the north and east coast of Mandioli and around Tg. Poan, north of Kaputusan Village. In the latter locality, pyroclastic rocks together with the conglomerates (Goro-goro Member) draped and asphyxiated older laminated reefs (Mandioli Member) which locally have steep dips. Fig.8.2 shows the regional extent of the Mandioli Member.

8.7.2 Microfacies
Petrographic analysis indicates that the Mandioli Member is a wackestone-packstone (Dunham, 1962). It contains a diverse bioclastic assemblage, dominated by planktonic foraminifera (10%), coral (10%), flattened benthonic foraminifera (5%), red algae (5%) and bivalve (5%) fragments. Minor constituents include broken echinoid and brachiopod fragments. Allochems include plagioclase laths (15%), elongated brown and rarely green amphibole (10%), glassy lithic fragments (10%), pyroxene (5%) and euhedral opaques (5%). Micrite forms the matrix, although locally it is recrystallized (20%). This rock shows textural inversion (Plate 8.8a).
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All bioclasts are disarticulated, abraded (subangular-subrounded) and poorly sorted. Volcaniclastic allochems are angular and moderately sorted (Plate 8.8b). There is no evidence for grading or preferential alignment of allochems.

8.7.3 Age Determination
Nine samples have been dated using large benthonic foraminifera and they yielded Late Miocene-Early Pliocene ages (F.T. Banner, pers.comm, 1992). Characteristic faunas are Amphistegina lessonii, Elphidium, Globigerina, Globigerinoides, Globoratalia tumida, G. menardii, Globorquadrina cf dehiscens, Globigerinella hirsuta, Lenticula, Orbulina universa, Neoratalia, Nephrolepidina trybliopidina nucleoconch and Sphaeroidinellopsis paenedehiscens. Fig.8.18 shows the distribution of micropalaeontological ages in the Kaputusan Formation.

8.7.4 Depositional Environment
Microfacies study reveals that this limestone possesses textural inversion, similar to the Ruta Formation, which represents final deposition in a low energy environment after transportation and previous deposition in a high energy environment. This condition can be achieved on an open shelf or open platform (Flügel, 1982). The abundance of euhedral, angular volcaniclastic debris demonstrates that the basin is proximal to a volcanic centre, suggestive of an open platform basin. The presence of bioturbation structures, abundance of coral, red algae, foraminifera, bivalves, brachiopod and echinoid fragments indicate a shallow marine (10-100 m), normal salinity environment (Flügel, 1982). The localised occurrence of this member suggests that it is a localised platform. A lateral transition to the Goro-goro Member indicates that the two members are contemporaneous.

This member is interpreted to be a product of local shallow marine carbonate build-ups during the Kaputusan Formation volcanism, probably representing small fringing coastal reefs similar to those along the present-day coast.

8.8 STRUCTURE
The Goro-goro Member is affected by gentle open folding. Fold axis orientations are variable (015, 050 and 087). There are two dominant dip directions on western coast of Bacan (090 and 140-200; n=9), whereas on eastern and southern Bacan the dip directions are variable (n=7). Most dips are <20°.
Plate 8.7a,b. PPL and XPL of Pacitak Member sandstone (BM7). Note the presence of elongated glass fragments.

Plate 8.8a. PPL of Mandioli Member limestone containing large benthonic foraminifera (BR209).

Plate 8.8b. XPL of Mandioli Member limestone with andesitic allochems (BR209).

Scale for all photomicrographs = 0.5 mm.
The Kaputusan Formation

The Pacitak Member is also locally gently folded (axis=170). The bedding attitudes are dominated by two directions (110-130 and 210-240; n=10), with most dips <30°. The exposures at northern Pacitak are jointed with 300 and 360 joint directions. Only two bedding attitudes are available from the Mandioli Member (160/30 and 325/48).

Similarity in bedding attitudes between the Goro-goro and Pacitak Members in western Bacan is further evidence of the co-genetic relationship. Folding and the inconsistency in bedding attitudes at east and south Bacan is attributed to recent volcanism and faulting.

8.9 SYNTHESIS AND TECTONIC SIGNIFICANCE

The Kaputusan Formation lies unconformably upon the Ruta Formation. Whole rock geochemistry, mineral chemistry and radiometric age data demonstrate that the Goro-goro Member is a product of complex island arc volcanism, erupting from early Late Miocene (Tortonian) to Pleistocene. It includes at least four eruption centres, each active for ~2 Ma, with a history of multiple eruptions. The oldest volcano occurs in the south Bacan region, followed by the one in the Goro-goro region, the North Mandioli region and then in the north Bacan area. Some of these volcanoes may have been active simultaneously. The South Bacan volcano (Group I) produces mostly HBIAN, the Goro-goro volcano (Group II) produces mostly TPAN, whereas the Mandioli and North Bacan volcanoes (Group III and IV) are largely HPAN and HBAN. Some of the rocks from these two latter groups may have been erupted in an extensional setting within the convergent zone. Magma diversification was achieved mainly by crystal fractionation (from TPAN to HPAN to HBAN to HBIAN), with some evidence of source heterogeneity, magma replenishment, immiscibility, resorption, assimilation and crustal contamination. Localised metamorphism to the zeolite facies is interpreted as due to local hydrothermal activity.

The poorly sorted, angular nature of clasts and presence of parallel, cross lamination and massive beds in the Goro-Goro Member is characteristic of a pyroclastic flow with an associated base surge deposit (Fisher & Schmincke, 1984). The absence of welded tuffs indicates low temperature deposition possibly in a subaqueous environment. This is supported by the presence of the breccia-conglomerate unit, which is interpreted as a lahar deposit. The Goro-Goro Member is therefore interpreted as the product of catastrophic and repeated explosive activity, which is typical of a calc-alkaline, island arc volcanic environment.
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The Upper Miocene-Upper Pliocene Pacitak Member consists of pyroclastics and reworked volcaniclastic material from the Goro-goro Member, deposited in an oxygen-rich, near shore, shallow marine environment, proximal to a volcanic source. The Mandioli Member is interpreted as small fringing coastal reefs developed during the Late Miocene-Early Pliocene, along the palaeocoast, contemporaneous with the Kaputusan Formation volcanism.

Sedimentological, petrographic, mineralogical and chemical characteristics suggest a correlation with the Weda Group Volcanic rocks of the western arm of Halmahera. The Kaputusan Formation is therefore interpreted to be the product of the Neogene Halmahera volcanic arc. Fig. 8.19 shows that when Bacan is rotated counterclockwise ~ 12.5°, it fits quite well with the eastern coast of Kasiruta, having the Kaputusan Formation as the boundary. It is speculated that rifting may be due to local rotation related with Kaputusan Formation volcanism.

This study provides evidence of the migration of volcanic centres northward. Similar rocks, attributed to the subduction of the Molucca Sea Plate under Halmahera can be found from Obi in the south to northern Halmahera (Hall et al., 1991). Current volcanic activity occurs only from P. Makian northward. The cause of cessation of activities south of Makian is speculated to be related to splaying of the Sorong Fault Zone.
Figure 8.19. Bacan rotated 12.5 degrees counterclockwise, showing that the western coast of Bacan fits with eastern coast of Kasiruta.
CHAPTER NINE
INTRUSIVE ROCKS

9.1. INTRODUCTION

9.2 THE NUSA BABI MONZODIORITE (NBM)

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9.2.2 Lithofacies
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9.3 THE SALEH DIORITE

9.3.1 Lithofacies
9.3.2 Petrography and Mineral Chemistry
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9.4 SYNTHESIS AND TECTONIC SIGNIFICANCE
CHAPTER NINE
INTRUSIVE ROCKS

9.1. INTRODUCTION
There are two main intrusive bodies in the Bacan area: the Nusa Babi quartz-monzodiorite and the Saleh diorite. The Nusa Babi intrusive consists mainly of plagioclase, amphibole, biotite and minor quartz with associated 'aplite' dykes. The Saleh diorite consists of euhedral amphibole interlocked with plagioclase. This chapter deals with the lithologies, petrography, mineral and whole rock chemistry and age determinations of the intrusive rocks.

9.2 THE NUSA BABI MONZODIORITE (NBM)

9.2.1 Aerial Photography
On aerial photographs, the quartz-monzodiorite is characterised by its low altitude undulating hills, with a dendritic drainage pattern and light colouration.

9.2.2 Lithofacies
Type Locality
The leucocratic-mesocratic quartz-monzodiorite found on P. Nusa Babi consists of amphibole, biotite, feldspar and quartz with a granular texture. It contains xenocrysts of coarse amphibole, up to 20 mm in length, with a locally subparallel alignment (Plate 9.1a). The rock is cut by leucocratic 'aplite' dykes consisting of fine granular amphibole, feldspar, quartz (Plate 9.1b) and quartz-monzodiorite xenoliths. Mineralogically, the 'aplite' is a monzogranite (LeMaitre, 1989).

Outcrops at the type locality are jointed pervasively with directions (009 & 096) similar to the Bacan and South Bacan Formations. Similarity in the joint directions between the intrusive and the host rock extends throughout the studied area.

Host Material
The NBM can clearly be observed to intrude the Bacan and South Bacan Formations at the upper parts of S. Kaputusan and at S. Rain, respectively. An angular feldspar-phyric South Bacan Formation xenolith is present within quartz-monzodiorite float from S. Mau (Plate 9.1c). Nowhere is there an intrusive contact with the Ruta Formation.

Lateral Extent and Variations
Along S. Kaputusan, finer grained quartz-monzodiorite cuts through the Bacan Formation as sills and dykes, typically with thicknesses of about 0.10 m. Similar relationships were observed along
Plate 9.1a. Coarse NBM with subparallel alignment of amphibole, exposed at P. Nusa Babi.

Plate 9.1b. NBM cut by 'aplite' dyke exposed at the type locality.

Intrusive Rocks

S. Bibinoi, S. Rain and S. Mau, where the intrusive rocks form small plugs (cf. Silitonga et al., 1981; Pudjowalujo & Bering, 1982) cutting through the South Bacan Formation. Along S. Rain and S. Mau, monzogranite dykes and veins cut through the quartz-monzodiorite which intrudes the South Bacan Formation. Float samples of fine grained quartz-monzodiorites were collected from S. Wayakuba, S. Wayamoa and S. Gofu all on south Bacan; S. Geti and along the track from Babang to Yaba, subparallel to S. Sayoang, both on north Bacan; and at P. Nusa Deket, north of Nusa Babi. Pudjowalujo & Bering (1982) and Bering et al. (1985) referred to the finer grained intrusive rocks as tonalite porphyry. Fig.9.1 shows the distribution of the NBM in the Bacan region.

Disseminated pyrite and galena mineralisation occurs at the contacts with the host formations, seen most clearly along S. Bibinoi and S. Kaputusan. The quartz-monzodiorite is locally altered, indicated by the presence of float containing secondary epidote and biotite replacing hornblende, as collected at S. Nyonyifi.

9.2.3 Petrography and Mineral Chemistry

9.2.3.1 Petrography

The monzodiorite consists of interlocking subhedral crystals (~1.5 mm), dominated by plagioclase (55%), green primary amphibole (15%), quartz (10%), biotite (10%), alkali feldspar (5%), opaques (5%) and both ortho- and clinopyroxene (< 5%) (Plate 9.2a,b). Zircon is present as an accessory phase. Plagioclase is zoned and twinned and contains glass inclusions in crystal cores. Quartz occurs in both monocrystalline and polycrystalline forms. Amphiboles are subhedral with well developed cleavage and locally are intergrown with pyroxene. Some samples show alteration of pyroxene to epidote. Platy biotite occurs as a probable alteration product of amphibole. The rock is classified as an amphibole-bearing quartz-monzodiorite (LeMaitre, 1989).

The finer grained (mode ~0.5 mm) monzogranite ('aplite') dykes consist of quartz (45%), plagioclase (40%), alkali feldspar (5%), hornblende (5%), biotite (5%) and minor opaque grains, with epidote and chlorite present as secondary minerals. The feldspars do not show twinning (Plate 9.2c,d).

9.2.3.2 Mineral Chemistry

Seven samples were selected for mineral chemistry analysis (BR79, BR80, BR267, BM260, BM285, BM460, and BM527). Samples BR79 and BM260 are monzogranite, whereas the others are quartz-monzodiorite. Fig.9.2 shows the locations of samples used for analytical work.
Figure 9.1. Distribution of intrusive rocks in the Bacan region.
Plate 9.2a-b. PPL and XPL of NBM quartz-monzodiorite (BR80); pyroxene intergrown with amphibole in a plagioclase and quartz matrix.

Plate 9.2c-d. PPL and XPL of 'aplite' monzogranite (BM399); biotite and amphibole in a quartz and plagioclase matrix.

Scale for all photomicrographs are 0.5 mm.
Figure 9.2 Location of samples from the Nusa Babi and Saleh Intrusives used in analytical work.
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The results are discussed in descending order of mineral modal abundance.

**Plagioclase**

Subhedral-anhedral plagioclase is ubiquitous in both monzodiorite and monzogranite dykes. In the monzodiorite, the plagioclase compositional range is $\text{An}_{49.69} \text{Ab}_{31.74} \text{Or}_{0.6}$ (mean $\text{An}_{54} \text{Ab}_{5} \text{Or}_{1}$; $n=21$), with the majority being $\text{An}_{54.69}$. In the monzogranite, compositions are $\text{An}_{35.62} \text{Ab}_{38.6} \text{Or}_{8.7}$ (mean $\text{An}_{42} \text{Ab}_{2} \text{Or}_{8.3}$; $n=17$), with the majority $\text{An}_{47.59}$. Oscillatory zoning is present (e.g. BM260P3-14, core-rim), with the rims being more sodic. Fig. 9.3a shows the plagioclase compositional variation, where monzogranite plagioclase compositions appear slightly less calcic than those in the monzodiorite, consistent with the derivation of the aplite dykes from residual differentiated magma. The spectrum of plagioclase composition suggests varying degrees of albitisation.

**Alkali feldspar**

Subhedral, K-rich alkali feldspar occurs in both monzodiorite ($\text{An}_{9.12} \text{Ab}_{18.20} \text{Or}_{0.92}$; $n=3$) and monzogranite ($\text{An}_{9.20} \text{Ab}_{8.20} \text{Or}_{80.92}$; $n=3$).

**Amphibole**

Green, anhedral primary amphibole is ubiquitous in the monzodiorite groundmass, locally intergrown with pyroxene. These are magnesian hornblende ($n=15$), with characteristics of orogenic andesites (Gill, 1981) which are: [1] $39-49$ wt% SiO$_2$; [2] $\text{Fe}_2\text{O}_3/\text{FeO} > 0.5$; [3] $\text{FeO}*/\text{MgO} 0.5-2.0$ and [4] Na+K correlating well with Al$^iv$ (Fig. 9.3b). Such amphiboles suggest a magma of $>3$ wt% $\text{H}_2\text{O}$, and the $< 0.5$ Al$^iv$ content suggests $\text{P}_\text{H}_2\text{O} < 9$ kb (Gill, op cit.).

**Micas**

Platy biotite occurs in BR80, BR267 and BM285 as part of the groundmass. All analyses ($n=13$) plot similarly in Al-Mg-Fe$^{2+}$ space, with the BR267 analyses being more aluminous. Biotite in the monzogranite (BR79) plots in the same region on this diagram. Muscovite occurs in BR267, as part of the groundmass ($n=3$). Compared to similar plutonic rocks from other parts of the world, NBM biotites have lower Fe and higher Mg compared to a granitic biotite from Southern California (Deer et al., 1992), whereas muscovites have lower Al and higher Fe compared to muscovite from a pegmatite in New Mexico (Deer et al., 1992), possibly indicating a lower temperature origin.

**Quartz**

Quartz is the dominant constituent of the groundmass in all samples. It is distinguished from
orthoclase by its undulose extinction.

*Fe-Ti Oxides*

There are three types of oxide present: [1] magnetite (no TiO$_2$), [2] titanomagnetite (TiO$_2$=10-18 wt%) and [3] high-Ti titanomagnetite (TiO$_2>$50 wt%). Magnetite (n=13) is found in both monzodiorite (BR80, BM285 and BM527) and monzogranite (BR79 and BM260), and titanomagnetite (n=6) in monzodiorite (BM285 and BM460). Both occur as subhedral-anhedral groundmass crystals and as inclusions in hornblende (BM285I1-2) and plagioclase (BM285I3-4), with no compositional difference. High-Ti titanomagnetite is present in BM285 (monzodiorite) and BM260 (monzogranite), closely associated with quartz.

*Pyroxene*

Orthopyroxene present in BR80 and BM285 has a composition En$_{52.58}$Fs$_{40.44}$Wo$_{2.6}$ (mean En$_{55}$Fs$_{42}$Wo$_{1.4}$; n=6).

*Apatite and Zircon*

Small euhedral grains of apatite occur as inclusions in amphibole (e.g. BM285) and euhedral zircon is occasionally found as inclusions in biotite (e.g. BR80).

*Chlorite*

Chlorite in monzodiorite replaces either biotite (e.g. BM285) or amphibole (e.g. BM460 and BM260). All chlorites are the high-Si variety (SiO$_2>$25 wt%; n=11), although those replacing amphibole are higher in SiO$_2$, FeO, MnO and lower in Al$_2$O$_3$, TiO$_2$, MgO (diabantites; after Hey, 1954) compared to those replacing biotite (pycnochlorite), which are similar to those in the monzogranite.

*Epidote*

Epidote occurs in BR267 and BM260 as anhedral grains or in radial clusters, replacing pyroxene. All analyses (n=8) are compositionally similar to the epidotes from low-grade metavolcanic rocks quoted by Coombs *et al.* (1976) and Offler & Aguirre (1984), differing from pegmatitic epidote (e.g. Deer *et al.*, 1992) in having slightly higher Al$_2$O$_3$ and lower Fe$_2$O$_3$ contents.

*9.2.3.3 Implications*

Based on the presence of inclusions, the order of crystallisation is Fe-Ti oxides, plagioclase, pyroxene (replaced by epidote), amphibole, biotite (replaced by chlorite), alkali feldspars and
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Figure 9.3a. Ternary diagram showing all plagioclase compositions from the NBM.

Figure 9.3b. Covariation of Na+K and Al" from the amphiboles (after Searle & Malpas, 1982) in the NBM.
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quartz. This order of crystallisation suggests a fractional crystallisation process, with the mafic minerals crystallising first followed progressively by more silicic minerals. Intergrown pyroxene and amphibole suggests that the magma evolved to a more acidic composition with the introduction of water. Locally there are rocks affected by alteration, with a secondary mineral assemblage of albite (in monzogranite) + chlorite + epidote, indicating low-temperature alteration in a hydrous environment, possibly due to auto-metasomatism during cooling of the intrusive body.

9.2.4 Whole Rock Chemistry

Five NBM samples (BR80, BR267, BM285, BM460 and BM527) were analysed for bulk chemistry, using the XRF technique, in an attempt to decipher the tectonic setting and also to compare them with extrusive rocks of similar age of the South Bacan Formation. Only BR80 and BM285 are in situ, the others are float samples. All analysed samples are quartz-monzodiorite.

9.2.4.1 Major Elements

Fig.9.4a can be used to assess the degree of alteration of the analysed NBM samples. In this diagram it is clear that the Y contents of NBM samples, except BR267, correlate well with Zr. The Rb and K₂O contents of BM527 do not correlate well with the Zr contents. These two samples are therefore interpreted to be affected by alteration.

Figs.9.4b-g are Harker-type diagrams, showing the covariation of SiO₂ with K₂O, Al₂O₃, CaO, Na₂O, TiO₂ and FeO*/MgO. All of these, except the FeO*/MgO ratio, correlate reasonably well with SiO₂, suggesting a cogenetic origin, related by crystal fractionation. The FeO*/MgO ratio variation is attributed to small variations in modal abundance of Fe-Ti oxides and possibly also due varying degree of alteration, controlled in these rocks by the amount of amphibole and biotite.

9.2.4.2 Trace Elements

Table 9.1 compares trace elements from the NBM samples with those of the South Bacan Formation. All NBM trace element contents closely resemble each other, except Sr which may be affected by alteration. The element ratios are also similar with the exception of Zr/Rb and Y/Nb, which can be explained by the anomalously low Rb content of BM527 and low Y content of BR267.

In comparison to the South Bacan Formation, most of the bulk incompatible trace element contents are lower, although the element ratios are comparable particularly with non-altered NBM. This is attributed to differences in the major elements between NBM (SiO₂=56.7-67.6 wt%; mean=62.0;
Figure 9.4a. Covariations of bulk alkali and LIL contents with HFS (Zr) to determine the degree of alteration in the NBM. Y serves to establish whether the samples are cogenetic.

Figures 9.4b-g. Harker-type diagram showing covariations of SiO₂ against K₂O, Al₂O₃, CaO, Na₂O, TiO₂, and FeO*·MgO in the NBM and Saleh Diorite.
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n=5) and the South Bacan Formation (SiO₂=50.1-61.4 wt%; mean=54.7; n=3), whereby the more acidic, thus more differentiated rocks such as the NBM, have greater contents of incompatible trace elements compared to the less differentiated volcanic rocks such as the South Bacan Formation.

Table 9.1. Comparison of selected NBM trace elements and their ratios with those of the South Bacan Formation. Values in parentheses are the ranges. Nusa Babi Intrusive consists of BR80, BM285 and BM460. Nusa Babi altered are BR267 and BM527.

<table>
<thead>
<tr>
<th></th>
<th>Nusa Babi Intrusive</th>
<th>Nusa Babi Intrusive altered</th>
<th>South Bacan Fm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sr</td>
<td>373 (345.7 - 389.2)</td>
<td>503 (409 - 597.3)</td>
<td>451 (245.6 - 600.9)</td>
</tr>
<tr>
<td>Rb</td>
<td>27.2 (11.2 - 40.3)</td>
<td>27.6 (2.1 - 53.1)</td>
<td>19.4 (5.5 - 33.1)</td>
</tr>
<tr>
<td>Zr</td>
<td>103 (58.8 - 155.4)</td>
<td>122 (89.9 - 154.5)</td>
<td>55 (40.1 - 62.0)</td>
</tr>
<tr>
<td>Nb</td>
<td>2.4 (1.3 - 4)</td>
<td>3.4 (3.1 - 3.6)</td>
<td>1.4 (1.1 - 1.6)</td>
</tr>
<tr>
<td>Y</td>
<td>27.6 (24.1 - 31.7)</td>
<td>24.7 (14.1 - 35.2)</td>
<td>19.2 (12.3 - 26.0)</td>
</tr>
<tr>
<td>La</td>
<td>10.4 (6.9 - 15.8)</td>
<td>12.9 (8.8 - 16.9)</td>
<td>5.7 (3.9 - 8.7)</td>
</tr>
<tr>
<td>Nd</td>
<td>15.8 (10.3 - 22.5)</td>
<td>16.0 (15.4 - 16.5)</td>
<td>10.3 (9.3 - 11.8)</td>
</tr>
<tr>
<td>Zr/Rb</td>
<td>4.1 (3.2 - 5.3)</td>
<td>37.6 (1.7 - 73.6)</td>
<td>4.7 (1.2 - 10.0)</td>
</tr>
<tr>
<td>Zr/Nb</td>
<td>44.1 (38.9 - 48.2)</td>
<td>37.4 (25 - 49.8)</td>
<td>40.7 (34.5 - 47.4)</td>
</tr>
<tr>
<td>Zr/Y</td>
<td>2.7 (1.5 - 4.1)</td>
<td>3.2 (2.4 - 4.1)</td>
<td>3.1 (2.3 - 5.0)</td>
</tr>
<tr>
<td>Y/Nb</td>
<td>13.3 (7.9 - 18.5)</td>
<td>7.6 (3.9 - 11.4)</td>
<td>14.5 (8.8 - 20.0)</td>
</tr>
<tr>
<td>La/Nb</td>
<td>4.5 (3.9 - 5.3)</td>
<td>3.8 (2.8 - 4.7)</td>
<td>4.2 (3.0 - 6.2)</td>
</tr>
<tr>
<td>Ba/La</td>
<td>26.3 (22.6 - 32)</td>
<td>20.7 (12.9 - 28.4)</td>
<td>37.2 (26.5 - 48.9)</td>
</tr>
</tbody>
</table>

9.2.4.3 Tectonic Discrimination Diagrams

Pearce et al. (1984) introduced empirical tectonic discrimination diagrams for granites based on analyses of over 600 samples from known tectonic settings. They divided all granites into [1] Ocean Ridge Granites (ORG), [2] Volcanic Arc Granites (VAG), [3] Within Plate Granites (WPG) and [4] Collision Granites (COLG), subdivided into syn-collision (SYNCOLG) and post-collision granites (POSTCOLG). It is, however, doubtful whether VAG can be distinguished from COLG, particularly if they formed immediately before or after collision.

When normalised against ORG (Pearce et al., 1984), the NBM samples show enrichment in K,
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Rb, Ba and Th and depletion in Nb and Y, all of which are characteristic of VAG and COLG (Fig.9.5a). The normalised Rb values of ~10, however, make the Nusa Babi rocks more likely to be of VAG origin. To distinguish VAG from COLG, plots of Y against Nb, and Rb against Y + Nb are used (Figs.9.5b,c). In the Y-Nb space, NBM plots in the field of VAG and SYNCOLG, while in the Rb - Y+Nb space, NBM plots in the VAG field.

There are limitations to this technique, particularly due to the effects of alteration (K, Rb, Ba are prone to alteration) and crystal accumulation (i.e. presence of aplite dykes). On all tectonic discrimination diagrams, non-altered and altered samples behave similarly, suggesting that these elements were not affected by alteration. The evidence from NBM is sufficient to rule out ORG and WPG as modes of genesis.

9.2.4.4 Implications

Whole rock chemistry suggests that the NBM is of plutonic arc or collisional magmatic origin, with the igneous diversity controlled by crystal fractionation. This intrusion is therefore interpreted as pre-collisional arc or post-collision magmatism, related to the arrival of continental crust in Bacan. These possibilities are discussed in more detail in Section 9.4.

9.2.5 Age Determination

K-Ar radiometric dating of a biotite separate from BR80 yields an Early Miocene age (19.80 ± 1.60 Ma). Isochron ages for BR80 and BM285 are 19.6 and 20.7 Ma (Fig.5.9a,b). The oldest date from BR80 (21.4 Ma) and both of the isochron ages suggests that this intrusion was formed after the collision of Australia and Philippine Sea Plate (dated by the regional unconformity at ~22 Ma). Dating of a hornblende separate from BM285 and of a biotite separate from BR267 yield Pliocene ages (< 6.155 Ma and 2.24 ± 0.89 Ma, respectively), which probably signify resetting due to the Quaternary volcanism in the vicinity of both samples.

Table 9.2 Summary of the K-Ar results from the Nusa Babi Monzodiorite.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Material</th>
<th>Grain size (µ)</th>
<th>%K (1σ error)</th>
<th>Wt for Ar (g)</th>
<th>40Ar* (nl/g, 1σ error)</th>
<th>40Ar/ 39Ar* (%)</th>
<th>Age (Ma, 2σ error)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BR80</td>
<td>hb + bt</td>
<td>125-250</td>
<td>1.061 ± 8.20%</td>
<td>0.4898</td>
<td>1.1895 ± 14.54%</td>
<td>92.96</td>
<td>28.6 ± 9.5</td>
</tr>
<tr>
<td>BR80dp</td>
<td>bt</td>
<td>125-250</td>
<td>4.291 ± 1.00%</td>
<td>0.1024</td>
<td>3.3251 ± 3.87%</td>
<td>68.74</td>
<td>19.8 ± 1.6</td>
</tr>
<tr>
<td>BM285</td>
<td>hb</td>
<td>125-250</td>
<td>0.407 ± 5.40%</td>
<td>1.1569</td>
<td>0.1020 ± 1159.23%</td>
<td>98.97</td>
<td>&lt; 6.16</td>
</tr>
<tr>
<td>BR267</td>
<td>bt</td>
<td>500-1000</td>
<td>6.690 ± 1.00%</td>
<td>0.1080</td>
<td>0.5861 ± 19.76%</td>
<td>94.99</td>
<td>2.2 ± 0.9</td>
</tr>
</tbody>
</table>
Figure 9.5a. Spider diagram for the NBM normalised against ORG (after Pearce et al., 1984).

Figure 9.5b. Covariation of Y and Nb for granitic tectonic discriminator applied to the NBM (after Pearce et al., 1984).

Figure 9.5c. Covariation of Y+Nb against Rb for granitic tectonic discriminator applied to the NBM (after Pearce et al., 1984).
9.3 THE SALEH DIORITE

9.3.1 Lithofacies

In the southern part of Saleh Kecil, the Saleh Metamorphic Complex is intruded by an amphibole-rich diorite named the Saleh Diorite. Cutting through the diorite are alternating epidote and amphibole lenses up to 0.25 m thick, possibly representing multiple stage intrusions with related mineral segregation processes. This intrusion has a similar texture to the hornblendites of the Sibela Complex which have cumulate textures.

Fine grained dioritic material, mineralogically similar to the Saleh Diorite but texturally different from the Sibela cumulates, was found as subrounded clasts (up to 0.25 m across) within recent, poorly consolidated coastal conglomerate around the northeastern coast of the Sibela Mountains. There are also microdiorite dykes, containing mainly amphibole and plagioclase, intruding the South Bacan Formation, along S. Wayakuba. Similar microdiorite are juxtaposed, as a fault bounded segment, against the the Sibela Complex at the southwestern corner of the Sibela Mountains. The exposure at southwestern Sibela Mountains consists of two subparallel, NW-SE trending, steep ridges with parallel drainage pattern perpendicular to the ridge top. These differ, aerial photographically, from the Sibela Continental Suite in having a lower elevation and a rougher surface. These intrusions are affected by hydrothermal veins consisting of epidote, malachite associated with quartz, calcite and minor metallic minerals.

9.3.2 Petrography and Mineral Chemistry

9.3.2.1 Petrography

The Saleh Diorites contain euhedral-subhedral green amphibole (65%), plagioclase (35%) and rare orthopyroxene and clinopyroxene. The amphiboles are euhedral and up to 25 mm across, with interstitial microcrystalline feldspar. Quartz is rarely present locally. Some of the amphiboles contain opaque inclusions (titanomagnetite and pyrite) and locally are replaced by epidote and chlorite. Plagioclase is locally replaced by zeolite, epidote and muscovite. Some samples have a strong igneous fabric, with the finer groundmass showing a poikilitic texture (Plate 9.3a,b).

The microdiorites have an intergranular texture, consisting of interlocking plagioclase, amphiboles, relict pyroxene, opaque and minor apatite (Plate 9.3c,d). Interstitial quartz is present. Amphibole is intergrown with pyroxene. Chlorite replaces amphibole. Zeolite and prehnite are present in veins.
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Plate 9.3a,b. PPL and XPL of Saleh Diorite (SR6).

Plate 9.3c,d. PPL and XPL of Saleh Microdiorite (B66).

Scale for all photomicrographs are 0.5 mm.
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9.3.2.2 Mineral Chemistry
Five samples (SR2, SR6, BR43, B66 and BM198) were analysed for mineral chemistry. SR6 was collected as an intrusion within SR2, and BR43 was a clast in the Sibela coastal conglomerate. B66 and BM198 are microdiorites intruding the Sibela Complex and the South Bacan Formation, respectively. Fig.9.1 shows the location of these samples.

Amphiboles
Amphibole occurs in all samples: in SR2 as magnesiohomblende (n=4), actinolitic hornblende (n=5), and minor paragasic hornblende (n=1); in SR6 as tschermakitic hornblende (n=4) and actinolite (n=1); in BR43 as magnesio-hastingsite and tschermakitic hornblende and in B66 and BM198 as green tremolite-actinolite to tremolitic-actinolitic hornblende (n=11). To distinguish metamorphic from igneous origin for these amphiboles, an Na+K against Al\textsuperscript{vi} diagram (Searle & Malpas, 1982) is used. Fig.9.6a shows that all amphiboles plot along the high temperature trend, which is supportive of a primary intrusive origin. These differ chemically from amphiboles in the similarly textured Sibela hornblendite (BM571 and BM572) in having lower stoichiometric Al\textsuperscript{vi} (<0.5).

Using the geothermometry calculation method of Blundy & Holland (1990), the Saleh Diorites amphiboles yield temperatures of 1004-1022°C at 1 kb, for BR43 and SR6, consistent with an intrusive origin. SR2 yielded a temperature of ~713°C, B66 ~598°C and BM198 ~517°C, all possibly lowered due to contact metamorphism with younger intrusions (e.g. SR2 intruded by SR6) or auto-metamorphism. It is, however, noted that compared to other geothermometers, the Blundy & Holland method appears to yield higher temperature estimations (cf. Sections 4.5.1.2 and 5.5.1.2).

Feldspars
Plagioclase occurs in the groundmass of all samples and in BR43 and BM198 as a phenocryst phase. All show simple twinning, although zoning is absent. There are two compositional groups of plagioclases, most are very calcic with a compositional range of An\textsubscript{14.9}Ab\textsubscript{3.2}Or\textsubscript{5} (mean An\textsubscript{04}Ab\textsubscript{4}Or\textsubscript{5}; n=24) and variations up to 12 mol% within a sample. The highly calcic nature of the bulk of the plagioclase is typical of orogenic rocks (Gill, 1981). Two analyses of plagioclase inclusions in amphibole are similar to the groundmass phase (An\textsubscript{81.3}Ab\textsubscript{15.3}Or\textsubscript{3.9}). Some analyses of phenocrysts in SR2, BR43, B66 and BM198 yield more sodic results (An\textsubscript{5.6}Ab\textsubscript{42.9}Or\textsubscript{5.12}; mean An\textsubscript{11}Ab\textsubscript{2}Or\textsubscript{5}; n=14), with variations up to 20 mol% in a sample. These sodic plagioclase are attributed to secondary albitisation during cooling in a hydrous environment. Four analyses from
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BR43 and B66 are orthoclase $\text{An}_{0.2}\text{Ab}_{2.14}\text{Or}_{84.98}$ (mean $\text{An}_{1}\text{Ab}_{7}\text{Or}_{92}$), interpreted as a primary K-bearing phase. Fig.9.6b shows the results of feldspar analyses.

Opaque Minerals

There are four types of opaque minerals observed: pyrite in SR2 and B66; possibly chalcopyrite in SR2; magnetite in SR6 and B66; and titanomagnetite in SR6 and BM198 (Fig.9.6c).

Pyroxene

Orthopyroxene ($\text{En}_{68}\text{Fs}_{30}\text{Wo}_{2}$) occurs in both radial and non-radial habit in SR6, with no compositional differences detected between the two. Clinopyroxene ($\text{En}_{42}\text{Fs}_{3}\text{Wo}_{45}$; $n=5$) occurs as a phenocryst in BM198 and as a rare accessory phase in BR43.

Phyllosilicates

Fine-grained muscovite ($n=5$) is present in SR2 and SR6 replacing plagioclase. Chlorite occurs in SR6 and B66 as a breakdown product of amphibole. Chlorites in SR6 are of similar composition to the high-Si chlorite of Hakim (1989), whereas the ones in B66 are of low-Si variety, both of which are typical of low-grade metamorphism.

Zeolite

Zeolite occurs in SR2 and BM198 as a low birefringence, platy mineral, forming veins and replacing plagioclase. The ones in SR2 are compositionally similar to errionite, whereas those in BM198 are of natrolite composition (Gottardi & Galli, 1985).

Other Minerals

Apatite is present in BR43 and B66 as microphenocrysts. Prehnite occurs in BR43 and BM198 replacing plagioclase and forming veins, typical of hydrothermal alteration. Epidote is present in SR2 and SR6, although was not analysed.

9.3.2.3 Implications

Based on the presence of inclusions, the order of crystallisation was Fe-Ti oxide, plagioclase, pyroxene and amphibole. Low temperature secondary minerals are present (e.g. albite, muscovite, chlorite and epidote), and are interpreted as the products of late stage hydrothermal alteration associated with cooling of the magma (auto-metamorphism). Prehnite and zeolites were produced during later stage alteration, possibly by hydrothermal fluid circulating in fractures, and may be related to Neogene volcanism.
Figure 9.6a. Covariation of Na+K and Al" from the amphiboles (after Searle & Malpas, 1982) in the Saleh Diorite.
Figure 9.6b. Ternary diagram showing plagioclase compositions in the Saleh Diorite.
Figure 9.6c. Ternary diagram showing Fe-Ti oxides compositions from the Saleh Diorite.
9.3.3 Whole Rock Chemistry

Three samples (SR2, SR6 and B66) were analysed, using XRF, to characterise the bulk chemical character and constrain the tectonic setting of the Saleh Diorite. SR2 and SR6 are diorites, whereas B66 is a microdiorite, all of which were collected in situ.

9.3.3.1 Major Elements

All samples have <1.5 wt% LOI, and therefore are considered fresh. Figs.9.4b-g are Harker diagrams which indicate that the Saleh Diorite is different to the NBM. Post-orogenic magmatic rocks often have a wide range of compositions, such as the appInitie intrusions of the Scotland Caledonides, attributed to gravitational differentiation and assimilation with previous magmatic bodies (e.g. French, 1976). It is therefore possible that the Saleh Diorite may be related to the NBM.

9.3.3.2 Trace Elements

Trace element geochemistry is used to establish the relationship between the Saleh Diorite and the NBM and Sibela hornblendites.

The higher LIL, HFS and REE contents and their ratios (particularly Y/Nb) clearly distinguish the Saleh Diorite from the Sibela hornblendites. In contrast, the Saleh Diorite has lower incompatible trace element contents compared to the NBM which may be due to its more basic composition (less differentiated).

9.3.3.3 Tectonic Discrimination Diagrams

Tectonic discrimination diagrams of Pearce et al. (1984) were produced for plutonic rocks with >5 modal % quartz (see Section 9.2.4.3). Although the Saleh Diorite is more basic (47-54 wt% SiO₂) than the samples used by Pearce et al. (1984) (~70 wt% SiO₂), the Saleh rocks locally contain ~5% quartz and therefore the diagrams may useful. On the ORG normalised diagram of Pearce et al. (1984), the Saleh Diorite shows a 'humped' enrichment of K₂O-Nb, a strong Nb depletion, concave-up Nb-Zr distribution and Y enrichment relative to Zr (Fig.9.7a), which closely resembles VAG pattern. In the Y-Nb space the Saleh Diorite plots in the field of VAG and SYNCOLG (Fig.9.7b), whereas in Y+Nb-Rb space it plots in the VAG field (Fig.9.7c). Differences between VAG and SYNCOLG may not be apparent for the reasons stated above (see Section 9.2.4.3).
Figure 9.7a. Spider diagram for the Saleh Diorite normalised against ORG (after Pearce et al., 1984).

Figure 9.7b. Covariation of Y and Nb for granitic tectonic discriminator applied to the Saleh Diorite (after Pearce et al., 1984).

Figure 9.7c. Covariation of Y+Nb against Rb for granitic tectonic discriminator applied to the Saleh Diorite (after Pearce et al., 1984).
Intrusive Rocks

Table 9.3 Comparison of the Saleh Diorite trace element contents and selected ratios with average NBM and Sibela hornblendites. Values in parentheses are the ranges. Saleh Microdiorite is B66. Hornblendites are BM571 and BM572. NBM consists of BR80, BM285 and BM460; for ranges refer to Table 9.1.

<table>
<thead>
<tr>
<th></th>
<th>Saleh Diorite</th>
<th>Saleh Microdiorite</th>
<th>Hornblendites</th>
<th>NBM</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sr</td>
<td>155 (95-215)</td>
<td>246</td>
<td>89 (71-108)</td>
<td>373</td>
</tr>
<tr>
<td>Rb</td>
<td>9.6 (4.4-14.7)</td>
<td>11.8</td>
<td>1.6 (1.5-1.6)</td>
<td>27.2</td>
</tr>
<tr>
<td>Zr</td>
<td>26.9 (20.3-33.4)</td>
<td>61.6</td>
<td>8.5 (6.6-10.3)</td>
<td>103.5</td>
</tr>
<tr>
<td>Nb</td>
<td>1.1 (0.8-1.4)</td>
<td>1.3</td>
<td>0.4 (0.3 - 0.5)</td>
<td>2.4</td>
</tr>
<tr>
<td>Y</td>
<td>18.2 (14.2-22.1)</td>
<td>26</td>
<td>10.4 (7.2-13.6)</td>
<td>27.6</td>
</tr>
<tr>
<td>La</td>
<td>2.6 (2.3-2.8)</td>
<td>3.9</td>
<td>-</td>
<td>10.4</td>
</tr>
<tr>
<td>Nd</td>
<td>5.9 (5.1-6.6)</td>
<td>9.3</td>
<td>1.8 (1.2-2.3)</td>
<td>15.8</td>
</tr>
<tr>
<td>Zr/Rb</td>
<td>3.4 (2.3-4.6)</td>
<td>5.2</td>
<td>5.4 (4.4-6.4)</td>
<td>4.1</td>
</tr>
<tr>
<td>Zr/Nb</td>
<td>24.6 (23.9-25.3)</td>
<td>47.4</td>
<td>21.3 (20.6-22)</td>
<td>44.1</td>
</tr>
<tr>
<td>Zr/Y</td>
<td>0.7 (0.5-0.9)</td>
<td>2.4</td>
<td>0.8 (0.8-0.9)</td>
<td>2.7</td>
</tr>
<tr>
<td>Y/Nb</td>
<td>16.8 (15.8-17.8)</td>
<td>20</td>
<td>25.6 (24-27.2)</td>
<td>13.3</td>
</tr>
<tr>
<td>La/Nb</td>
<td>2.4 (2-2.9)</td>
<td>3</td>
<td>-</td>
<td>4.5</td>
</tr>
<tr>
<td>Ba/La</td>
<td>26.4 (23.4-29.5)</td>
<td>26.5</td>
<td>-</td>
<td>26.3</td>
</tr>
</tbody>
</table>

Although these are intrusive rocks, their compositions would classify them as basalt and basaltic andesite (Le Bas & Streckeisen, 1991) had they been erupted as volcanic rocks. Tectonic discrimination diagrams for basaltic rocks are therefore applied to these rocks to indicate the likely tectonic setting for the magma. On the Ti-Zr-Y diagram of Pearce & Cann (1983), the SR2 and SR6 plot outside the Low-K Tholeiite (LKT) field, whereas B66 plots in the overlap field of LKT, Ocean Floor Basalt (OFB) and Calc-Alkaline Basalt (CAB). This diagram is therefore inconclusive (Fig.9.8a). On the Nb-Zr-Y diagram of Meschede (1986) all Saleh Diorite samples plot in the N-MORB and Volcanic Arc Basalt (VAB) overlap field (Fig.9.8b). On the TiO₂-MnO-P₂O₅ diagram of Mullen (1983), SR2 plots in the CAB field, while SR6 and B66 plot in the Island Arc Tholeiite field (Fig.9.8c).

When normalised against MORB (Pearce, 1982), the Saleh Diorites show a humped Sr-Th distribution; a strong Nb and Cr depletions, typical of volcanic arc basalts (Fig.9.9a). When normalised against chondrite (Sun, 1980), the Saleh Diorites have a spiky distribution with
pronounced depletions of Nb and Ce; and enrichments of K and Sr, similar to arc basalts (Fig.9.9b).

9.3.3.4 Implications
Trace element chemistry shows that the Saleh Diorite is different to either the Nusa Babi Monzodiorite or the Sibela hornblendites. The Saleh Diorite could be related to the NBM, much like the appinitic rocks of Scotland, however in Harker-type diagrams, their major elements do not correlate well possibly indicating different magma genesis. Tectonic discrimination diagrams suggest that the intrusive body is related either to a volcanic arc or post-collisional magmatism.

9.3.4 Age Determination
K-Ar results (Table 9.4) from Saleh Diorite hornblende separates yielded ages of 9.29 ± 1.58 Ma (SR6), 10.80 ± 8.70 Ma (SR2) and 15.1 ± 1.6 Ma (B66). The ~15 Ma age is interpreted as the age of intrusion in southwest Sibela Mountains, although there is nothing to rule out the possibility that this age may also be reset from an older age, particularly as there are some reset ages at ~15 Ma in the Bacan region (see Sections 4.7).

The ~11 Ma age is unreliable because of its large error, attributed to partial loss of $^{40}$Ar* (hence lower $^{40}$Ar*, higher $^{40}$Ar$_{sm}$) due to intrusion of the SR2 body by SR6. The ~9 Ma age may indicate partial resetting of age or the age of intrusion in the Saleh Islands, both of which could be related with the initiation of the Halmahera arc.

Table 9.4 Summary of K-Ar results from the Saleh Diorite. SR samples are from Saleh Islands, B66 is from southwest Sibela Mountains.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Material</th>
<th>Grain size (μ)</th>
<th>%K (1σ error)</th>
<th>Wt for Ar (g)</th>
<th>$^{40}$Ar* (nl/g, 1σ error)</th>
<th>$^{40}$Ar$_{sm}$ (%)</th>
<th>Age (Ma, 2σ error)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SR2</td>
<td>amph</td>
<td>250-500</td>
<td>0.184 ± 1.00%</td>
<td>0.2027</td>
<td>0.0774 ± 40.24%</td>
<td>97.36</td>
<td>10.8 ± 8.7</td>
</tr>
<tr>
<td>SR6</td>
<td>amph</td>
<td>125-250</td>
<td>0.364 ± 1.00%</td>
<td>1.0611</td>
<td>0.1318 ± 8.49%</td>
<td>89.20</td>
<td>9.3 ± 1.6</td>
</tr>
<tr>
<td>B66</td>
<td>wr</td>
<td>125-250</td>
<td>0.656 ± 1.00%</td>
<td>1.1974</td>
<td>0.3865 ± 5.23%</td>
<td>83.42</td>
<td>15.1 ± 1.6</td>
</tr>
</tbody>
</table>
Figure 9.8a-c. Basaltic ternary discrimination diagrams applied to the Saleh Diorite. See text for details.
Figure 9.9a. Spider diagram for the Saleh Diorite normalised against MORB (after Pearce, 1982).

Figure 9.9b. Spider diagram for the Saleh Diorite normalised against chondrite (after Sun, 1980).
9.4 SYNTHESIS AND TECTONIC SIGNIFICANCE

Nusa Babi Monzodiorite

The similarities in petrography, mineral chemistry and whole rock chemistry indicate that most of the quartz-monzodiorites in the area are cogenetic. Field evidence such as 'aplite' dyke intrusion, widespread mineralisation and relationships with host rocks support this interpretation. Mineral and bulk chemistry data suggest that the NBM magma has been affected by crystal fractionation and the addition of water. While textural and mineralogical consistency are recognised, the grain size variation in the monzodiorite suggests that there are several levels of intrusion. The coarse grained amphibole-rich rocks probably represent the deeper level intrusion, with the finer grained leucocratic rocks as hypabyssal intrusions. The 'aplite' dykes represent residual magma, a product of differentiation, which has subsequently intruded the quartz-monzodiorite. Locally, NBM rocks are affected by low-temperature alteration, possibly due to auto-metasomatism.

Tectonic discrimination diagrams suggest that NBM is of arc or collisional magmatic origin. However whole rock major and trace element chemistry indicates that there are considerable differences between the NBM and the similar age South Bacan Formation arc rocks. The fact that the NBM cuts through the folded South Bacan Formation indicates that NBM must have been younger than the South Bacan Formation and therefore the NBM is interpreted as a product of post-collisional magmatism. This is further supported by the K-Ar dates, indicating that the NBM was formed after collision of Australia and Philippine Sea Plate. The occurrence of NBM plugs intruding the South Bacan Formation is also consistent with this interpretation, as it would be unlikely that volcanic and plutonic equivalents would be juxtaposed at the same structural level. Post-collisional magmatism is also supported by the reported occurrence of the NBM intruding the Sibela Metamorphic Complex (Yasin, 1980), thus providing a latest constraining date for the arrival of the Sibela Metamorphic Complex in the Bacan region. The fact that the NBM is the most acidic intrusion in the Halmahera and Bacan regions may indicate the involvement of continental crust in the generation of the magma. This intrusive body is interpreted to be related to the arrival of continental crust in the region.

Saleh Diorite

Petrographical and mineral chemical studies show an order of crystallisation typical of arc-related diorite. Low temperature metamorphic effects are probably due to auto-metasomatism during cooling of the magma followed by hydrothermal circulation associated with brittle deformation. Trace
Intrusive Rocks

element geochemistry demonstrates that the Saleh Diorite is different from either the Nusa Babi Monzodiorite or the Sibela hornblendites. This intrusive body is associated with a volcanic arc or collisional magmatism. There are two possible interpretations for the genesis of the Saleh Diorite: [1] the Saleh Diorite represents post-collision magmatism related to the NBM, such as the appinites of Scotland or [2] the Saleh Diorite is a different magmatic body than the NBM and is associated with the initiation of the Halmahera arc. The first hypothesis implies that the ~15 Ma K-Ar age is a reset age, similar to the reset age of the Bacan Formation. The second possibility, supported by the whole rock geochemistry, indicates that the ~15 Ma age is formational age which is consistent with the presence of volcanic rocks from the Halmahera arc of similar age (~12 Ma) in Halmahera and Obi (Hall et al., 1992). The second interpretation is preferred here.
CHAPTER TEN
QUATERNARY DEPOSITS

10.1 INTRODUCTION

10.2 QUATERNARY VOLCANIC ROCKS
  10.2.1 Aerial Photography
  10.2.2 Lithofacies
  10.2.3 Petrography and Mineral Chemistry
    10.2.3.1 Petrography
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  10.2.4 Whole Rock Chemistry
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    10.2.4.3 Tectonic Discrimination Diagrams
  10.2.5 Synthesis and Tectonic Significance

10.3 QUATERNARY LIMESTONES

10.4 QUATERNARY DELTA

10.5 QUATERNARY SCREE

10.6 QUATERNARY ALLUVIUM AND FLUVIUM
10.1. INTRODUCTION
Five types of Quaternary deposits can be distinguished around Bacan: lavas, reef limestones, deltaic, scree, and alluvial deposits. None of these rock units has been considered in any detail in this work, but where appropriate they have been included in the mapping of the region (Fig. 10.1). This chapter deals with the lithology, petrography, mineral and whole rock chemistry of the Quaternary volcanic rocks. The lithologies of the other Quaternary deposits are also reported.

10.2 QUATERNARY VOLCANIC ROCKS
Quaternary volcanic rocks are exposed at Mts. Amasing, Suangi, Bibinoi and on the coast north of Wayatim Village. They are most accessible in the Mt. Suangi area, along a logging track from Babang to Yaba.

10.2.1 Aerial Photography
Aerial photography reveals that this unit is represented by smooth, steep and high volcanic cones with radial, deeply incised drainage patterns. Locally, lakes are formed at the caldera centres. The relatively steep conical shape of the volcanic centres is unusual for an extensive, mobile basaltic flow.

10.2.2 Lithofacies
The Quaternary volcanic rocks are grey-black and olivine, clinopyroxene and plagioclase-phyric, with grains up to 10 mm across (mode 0.3 mm). Both pyroxene and olivine are fresh and euhedral, locally forming glomerocrysts. The matrix is microcrystalline, cryptocrystalline or glassy. The rocks contain vesicles, flow banding and elongated parallel gas escape structures (Plate 10.1a), analogous to stretched vesicles, with fine-grained basaltic xenoliths in some samples. In the Mt. Bibinoi area, there is an exposure of poorly sorted, slightly graded volcanic breccia, possibly representing lahar.

The Quaternary exposures at Mts. Bibinoi, Amasing, Suangi and Wayatim lie unconformably upon the Bacan and South Bacan Formations.

10.2.3 Petrography and Mineral Chemistry
10.2.3.1 Petrography
The pyroxenes (15%) are euhedral, except when they form glomerocrysts, which normally
Figure 10.1 Distribution of Quaternary deposits in the Bacan region.
aggregate around large crystals. Zoning is common in large crystals. Plagioclase laths (15%) show complex zonation, twinning and contain glass inclusions. Olivines (15%) are mostly fresh and euhedral, locally showing conchoidal fracturing. In places they are replaced by red-brown 'iddingsite'. Oxides occur as euhedral microphenocrysts (5%) and there are embayed polycrystalline quartz xenocrysts. Some samples have been quenched and some have developed a trachytic fabric (Plate 10.1b,c).

10.2.3.2 Mineral Chemistry

Four samples were used for mineral chemistry study (BR256, BR268, BM202 and BM235; Fig.10.2). None of these samples were collected in situ, however they were all collected from angular blocks, ~2 m across, and were therefore near in situ. The aim of this study was to characterise the deposit and observe any chemical correlation with the Kaputusan Formation. The results are discussed in descending order of abundance.

**Plagioclase**

Plagioclase is an ubiquitous primary mineral occurring as phenocrysts, groundmass microlites and inclusions in olivine. Phenocryst compositions are An$_{62.84}$Ab$_{6.37}$Or$_{6.2}$ (mean An$_{75}$Ab$_{24}$Or$_{1.2}$; n=11), and the groundmass phase is An$_{44.7}$Ab$_{39.53}$Or$_{1.7}$ (mean An$_{60}$Ab$_{34}$Or$_{2}$; n=12). Only two analyses of inclusions are available, both from BM235, and these are An$_{50.6}$Ab$_{39.4}$Or$_{2}$. Compositional variation within a sample is mostly <10 mol%, except in BR268 (up to 20 mol%). In all samples, groundmass microlites are more sodic than the phenocrysts, suggestive of progressive fractionation. Fig.10.3a shows the variation of plagioclase composition.

**Pyroxenes**

Pyroxene is a common phenocryst phase, mostly as clinopyroxene with minor orthopyroxene. The composition of the clinopyroxene is En$_{42.5}$Fs$_{6.18}$Wo$_{34.46}$ (mean En$_{47}$Fs$_{19}$Wo$_{34}$; n=19) and the orthopyroxene is En$_{73.4}$Fs$_{17.24}$Wo$_{2.1}$ (mean En$_{77}$Fs$_{30}$Wo$_{1}$; n=5). Zoning is observed in some samples (e.g. BR256PX5-7, core-rim), with the core more calcic than the rim. Phenocrysts are diopside, endiopside, augite and hypersthene, whereas groundmass grains are all augite (Fig.10.3b). On the Nisbet & Pearce (1977) diagram, the clinopyroxene plots in the VAB field (Fig.10.3c). Estimated crystallisation temperatures based on two pyroxene thermometry (Lindsley, 1983) are 700-1000°C.

**Olivine**

Olivine occurs as phenocrysts and as inclusions in pyroxene. It is fresh, euhedral-subhedral and of composition Fo$_{70.85}$ (mean Fo$_{77}$; n=20). Locally it has pleochroic altered margins of red-brown.
Plate 10.1a. Boulder of Quaternary volcanic rock at S. Bibnoi showing stretched vesicles.

Plate 10.1b-c. PPL and XPL of Quaternary volcanic rock (BR268). Scale = 0.5 mm.

Plate 10.1d. Typical outcrop of Quaternary deltaic deposit outside of Amasing Kota Village.

Plate 10.1e. Monzodiorite, limestone, volcanic and high-grade metamorphic clasts within the Quaternary deltaic deposit at Amasing Kota Village.
Figure 10.2 Location of samples from the Quaternary Volcanics used in analytical work.
Quaternary Deposits

hyalosiderite (e.g. BM202OL1-2). Zoning may be present (e.g. BR268OL2-4, core-rim), with Fe enrichment towards the rim.

**Spinels**

Fe-Ti oxides occur as square-sectioned microphenocrysts (titanomagnetite; n=16) and as inclusions in olivine (chrome spinel; n=3). Titanomagnetite plots between ulvöspinel and magnetite (Fig. 10.4a). On the covariation diagram of Cr# and TiO₂ wt%, the Cr-spinel plots in the VAB field (Fig. 10.4b), but on the Fe³⁺ versus TiO₂ wt%, it plots in the area which includes both intraplate and arc basalt (Fig. 10.4c; after Arai, 1992).

**Quartz**

Embayed monocrystalline and polycrystalline quartz xenocrysts are present in most samples. Locally they form glomerocrysts, up to 3 mm across.

**Glass**

Glass can be distinguished from plagioclase by its higher SiO₂ (>60 wt%), lower Al₂O₃ (<25 wt%) and erratic alkali contents. All glasses analysed were dark brown to opaque with no systematic chemical differences detected.

10.2.3.3 Implications

Based on the presence of inclusions, the order of crystallisation is Cr-spinel and Fe-Ti oxide, olivine, pyroxene and plagioclase. Plagioclase inclusions in olivine are interpreted as xenocrysts. Clinopyroxene and chromespinel tectonic discrimination diagrams suggest an arc related and/or possibly intra-plate origin for these basalt and the presence of quartz xenocrysts suggests that this unit was erupted through a quartz-rich basement (?Sibela continental fragment). The rocks differ from the Kaputusan Formation in lacking amphibole and biotite, and are also distinguished by having olivine phenocrysts and chromespinel inclusions.

10.2.4 Whole Rock Chemistry

The purpose of this study was to compare these rocks to the Kaputusan Formation. Four samples (BR256, BR268, BM202 and BM258) were analysed, using the XRF method, none of which were found in situ.

10.2.4.1 Major Elements

All samples analysed are fresh with <0.7 wt% LOI. Following Le Bas & Streckeisen (1991), these
Quaternary Deposits

Figure 10.3a. Plagioclase compositions from the Quaternary volcanic rocks.
Figure 10.3b. Pyroxene compositions from the Quaternary volcanic rocks.
Figure 10.3c. Tectonic discrimination diagram of Nisbet & Pearce (1977) applied to the Quaternary volcanic clinopyroxenes.
Figure 10.4a. Fe-Ti spinel compositions from the Quaternary volcanic rocks.
Figure 10.4b-c. Chrome spinel tectonic discriminants of Arai (1992) applied to the Quaternary volcanic rocks.

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Quaternary Deposits

samples are classified as calc-alkaline basaltic-andesite (Fig. 10.5a). In all samples, except BM202 which is distinct petrographically in having a coarser groundmass, CaO, Na₂O, Al₂O₃, Fe₂O₃, TiO₂, MgO and FeO*/MgO are well correlated with SiO₂ (Figs. 10.5b-h).

Hatherton & Dickinson (1969) used seismic and geochemical data from volcanoes in Indonesia, including the Halmahera Arc, to propose a correlation between depth to the top of the Benioff zone and K₂O content. Hutchison (1976) developed this concept and introduced an empirical equation relating depth of Benioff zone to K, Sr and Rb contents. Although these elements are all LIL and therefore prone to alteration, they can be used for the Quaternary volcanic rocks, due to their extreme freshness and young age. The average calculated depth to the top of the Benioff zone is [1] for K₂O 150 km (range 139-160 km); [2] for Rb 157 km (range 147-163 km); [3] for Sr 188 km (range 170-218 km). The error for each of these is 30 km therefore most results are within range of each other. The depth of the top of the Benioff zone is therefore interpreted as 150-180 km, which is comparable to results derived from the Halmahera Arc (Hatherton & Dickinson, 1969; Hutchison, 1976).

Trace Elements

A comparison of selected trace elements and ratios from the Quaternary basalts with rocks of known tectonic settings and the Kaputusan Formation is presented in Table 10.1. KFm I-IV denotes Kaputusan Formation Groups I-IV respectively.

The Quaternary Volcanic rocks’ trace elements and ratios are most similar to IACA. Comparison with the Kaputusan Formation shows that, although there are many differences, the Quaternary Volcanic rocks are most similar to Kaputusan Formation Group II, which may be attributed to similarity in petrography (basalt and TPAN). The Ba/La ratio (>3) is typical of Island Arc Basalt (Arculus & Powell, 1986) as is the La/Nb ratio (2-6; Saunders et al., 1980).

Tectonic Discrimination Diagrams

On the Ti-Zr-Y diagram of Pearce & Cann (1983), the Quaternary volcanic rocks plot in the OFB and CAB overlap field (Fig. 10.6a). On the Nb-Zr-Y diagram of Meschede (1986), they scatter in the fields of VAB & WPT and N-MORB & VAB (Fig. 10.6b). On the TiO₂-MnO-P₂O₅ diagram of Mullen (1983), they plot between the CAB and IAT fields (Fig. 10.6c).

When normalised against MORB (Pearce, 1982), the Quaternary volcanic rocks show a humped pattern for LIL (Sr-Th); a sharp depletion of Nb; Ce enrichment; a fairly flat, near MORB P to
Figure 10.5a. Quaternary volcanic rocks in the classification diagram of Le Bas & Streckeisen (1991).
Figures 10.5b-d. Covariations of SiO$_2$ against CaO, Na$_2$O and Al$_2$O$_3$. 
Figures 10.5e-h. Covariations of SiO₂ against Fe₂O₃, TiO₂, MgO and FeO*/MgO.
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Zr distribution; Ti depletion and a flat, near MORB Ti-Cr distribution. Although all samples have similar patterns, the BR samples (north Bacan) are enriched in most elements relative to the BM samples (south Bacan), and this may be attributed to the presence of continental crust under south Bacan. All patterns closely resemble VAB (Fig. 10.6d).

Table 10.1. Comparison of the average Quaternary volcanic rocks' trace elements and their ratios with IACA, Within Plate Alkaline (WPA) and average Kaputusan Formation. IACA and WPA data taken from Wilson (1989).

<table>
<thead>
<tr>
<th></th>
<th>QV</th>
<th>IACA</th>
<th>WPA</th>
<th>KFm I</th>
<th>KFm II</th>
<th>KFm III</th>
<th>KFm IV</th>
</tr>
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<td>Sr</td>
<td>569.5</td>
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<td>800</td>
<td>474.4</td>
<td>394.9</td>
<td>2202.5</td>
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<tr>
<td>Rb</td>
<td>29.5</td>
<td>14</td>
<td>22</td>
<td>128.2</td>
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<td>33.3</td>
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<td>Zr/Nb</td>
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<td>17.6</td>
<td>52.9</td>
<td>61.1</td>
<td>37.8</td>
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<tr>
<td>Zr/Y</td>
<td>4.2</td>
<td>2.7</td>
<td>7.3</td>
<td>7.1</td>
<td>3.8</td>
<td>9.1</td>
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<tr>
<td>La/Nb</td>
<td>4.5</td>
<td>7.1</td>
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<td>6.5</td>
<td>19.5</td>
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<tr>
<td>Ba/La</td>
<td>15.9</td>
<td>30</td>
<td>10.9</td>
<td>17.8</td>
<td>18.6</td>
<td>9.6</td>
<td>17.6</td>
</tr>
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</table>

On covariation diagrams of Ti against Zr; Cr against Y and Ce/Sr; Ti/Y against Nb/Y (Pearce, 1982) and Zr against Zr/Y (Pearce & Norry, 1979), the Quaternary volcanic rocks mostly plot in the VAB field (Figs. 10.7a-e).

10.2.5 Synthesis and Tectonic Significance

Field and petrographic evidence point to a rapidly cooled, mobile, low viscosity lava flow with associated possible laharc deposits. Mineral and whole rock chemistry reveal arc characteristics, with some indication of a within-plate component. The magma source may be related to the Kaputusan Formation (eastward subduction of the Molucca Sea Plate), consistent with the calculated depth to the top of the Benioff zone (150-180 km). The sample freshness and the smooth conical shape of the volcanoes indicate a very young age for these volcanic rocks, and
Figures 10.6a-c. Ternary tectonic discrimination diagrams applied to the Quaternary volcanic rocks.

Figure 10.6d. Spider diagram showing Quaternary volcanic rocks normalised against MORB.
Figures 10.7a-e. Covariations of trace elements used to decipher the tectonic setting of the Quaternary volcanic rocks.
Quaternary Deposits

imply a reactivation of volcanism after the cessation of that giving rise to the Kaputusan Formation in the Late Pliocene. The Quaternary volcanic rocks are therefore interpreted as a product of reactivated volcanism related to splaying of the Sorong Fault Zone. This is further supported by the linear distribution of Quaternary volcanic centres in Bacan, which coincides with major faults, and the absence of equivalent Quaternary volcanism between Bacan and Makian (southern-most active volcano in the Halmahera arc). This conclusion is in accordance with that of Silitonga et al. (1981). Froidevaux (1978) reported the presence of basalts in the Salawati and Misool region (Bird’s Head of Irian Jaya), related to movement along the Sorong Fault Zone. The presence of quartz xenocrysts and the suggestion of a WPT component on the discrimination diagrams may indicate the extent of continental basement under Bacan.

10.3 QUATERNARY LIMESTONES
Reef Limestones may be divided into two groups: uplifted masses of reef limestone presently exposed on land, and actively forming or very recently formed coral reefs which occur around the coastline. The appearance of the limestones is very distinctive on aerial photographs, having a flat and slightly karstic nature with a light colour shading. The limestones are well lithified and locally crystalline, cut by calcite veins and consisting of corals, up to 0.45 m across, shells and bivalve fragments.

The degree of uplift may be considerable, locally over 30 m. It has not been possible to date these limestones precisely within the Quaternary and so rates of uplift and tilting cannot be determined. Wave-cut platforms of dead coral are locally present around coasts. This indicates that uplift is continuing at the present day, but the pattern of uplift and subsidence is complex.

10.4 QUATERNARY DELTA
A 10 m deposit of conglomerate occurs just outside of the Amasing Kota village (Plate 10.1d). The clasts are up to 0.10 m across, subrounded, imbricated and consist of limestone (?Ruta Formation), volcanic rocks (the Bacan, South Bacan and Kaputusan Formations), quartz-monzodiorite and high grade continental metamorphic rocks (Plate 10.1e). The outcrop is poorly lithified and moderately sorted, forming bedding ~5 cm thick. Scouring is absent. The bedding attitude is 180/17 and imbrication direction is 180. A similar deposit at Tg. Joranga has a 30° dip, with comparable bedding and imbrication directions. Consistency of dip and imbrication directions indicates that dip is depositional, with sediment derived from the north. This deposit is interpreted as a river-dominated, Gilbertian-type delta front/plain deposit (Miall, 1984). The age of this deposit is believed to be very young, probably Pleistocene.
Quaternary Deposits

The presence of Sibela Metamorphic Complex clasts in this deposit signifies their first occurrence in younger deposits in the studied area. This implies that the Sibela Metamorphic Complex has been subaerial only recently, indicating an uplift rate in excess of 1 mm/a (1 km/Ma).

10.5 QUATERNARY SCREE
Scree is found in the northern foothills of the Sibela Mountains, as angular blocks of Sibela Metamorphic Complex rocks, up to 2 m across, in relatively flat, heavily vegetated ground. On aerial photographs, it forms very large (up to 8 km x 6 km across) fan-shaped, smooth ground with a dendritic drainage pattern.

10.6 QUATERNARY ALLUVIUM AND FLUVIUM
The fluvial deposits are poorly consolidated, grey, yellow and reddish-pink mud, laminated on a mm scale, occurring along most rivers. Alluvium is present near the mouths of rivers and is flat-lying and heavily vegetated, mostly by mangroves. Aerial photographically they are characterised by flat topography, association with rivers, near coast situation and light colour shading. Both deposits in many cases are controlled by young faults.
CHAPTER ELEVEN
THE GEOLOGY AND TECTONIC EVOLUTION OF THE BACAN REGION

11.1 INTRODUCTION

11.2 THE GEOLOGY AND TECTONIC EVOLUTION OF THE BACAN REGION

11.3 CORRELATION WITH REGIONAL EVENTS

11.4 IMPLICATIONS ON REGIONAL AND GLOBAL TECTONICS

11.5 RECOMMENDATIONS FOR FURTHER STUDIES

11.6 CONCLUDING REMARK
CHAPTER ELEVEN
THE GEOLOGY AND TECTONIC EVOLUTION OF THE BACAN REGION

11.1 INTRODUCTION
Previous chapters have described and interpreted the geology of each of the formations encountered in the Bacan region. This chapter summarises this information, discusses the geology of Bacan in the light of regional tectonic events and proposes a model for the tectonic evolution of the Bacan region. The implications of this model for regional and global tectonic processes are also considered. Finally recommendations for further studies of the Bacan region are put forward.

11.2 THE GEOLOGY AND TECTONIC EVOLUTION OF THE BACAN REGION

Sibela Metamorphic Complex, Continental Suite

The oldest rocks in the Bacan region are thought to be those of the Sibela Metamorphic Complex, Continental Suite. Based on regional stratigraphical arguments, these are postulated to be of Palaeozoic age (Hamilton, 1979). The complex includes continental phyllites, schists and gneisses of upper amphibolite-lower granulite facies (~600°C, ~5 kb), with strong penetrative fabrics indicative of a polyphase deformation and recrystallization history, typical of Barrovian type dynamo-thermal metamorphism. There is also evidence for retrograde metamorphism associated with post peak-metamorphism deformation. The protoliths are pelites with minor amounts of sandy and carbonate/marly horizons. Whole rock chemistry of the pelites suggests that they were derived from the Australian craton and was deposited on an active margin (?continental arc). A Palaeozoic age has not been confirmed isotopically, as K-Ar and Ar-Ar step heating results yielded extremely young ages (<0.21 Ma). These ages are interpreted to be a result of interaction with a hydrothermal fluid carrying fractionated argon, which was subsequently concentrated between the sheet silicate layers, causing an apparent young age. This mechanism may account for other unusually young ages from high-grade metamorphic rocks in the region.

The simplest interpretation is that the Continental Suite was derived from the Australian Plate, where these rocks formed either part of the passive margin of northern Australia or fragments rifted from Australia during Gondwana breakup. Many people have suggested or implied that the continental rocks in Bacan were introduced fairly recently (Pliocene) into the region via strike slip motion of the Sorong Fault Zone (e.g. Hamilton, 1979; Silver et al., 1985) from the Bird’s Head region of Irian Jaya or Central Papua New Guinea (Pigram & Panggabean, 1984). However, Forde (pers.comm, 1993), based on $^{87}$Sr/$^{86}$Sr isotopes, deduced that the Upper Neogene Kaputusan Formation on south Bacan was
erupted on a continental crust, indicating the presence of continental material under the Bacan region by Late Miocene. The presence of Lower Miocene post-collisional intrusion (the Nusa Babi Monzodiorite) in Bacan, also suggested the presence of continental crust by Early Miocene (discussed below).

**Sibela Metamorphic Complex, Ophiolitic Suite**

Juxtaposed against the Continental Suite is the Sibela Metamorphic Complex, Ophiolitic Suite. There are two types of rocks recognised in this suite: [1] ophiolitic rocks, most of which are of lower crustal, plutonic origin with very little volcanic component (e.g. harzburgites, gabbros, microgabbros) and [2] spatially related rocks including strange amphibole cumulates and many rocks with unusual magmatic-tectonic fabrics. This dismembered, incomplete ophiolitic complex is mostly unmetamorphosed, although locally there are mylonitic rocks which have been affected by upper amphibolite-lower granulite facies metamorphism (~1000°C, ~5 kb) related to ductile deformation of hot rocks at or near their place of formation, and local recrystallization in shear zones.

The petrography and whole rock chemistry reveals that most of the cumulate rocks consist mainly of amphibole with minor pyroxene and intercumulate plagioclase, indicating crystallisation under hydrous conditions, implying an arc-related setting. Pyroxene and chrome-spinel compositions suggest that the cumulate may have formed in an arc setting. This is further supported by geochemical evidence derived from the cogenetic metagabbro.

The cumulate rocks in the Sibela ophiolite are very different from those of the Halmahera ophiolite, which consists of olivine, orthopyroxene and clinopyroxene (Ballantyne, 1990; Ballantyne, 1992). The Sibela cumulate is similar to those present in P. Tapas, north of Obi (Hall et al., 1992). The Halmahera cumulates are genetically related to the ophiolite which is interpreted to have formed in a Supra Subduction Zone, normally associated with the initiation of a subduction zone, from the intra-Pacific subduction (Ballantyne, 1990; Ballantyne & Hall, 1990; Ballantyne, 1991b). Preliminary ages on the Halmahera cumulates are Jurassic (Thirlwall pers.comm, 1992).

Ar-Ar and K-Ar dating of the Sibela cumulates and metagabbro yielded Mid-Cretaceous (97-94 Ma) and Oligocene-Miocene (37-25 Ma) ages, although these plateaux ages contain excess argon and therefore will be slightly older than their closure ages (Lanphere & Dalrymple, 1976). The Mid-
Cretaceous age can be correlated with the volcanic activity affecting the eastern Halmahera Ophiolite (80-94 Ma; Ballantyne, 1990), suggesting a link between the Sibela ophiolite and the east Halmahera ophiolite. The Oligocene-Miocene age is related to the Oligocene volcanism and extension event (discussed below).

There are two possible interpretations for the occurrence of the ophiolitic and cumulate rocks. The first possibility is that the ophiolitic rocks represent older crust of possibly Jurassic age (cf. East Halmahera-Gag ophiolite). This ophiolite was intruded by Mid-Cretaceous arc cumulates related to intra-oceanic subduction. This is supported by similarities between the cumulates and zoned ultramafic complexes, normally associated with arc magmatism. Alternatively, the ophiolitic and arc cumulate rocks are all related to the same intra-oceanic subduction during the Mid-Cretaceous. This possibility does not answer the question of the original crust upon which the arc was built.

**Saleh Complex**

The Saleh Complex consists of two types of rocks: metabasites and foliated metasedimentary rocks, mostly phyllites. The metabasites were metamorphosed at conditions between upper prehnite-pumpellyite and lower greenschist facies (∼250-360°C, ∼4 kb). The character of metamorphism suggests static, regional metamorphism. Geochemical and lithological evidence from the metabasites suggests that they represent an arc related calc-alkaline sequence, possibly formed in a back-arc, and may be cogenetic with the Bacan and South Bacan Formations, but distinct from the Sibela ophiolite.

The metasedimentary rocks are highly foliated and have a very different character (higher degree of metamorphism) than the metabasites. These rocks have a similar character to the Sibela Continental Suite, particularly in having brown metamorphic biotite which throughout the whole region is found only in the two complexes, but are possibly of lower metamorphic grade. Isotopic dating shows that the Saleh Complex was affected by a thermal event at ∼12 Ma, probably related to the initiation of the Halmahera arc.

Saleh Complex includes two groups of rocks of different origin, possibly juxtaposed tectonically. These are arc rocks of ?Eocene-Miocene age and continental rocks of Sibela (Australian) origin, and in the whole region, this complex is the only place where high grade continental metamorphic rocks are in contact with arc-related rocks.
The Geology and Tectonic Evolution of the Bacan region

The Bacan and South Bacan Formations

In north Bacan the oldest formation exposed is the Upper Eocene Bacan Formation. This formation comprises interbedded basic-intermediate volcanic and turbiditic volcaniclastic rocks, all of which are folded. The Bacan Formation was metamorphosed under conditions transitional between the prehnite-pumpellyite and pumpellyite-actinolite facies (250-330°C, ~2 kb upwards), with characteristics of burial metamorphism. Retrograde hydrothermal alteration may be related to Neogene Kaputusan volcanism. Mineral and whole rock geochemistry indicates that the volcanic rocks were erupted in an arc setting. Turbidite rocks were deposited by dilute, low concentration, possibly distal or low energy currents. The Bacan Formation has also been affected by the ~15 Ma thermal event.

The oldest rocks in south Bacan are assigned to the Lower Miocene South Bacan Formation; an interbedded sequence of volcanic and volcaniclastic rocks. These rocks were metamorphosed to prehnite-pumpellyite facies (~240-330°C, ~2 kb), with their metamorphic character indicating burial metamorphism. There is local hydrothermal alteration. Whole rock and mineral chemistry indicates an arc origin for the volcanic rocks, similar to the Bacan Formation and the Saleh metabasites. Facies analyses of sedimentary rocks demonstrate deposition by intermediate turbidity currents which dissipated their energies as they travelled down slope. Isotopic dating showed that this formation has been affected by a thermal event at ~8 Ma.

The Bacan and South Bacan Formation are similar in all aspects, with age differences as the only distinction between the two formations (Bacan Formation ~39-35 Ma, South Bacan Formation ~23-22 Ma). There are two possible tectonic scenarios for the deposition of these formations: [1] the Bacan and South Bacan Formations represent different parts of a single arc active from Late Eocene until Early Miocene (Fig. 11.1a) or [2] the Bacan and South Bacan Formations are two separate arcs active at different times (Fig. 11.1b). More dating is needed, although this could prove difficult. Despite examination of nearly two hundred thin sections, only one sample was suitable for dating. This is attributed to lack of fossils in the original sediments, metamorphism which destroyed fossils and thermal overprint on volcanic rocks.

In the first explanation, the Bacan and the South Bacan Formations represent a long-lived or an intermittently active arc which existed from Late Eocene to Early Miocene. The Oligocene Tawali Formation of Kasiruta may represent part of the same arc. Continental crust was subsequently thrust
Figure 11.1 Different possibilities for the relationship between the Bacan, South Bacan and Tawali Fms.
under southern Bacan. Although this alternative is plausible, there are several arguments against this: stratigraphical evidence, such as lack of Oligocene rocks in Bacan, and the differences in lithology and metamorphic character (particularly the lower grade of metamorphism suffered by the Tawali Formation relative to the older and younger formations) and trace element differences between the Tawali and the Bacan/South Bacan Formations, argues against it. The continuous arc interpretation is favoured here because all of these formations post-date the Eocene Pacific Plate reorganisation event and they pre-date the ~22 Ma unconformity (discussed below) interpreted to be related to the arrival of the continental crust in the region. Therefore it is simplest to attribute all of these formations to a single arc event. Differences in lithology and metamorphic character and geochemistry may be due to different positions that these formations occupy within the arc-system or difference structural levels that these formations occupy in the collisional stack.

**The Tawali Formation**

The oldest rocks on Kasiruta belong to the previously unrecognised Tawali Formation, consisting of pillow lavas (Jojok Member) and volcaniclastic turbidites (Marikapal Member). This formation is affected by burial metamorphism transitional between low and high temperature zeolite facies (~180°C, <2 kb). Whole rock chemistry indicates that the Lower Oligocene basalts are highly differentiated arc lavas, erupted in a deep, open marine environment above the CCD. The Upper Oligocene volcaniclastics are products of repeated, high density, proximal turbidity currents with associated slumped deposits. The thickness of the two members and the narrow age range indicate rapid deposition.

Similar rocks to the Tawali Formation have now been identified throughout the Halmahera-Waigeo region (the Tawali Formation in NW Halmahera, the Anggai River Formation in Obi and the Rumai Formation in Waigeo; Hall et al., 1992). All of these rocks may be equivalents in age and in their arc origin. Their differences in lithology could be explained by the different formations representing different parts of the arc (e.g. arc slopes, back arc basin, forearc slopes). Differences in the grades of metamorphism could be a function of different structural positions during collision (with the Australian continent) or different thermal environments due to different positions in the pre-collision arc. The Bacan and South Bacan Formations may also be part of this arc, as is the Oha Formation which is probably not of Upper Cretaceous-Eocene age as previously suggested (Hall et al., 1988a; Hakim & Hall, 1991). The simplest model to account for all these Upper Eocene, Oligocene and
The Geology and Tectonic Evolution of the Bacan region

Lower Miocene volcanic arc formations in the region, is that they all represent arc volcanism at the edge of the Philippine Sea Plate due to the subduction of Indian Plate (the leading edge of the Australian Plate) under the Philippine Sea Plate (Fig. 1.2). This is supported by the fact that all of these formations record similar palaeomagnetic declinations and inclinations which are characteristic of the Philippine Sea Plate (Hall et al., 1993).

**Early Miocene Unconformity**

Following the deposition of the Bacan, Tawali and South Bacan Formations, there is a major regional unconformity. This is characterised by a major hiatus, change in lithological, metamorphic and structural character. Rocks below the unconformity are folded arc volcanic and turbidites, metamorphosed up to greenschist facies, whereas the rocks above it are mostly unfolded, unmetamorphosed carbonates. This unconformity has been dated throughout the Halmahera region at ~22 Ma (Hall et al., 1992).

**Nusa Babi Intrusive Rocks**

The Nusa Babi Monzodiorite (NBM) includes quartz-monzodiorite dykes and plugs with associated 'aplite' dykes of monzogranite composition. These intrude both the Bacan and the South Bacan Formations and have been reported to intrude the Sibela Complex (Yasin, 1980), but did not intrude the Ruta and Amasing Formations. The NBM is locally altered and these may be due to auto-metasomatism. Magma evolution is primarily controlled by fractionation and the addition of water. Whole rock geochemistry indicates that the NBM is of plutonic arc or collisional magmatism origin. Field and isotopic age evidence indicate a post-collisional origin, related to the arrival of Sibela Continental Suite in the region. K-Ar dates show an Early Miocene age with a thermal overprint of Pliocene age.

**The Ruta and Amasing Formations**

The Lower-Middle Miocene Ruta Limestone lies unconformably above the Bacan, Tawali and South Bacan Formations. There are four microfacies recognised in the Ruta Limestone: skeletal wackestones-packstones, algal boundstones, foraminiferal packstones and bioclastic-lithoclastic packstones. These represent deposition on open platform, platform margin build up (patch reef), tidal bar downslope from patch reef (open platform) and foreslope talus, respectively, forming a shallow marine carbonate platform. The deposition of the Ruta Formation began in Early Miocene times, predominantly as an
Late Eocene - Early Miocene Arc, due to subduction of the Indo-Australian Plate under the Philippine Sea Plate.

No Scale Intended

Figure 11.2 Different positions of the volcanic formations in the region within the L. Eocene - E. Miocene arc.
open platform, platform margin buildup and foreslope talus facies. Between Early Miocene and late Middle Miocene times, deposition was locally interrupted by sudden and high influxes of volcaniclastic material forming the Amasing Formation. Three facies in the Amasing Formation have been recognised: shallow marine with storm horizons, shoal or estuarine, and beach deposits. These facies represents shallowing upward from the open carbonate platform to beach deposit. Deposition of the Ruta Formation continued until late Middle Miocene times and was dominated by tidal bar facies.

There is widespread development of carbonate platforms at this time throughout the Halmahera-Waigeo region (Hall et al., 1992), along the northern Irian Jaya (Pigram & Davies, 1987; Pigram et al., 1990) and in the Philippines (Mitchell et al., 1986). There was no active arc in northeastern Indonesia during this time, except minor local activity. Widespread carbonate platform development suggests that these regions were in an equatorial position and were part of an extensive shallow marine area.

The Kaputusan Formation
Following the deposition of the Ruta and Amasing Formations, there is a Late Miocene regional unconformity. The Kaputusan Formation was deposited above the unconformity. This formation consists of three members: the Goro-goro Volcanic, the Pacitak Volcaniclastic and the Mandioli Limestone.

The Goro-goro Member consists of Two Pyroxene Andesites (TPAN), Hornblende Pyroxene Andesites (HPAN), Hornblende Andesites (HBAN) and Hornblende Biotite Andesites (HBIAN). Associated with the volcanic rocks are subaqueous pyroclastic flows with related base surge deposits. The petrography, mineral chemistry, whole rock major and trace element chemistry of these rocks are typical of island-arc volcanic rocks. They are mostly fresh, although locally they are metamorphosed to the zeolite facies, interpreted as due to hydrothermal activity. Magma diversification of the Goro-goro Member was achieved mainly by fractionation of plagioclase, pyroxene, Fe-Ti oxide, amphibole and biotite, with evidence for replenishment, immiscibility, resorption and assimilation. Magmatic conditions were <8 kb and 2-10 wt% H₂O with P(H₂O)<Ptotal. The Goro-goro Member includes at least four eruption centres (as defined by bulk trace element ratios): in south Bacan (mostly HBIAN), the Goro-goro area (mostly TPAN), the North Mandioli and the Kaputusan areas (both are largely HPAN and HBAN).
The Geology and Tectonic Evolution of the Bacan region

The Mandioli group are shoshonitic rocks, which may indicate eruption in an extensional setting within a dominantly convergence zone, such as a strike-slip zone where similar rocks commonly occur (e.g. Ellam et al., 1988). K-Ar ages indicate that these centres were erupting from Late Miocene to Pleistocene. Each of these centres was active for ~2 Ma, with a history of multiple eruptions. The oldest volcano occurs in south Bacan, and the arc migrated northwards. Some of these volcanoes may have been erupted simultaneously.

The Upper Miocene-Upper Pliocene Pacitak Member consists of reworked pyroclastic and volcaniclastic material from the Goro-goro Member, deposited in an oxygen-rich, near shore, shallow marine environment, proximal to a volcanic terrain. The Upper Miocene-Lower Pliocene Mandioli Member consists of wackestones-packstones which formed fringing coastal reefs.

The Kaputusan Formation is interpreted to be the product of the eastward subduction of the Molucca Sea Plate under Halmahera, and can be correlated with the Weda Group of Halmahera and the Woi Formation of Obi (Hall et al., 1992). The arc initiated at ~15 Ma, recorded by thermal events affecting the Bacan Formation and the Oha Formation in Halmahera (Hall et al., 1992). The oldest recorded regional isotopic age of this arc volcanic is from Obi at ~12 Ma and the age becomes younger northward (S.J.Baker pers.comm, 1993), consistent with the results from Bacan. Small scale rifting separating Bacan and Kasiruta is interpreted to be related to Kaputusan volcanism between the two islands.

**The Saleh Diorite**

The Saleh Diorite intrudes the Saleh Complex and is juxtaposed against the Sibela Continental Suite. This intrusion consists of amphibole-bearing diorite and micro-diorite. Locally these rocks suffered from low temperature alteration, probably due to auto-metamorphism, and subsequent hydrothermal alteration. Whole rock geochemistry indicates that the Saleh Diorite is different from the NBM and is of arc or post-collisional magmatism origin. K-Ar dates show an ~15 Ma age is related to the initiation of the Halmahera arc, and ~9 Ma reset age related to the Neogene volcanism.

**Quaternary Deposits**

Although currently there are no active volcanoes at Bacan, Quaternary volcanic rocks have been documented. These consists of fresh olivine, pyroxene and plagioclase-phyric basalts. Mineral and
whole rock geochemistry indicate that these are arc basalts, with 150-180 km distance to the top of the Benioff zone, erupted through a quartz-rich basement (?Sibela Continental Suite). There is evidence of a within-plate basalt component from the whole rock and mineral chemistry, possibly indicating magmatism related to movements along the sinistral Sorong Fault Zone. The linear distribution of the volcanic centres also supported this.

Detritus from the Sibela Continental Suite can be found solely in the Quaternary deposits, indicating that it has become available for erosion only recently, demonstrating extremely fast rates of uplift (~1 mm/yr). It is postulated that this uplift rate may be related to the effect on the mantle from the collision of the Halmahera and Sangihe arcs.

11.3 CORRELATION WITH REGIONAL EVENTS

Regional Setting

The Bacan region is located within the convergence zone of Indo-Australian, Philippine Sea and Eurasian Plates. The Indo-Australian Plate has been moving northward since the Cretaceous, subducting under the southern edge of the Philippine Sea Plate from Cretaceous to early Tertiary and later in some places (Hall et al., 1992). The Eurasian Plate in the region is broadly a continental margin (including continental crust in the Philippine) with areas of volcanic arc (e.g. the Sangihe arc) (Rangin, 1991).

Following the collision of India and Eurasia at ~40-45 Ma, Indochina was extruded into Southeast Asia, causing the opening of South China Sea (Taponnier et al., 1982). At the same time the Pacific Plate was experiencing a major plate reorganisation (Uyeda & Ben Avraham, 1972; Seno & Maruyama, 1984) accompanied by a change in the plate motion, from northward to westward, indicated by the bend in the Hawaii-Emperor Seamount chain (Clague & Jarrard, 1973). This change may be related to the collision of India and Eurasia (Dalrymple & Clague, 1976).

The motion of the Philippine Sea Plate has been complex, although recent palaeomagnetic results are developing a good clear picture. The work by Hall et al. (1993) shows that between ~50-40 Ma, the plate experienced a ~50° clockwise rotation with a southward translation; no rotation was detected from 40-25 Ma and between 25-0 Ma, the plate rotated clockwise 35° with northward translation (Fig.11.3). The first rotation may be related to the change of motion in the Pacific Plate, whereas the
second may be related to the collision of Australia with Philippine Sea Plate (discussed below).

Figs. 11.4-11.9 are simplified cartoon diagrams for the tectonic evolution of the Bacan region, which incorporate the palaeomagnetic results of Hall et al. (1993). The pre-Neogene reconstructions are less reliable due to the different possibilities of the tectonic setting of the Bacan, South Bacan and Tawali Formations (Chapters Four-Six). The pre-Tertiary model is more speculative because of lack of data.

**Pre-Tertiary Evolution of Bacan**

There are two types of rocks in the pre-Tertiary Bacan: continental crust and ophiolitic-arc material. The continental crust is presumably of Australian margin or microcontinents origin. The leading edge of the Australian Continent was the Indian Oceanic Plate.

The arrival of the Australian continental rocks in Bacan (Sibela Continental Suite) is due to [1] pre-Late Miocene collision, [2] a strike-slip translation and subsequent thrusting under Bacan in the pre-Late Miocene time, or [3] collision followed by strike-slip translation and subsequent thrusting. Although there is no clear evidence to favour one of the options stated above, the proximity of the Sibela Continental Suite to the arc interpreted to be the result of subduction of the Indo-Australian Plate under the Philippine Sea Plate (Chapters Four-Six) and the juxtaposition of the continental rocks with the ophiolite interpreted to represent Philippine Sea Plate material (see below), favours the collisional interpretation.

Ophiolite-arc material in Bacan recorded Cretaceous and Oligo-Miocene ages, both of which can be correlated with ages recorded by the Philippine Sea Plate. This implies that the ophiolitic rocks of the Bacan region have been part of the Philippine Sea Plate since at least the Mid-Cretaceous time (Fig. 11.4).

**Early Tertiary**

There is no lithostratigraphic evidence in the Bacan region between Mid-Cretaceous to Late Eocene. Lack of deposition may be due to uplift following the regional unconformity related to the ~45 Ma Pacific Plate reorganisation (Uyeda & Ben Avraham, 1972; Seno & Maruyama, 1984).
Figure 11.3 Tectonic reconstruction of the Philippine Sea Plate (after Hall et al., 1993).
There was extensive arc activity throughout the Halmahera region, including the Bacan area, in Late Eocene-Early Miocene. Sukamto et al. (1981) suggested that an Oligocene arc in the Halmahera-Bacan region is a result of westward-dipping subduction of the Philippine Sea Plate, implying that the region lay on the Eurasian Plate. Recent palaeomagnetic evidence (Hall et al., 1993) from the Tawali Formation shows that the region was actually part of the Philippine Sea Plate. The Oligocene Tawali Formation is therefore interpreted as the product of arc volcanism, along the edge of the Philippine Sea Plate, resulting from the rapid northward subduction of Indo-Australia Plate (Ben Avraham, 1978) under the newly formed Philippine Sea Plate (Hilde & Lee, 1984), following the Pacific Plate reorganisation.

The Upper Eocene Bacan Formation and the Lower Miocene South Bacan Formation are either [1] genetically related to the Tawali Formation and represent temporal and spatial variations of the same continuous arc or [2] they represent separate arcs from the Tawali Formation, each different from each other. Arguments for a continuous arc hypothesis (Fig.11.5) have been outlined in Section 11.2.

There are two possible correlatable volcanic arcs in the northeastern Indonesia - southwest Pacific area: the Oligocene-Miocene volcanic rocks of Irian Jaya and Papua New Guinea (e.g. Pigram & Davies, 1987) or the Eocene-Oligocene Palau-Kyushu remnant arc (Sutter & Snee, 1980). The Palau-Kyushu and the West Mariana ridges are associated with opening of the Parece-Vela Basin (at 30-17 Ma) and these remnant arcs are attributed to subduction of Pacific ocean (e.g. Uyeda & Ben-Avraham, 1972; Seno & Maruyama, 1984) and are therefore a different arc than the Tawali Formation’s (Fig.11.3). The arc volcanic rocks of northern Irian Jaya, e.g. the Arfak (Ratman & Robinson, 1981; Pieters et al., 1982), Batanta (Sanyoto et al., 1985), Yapen (Atmawinata et al., 1989) and Mandi Formations (Hartono et al., 1989) and Papua New Guinea, e.g. the Bismarck Volcanic Province (Dow, 1977), are interpreted to be related to the subduction of the Indo-Australian Plate under the Philippine Sea Plate (Dow, 1977; Pigram & Davies, 1987) and may be a continuation of the Bacan-Tawali-South Bacan Formations arc. Correlation of volcanic rocks in Bacan and the Bird’s Head had been indicated earlier by Van Bemmelen (1949) and Verstappen (1964) based on lithological and aerial photographic characteristics, respectively. Rangin (1991) proposed that the Oligocene volcanic rocks of Luzon and Bicol in the Philippines are similarly related to the northward subduction of the Indo-Australian Plate under the Philippine Sea Plate.
The Geology and Tectonic Evolution of the Bacan region

22 Ma Unconformity
Following the deposition of the South Bacan Formation, there is a regional unconformity (~22 Ma; Hall et al., 1992) interpreted as resulting from the collision of Australian and Philippine Sea Plates (Fig. 11.6). This is supported by the common lithostratigraphy, represented by carbonate rocks, above the Upper Eocene-Lower Miocene arc rocks (Philippine Sea Plate affinity) and those of Australian continental affinity; changes from folded, metamorphosed arc rocks to unmetamorphosed carbonates and possible stitching (intrusion) of Philippine Sea Plate rocks (the Bacan and South Bacan Formations) and Australian Plate rocks (Sibela Continental suite) by the Nusa Babi Monzodiorite at 19.8 ± 1.6 Ma.

The collision of the Australian Plate with the Pacific or Philippine Sea Plates has been dated at Early Miocene on northern Papua New Guinea (Jaques & Robinson, 1977) and tentatively at Middle Oligocene on Waigeo (Charlton et al., 1991), however this is now revised to an Early Miocene age (Hall et al., 1992). In contrast, Pigram & Davies (1987) suggested that there were several terranes in the Pacific Plate which started to collide with the Australian Plate in the Middle-Late Oligocene and continued until the Pliocene. Pigram & Symonds (1991) produced a compilation of suggested times of collision, by different workers, which range from Eocene to Late Miocene, although most workers in New Guinea recognise Early Miocene as the key event (e.g. Dow, 1977; Jaques & Robinson, 1977).

19-14 Ma
After the ~22 Ma unconformity, the region was uplifted and deposition of shallow marine carbonate sequences followed. Strike slip related basins may explain the development of a thick, laterally extensive carbonate platform. This is consistent with the palaeomagnetic results (Hall et al., 1993) showing the Halmahera region, including Bacan, sliding along the northern margin of the Australian Plate along a proto-Sorong Fault system. Localised uplift and volcanic events may have contributed influxes of volcaniclastic material onto the carbonate platform (Fig. 11.7).

Initiation of Molucca Sea subduction at ~15-12 Ma
There is another regional unconformity of early Late Miocene age following the Early to Middle Miocene carbonate platform development. The Upper Miocene Kaputusan Formation was deposited after the unconformity. This formation is the equivalent of other Upper Miocene arc sequences in Halmahera and Obi and they are interpreted as the product of arc volcanism resulting from the
eastward subduction of the Molucca Sea Plate under Halmahera (the Philippine Sea Plate) at the Halmahera trench (Fig.11.8).

Although the oldest isotopic age obtained from the Kaputusan Formation is ~7.5 Ma, there are thermal events at ~15 Ma affecting the Bacan and Oha Formations. The oldest isotopic age obtained from similar sequences on Obi yielded an ~12 Ma age. The Saleh Diorite, interpreted to be a precursor to the Kaputusan Formation, was intruded at ~15 Ma. Initiation of subduction possibly started at ~15 Ma and the first product of volcanism was erupted at ~12 Ma.

There are major plate configuration changes in the region occurring between 15 and 12 Ma which may have triggered the Molucca Sea Plate eastward subduction. Spreading in the South China Sea and Shikoku Basin ended at ~15 Ma (Jolivet et al., 1989; Briais et al., 1993; Taylor, 1992). At ~14 Ma the Cagayan Arc collided with Eurasia and the Palawan Arc collided with Mindoro (Jolivet et al., op. cit). From the Early Miocene, the east-west compression between the Eurasia and the Philippine Sea Plates was accommodated by subduction of the Philippine Sea at the Sangihe, Ryukyu and Manila Trenches (Jolivet et al., op. cit). Following the end of spreading and collision, compression between Eurasia and the Philippine Sea Plate was taken up by the Halmahera Trench. This differs from the model of Hall (1987) who, based on younger biostratigraphic dates, attributed subduction under Halmahera to the collision of East and West Mindanao during the Pliocene.

At ~3 Ma there is 60 km shortening between east and west Halmahera, which is attributed to movement along the Sorong Fault Zone (Nichols & Hall, 1991). Splays of the Sorong Fault Zone running through the Bacan region may have accommodated the Molucca Sea - Philippine Sea Plates compression and thus ended Kaputusan volcanism. Quaternary volcanism was later reinitiated along the fault splays (Fig.11.9). These faults are also responsible for the shaping of Bacan coastlines and the creation of deep basins surrounding Bacan.

Fig.11.10 is a summary of the stratigraphy of Bacan with global and regional tectonic events which may have influenced its tectonic development. A simplified geological map of the Bacan region is provided in Plate 11.1. This map is based on aerial photographic interpretation, constrained by field data and superimposed on a digitised 1:250,000 topographic map obtained from GRDC, Bandung.
Figure 11.4. Simplified tectonic setting of the Bacan region during the Mid-Cretaceous.
Late Eocene

Subduction of Indo-Australian Plate under the Philippine Sea Plate. Bacan region is the arc.

Oligocene

Continuation of subduction of Indo-Australian under the Philippine Sea Plate. Bacan region is the back-arc.

Early Miocene

Continuation of subduction of Indo-Australian under the Philippine Sea Plate, culminating with collision of continental crust with the arc.

Figure 11.5. Simplified tectonic setting of the Bacan region during the Late Paleogene - Early Miocene.
Arc-continent collision. Transfer of Australian leading fragments to Philippine Sea Plate which subsequently rotated westward.

Figure 11.6. Simplified tectonic setting of the Bacan region during the Early Miocene.
Northward movement of Australia is accommodated by transform boundary which moves north at similar speed to Australian Plate. Philippine Sea Plate is rotating clockwise due to subduction on all sides. Widespread development of carbonate platform.

Figure 11.7. Simplified tectonic setting of the Bacan region during the Middle Miocene.
Figure 11.8. Simplified tectonic setting of the Bacan region during the Late Miocene.
Sorong Fault Zone transform boundary cuts through Bacan region. Cessation of volcanism due to subduction of Molucca Sea Plate. Start of volcanism along SFZ splays. Fault controlled coast lines and basin development between strands of SFZ. Some composite fragments rotated counterclockwise in megashear (e.g., Obi).

Figure 11.9. Simplified tectonic setting of the Bacan region during the Late Pliocene - Quaternary.
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<table>
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<tr>
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<td>Collision of east and west Malahera, movement on splay of Sorong Fault Zone.</td>
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<tr>
<td>Start of volcanic in Bacan (~7 Ma). Malahera and Obi.</td>
</tr>
<tr>
<td>Four eruptive centres which migrated northwards with time.</td>
</tr>
<tr>
<td>Associated subaerial and submarine (shallow marine) pyroclastic deposits, and fringing reefs.</td>
</tr>
<tr>
<td>Collision of Cagayan with Eurasia, Pacific with Mindoro, end of spreading in Shikoku and South China Sea.</td>
</tr>
<tr>
<td>Initiation of eastward Molucca Sea Plate subduction under Malahera.</td>
</tr>
<tr>
<td>Thermal event (~15 Ma).</td>
</tr>
<tr>
<td>Shallow marine carbonate platform as part of the widespread regional carbonate buildup.</td>
</tr>
<tr>
<td>Open platform, patch reef, tidal flat, and foreslope talus.</td>
</tr>
<tr>
<td>Local development of shallow marine volcanoclastic deposit (shallow marine, shoal, and beach).</td>
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<tr>
<td>Sedimentation on strike-slip zone.</td>
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<td>Arc volcanic and turbidite.</td>
</tr>
<tr>
<td>Metamorphosed to Prehnite-Pumpellyte facies.</td>
</tr>
<tr>
<td>Repeated high density, proximal volcaniclastic turbidite.</td>
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<td>Pillow lavas, zoic facies metamorphism.</td>
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<td>Geochemical data suggests a volcanic arc origin.</td>
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<tr>
<td>Arc volcanic and turbidite.</td>
</tr>
<tr>
<td>Metamorphosed to Prehnite-Pumpellyte - Lower Greenschist facies.</td>
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<tr>
<td>India-Eurasia collision (~ 45 Ma) and major plate reorganization in the Pacific.</td>
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<tr>
<td>Opening of the South China Sea. Isolation of Philippine Sea Plate.</td>
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<tr>
<td>Thermal event in the Sibea Ophiolite (~ 3 Ma).</td>
</tr>
<tr>
<td>Metasomatised and phyllite, dynamothermal metamorphism (360°C, 4kb). Arc (back-arc) origin for basalt, phyllite related to Sibea Complex. Thermal event at ~12 Ma.</td>
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<tr>
<td>Related to Bacan Fm.</td>
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<td>Continental dynamothermal metamorphic (~ ~550°C, 5kb). Arc protolith, Sibea Complex.</td>
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<tr>
<td>Mytilus gigus, age Philippine Sea Plate affinity.</td>
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</table>

Figure 11.10. Stratigraphy of the Bacan region with related regional and global tectonic events.
Stratigraphical relationship is depicted in Fig.11.10
Structural cross-sections are presented in Fig.11.11

Some areas may be speculative due to lack of data.

Figure 11.11 Structural cross-sections from the geological map of the Bacan region.
11.4 IMPLICATIONS ON REGIONAL AND GLOBAL TECTONICS

This study has provided important geological data and elucidated the tectonic development of the Bacan region. On the regional context, this study has contributed to the tectonic history of the zone of convergence between the Australian, Philippine Sea and Eurasian Plates in eastern Indonesia. Some of the important implications are:

[1] The Sibela Continental Suite has arrived in the region due to collision of Australia or rifted margin of Australia with the Philippine Sea Plate, not by movement along the strand of the Sorong Fault Zone (e.g. Hamilton, 1979; Silver et al. 1985). This is in accordance with the view of Charlton (1986) and Hall et al. (1988a).

[2] The evolution of the southern part of the Philippine Sea Plate has been elucidated by this study. The Bacan Region has always been part of the Philippine Sea Plate since Mid-Cretaceous demonstrated by the presence of the Sibela Ophiolite Suite. The Bacan, South Bacan and Tawali Formations are most likely to be an arc-related deposit due to the subduction of Australia under the Philippine Sea Plate. Collision of Australia with the Philippine Sea Plate can be dated at ~23 Ma in the Bacan region.

[3] Presently there are different models for the tectonic development of northeastern Indonesia, e.g. the indentor model of Charlton (1986); the terrane model of Silver et al. (1985); and the marginal basin model of Ben-Avraham (1978), none of which are based on sufficient geological data. The Ben-Avraham model puts Halmahera and Bacan as part of the Australian continent, which is wrong as most of the region has been part of the Philippine Sea Plate since Mid-Cretaceous. Both the Charlton and Silver models predicted that the Sibela continental block was translated by a very recent strike slip motion to the Bacan region, which has been disputed in this study. Hall et al. (1993) provide the only model based on sound geological and geophysical data.

In a global context, if one follows the terrane definition of Howell et al. (1985), the Sibela Continental Suite, the Sibela Ophiolite Suite, the Bacan, Tawali and Kaputusan Formations will be considered as five different tectono-stratigraphic terranes, and may lead to erroneous interpretation in assigning separate histories for each terrane. Karig et al. (1986) for instance attributed the rocks in the northern Philippines which have a similar character to the Sibela Continental Suite, Sibela Ophiolitic Suite,
The Geology and Tectonic Evolution of the Bacan region

Bacan and Kaputusan Formations as separate allochthonous terranes juxtaposed by strike-slip faulting. Whilst not arguing against their interpretation, their study can be used as an example of terrane tectonic concept, whereas this study has shown that having rocks of different character juxtaposed against each other does not necessitate the notion of allochthonous terranes. In the Bacan region, the only fragment that can be considered an allochthonous terrane is the Sibela Continental Suite. The region has always been at the edge of the Philippine Sea Plate, and the different 'tectono-stratigraphic terranes' reflects different tectonic regimes at the edge of the plate.

The use of stratigraphical similarities is often used in the literature to correlate tectono-stratigraphic terranes and ultimately tectonic history (e.g. Hamilton, 1979; Pigram & Panggabean, 1984). Using this technique, the Bacan region could be correlated with parts of Papua New Guinea (Eocene pillow lavas against continental basement; Dow, 1977), Zamboanga Peninsula of Mindanao (high grade continental rocks juxtaposed against metaophiolite unconformably overlain by Middle Miocene volcanic rocks; Rangin, 1991) and northern Philippines (see above; Karig et al., 1986). This type of correlation will lead to erroneous interpretations as each of these regions has different tectonic affinities (Bacan was and is still part of the Philippine Sea Plate; Papua New Guinea was part of the Philippine Sea Plate, but now is part of the Australian Plate; Zamboanga and northern Philippines were and are still part of the Eurasian Plate). Similarities in the stratigraphy is a function of similar tectonic histories (areas recording the collision of continental with oceanic plate), and one should be cautious in correlation and interpretation of movement of terranes along strike slip fault.

All too often tectonic reconstruction of a region is based on meagre geological data, such as in the Bacan region, resulting in a collage of interpretations (Fig.2.3). Although this study has provided the most comprehensive geological data from the Bacan region, the history of the region at the Early Miocene and before the Eocene is still unclear. Bearing this in mind, one should be careful in interpreting the tectonic evolution of other, similarly complex, older regions.

11.5 RECOMMENDATIONS FOR FURTHER STUDIES

The purpose of this study was to elucidate the geology and tectonic evolution of the Bacan Region. Although this study attempts to be as comprehensive and multi-disciplinary as possible, time constraints on this research prevented further development of some of the more interesting aspects of the study. Recommendations for further work are listed in order of relative ages of sequences.
For the Sibela Metamorphic Complex, Continental Suite, the study of the metamorphic facies and their distribution will be of interest and may shed light into the relationship of the continental and ophiolitic suites. The effect of hydrothermal fluids on the K-Ar system is of importance to further the understanding of the K-Ar isotope systematics. The isotopic age of both the continental and ophiolitic suites, to decipher the formational age of these complexes, will be important in understanding the pre-Cainozoic history of the region. The relationships between the ophiolitic and cumulate rocks and between the deformed metamorphosed rocks and the ones which have retained their magmatic fabric, in the Sibela ophiolite may further refine the geological history of the region.

The relationship, similarities and differences between the Saleh Complex and the Bacan and South Bacan Formations is critical in understanding the pre-Neogene history of the region. Radiogenic and stable isotope work may clarify this problem. Study of the metamorphic facies, of both the Bacan and South Bacan Formations, and their distribution will further the understanding of metamorphic processes of arc-related basins. The correlation of the Tawali Formation with the arc in the Philippines, northern Irian Jaya and Papua New Guinea needs to be confirmed. Work on the petrology and geochemistry of these proposed equivalent arc will be very useful.

The distribution and control of carbonate facies variations of the Ruta Formation in the context of the whole eastern Indonesia Lower Miocene carbonate rocks may help the understanding of post-collisional carbonate platform development. Isotopic studies on the Kaputusan Formation may reveal the extent of the continental crust underneath Bacan. Comparison of the Quaternary volcanic rocks with volcanic rocks along the Sorong Fault Zone (e.g. Misool and Salawati) may provide an insight into processes of volcanism along a major strike-slip plate boundary.

Detailed structural studies on each of the formations will further refine the tectonic model and identified suture zones, although poor exposure and complication due to younger deformation may hamper structural interpretation.

11.5 CONCLUDING REMARK
That's it!!
APPENDIX A

MICROPROBE RESULTS

Mineral formula proportions were calculated on the basis of the following number of oxygens: olivine (4), pyroxene (6), analcite (7), glass, ankerite and errionite (10), prehnite (11), Fe-Ti oxides (16), sphene (20), sheet silicate except chlorite (22), amphibole (23), garnet (24), epidote (25), apatite (26), chlorite (28), feldspar (32), mesolite/thomsonite (30), staurolite (48), pumpellyite (49), heulandite/stilbite/ferriarite (72) and natrolite (80). Calculation for Fe$^{3+}$ was performed assuming 12 cations.

The microprobe results are presented in the enclosed Microfiches and these are arranged according to the order of discussion in the thesis.
APPENDIX B

Results of Whole Rock Chemical Analyses using X-Ray Fluorescence Technique

Loss On Ignition (LOI) were calculated at ~1100°C, expressed in weight percent. Major elements are in weight percent, trace elements are in parts per million (ppm) calculated on a volatile free basis.

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APPENDIX C

Results of the Bacan Formation Whole Rock Chemical Analyses using Induced Coupled Plasma (ICP) Technique.

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APPENDIX D
ISOTOPIC DATING

D.1 SAMPLE SELECTION PROCEDURES
The following criteria are used in selecting samples: [1] samples are as fresh as possible; [2] xenoliths or rocks which do not characterise the formation are avoided, in situ samples are preferred to ensure that samples are exemplary of formation; [3] although only several grams are required for both the potassium and argon analyses, it is better to have large samples (~2 kilograms) so that petrographic and chemical analyses can also be performed; [4] minerals with high K content, such as mica and hornblende, are ideal for dating. Whole rock analyses can be performed with the risk of heterogeneous argon retentivity from different minerals. In the case of volcanic samples, vesicular glassy material is best avoided due to possible excess argon from the mantle (Faure, 1986). Devitrified or hydrated glass will not be suitable because of its potential argon loss. However, young, fresh and glassy volcanic rocks are suitable, as the argon loss is minimal, due to rapid solidification which effectively creating one blocking temperature. In these instances the age derived is as reliable as the mineral separate.

D.2 ANALYTICAL PROCEDURE
The K-Ar analyses are performed at NERC Isotope Geology Laboratory (NIGL) at Keyworth, Nottingham, under the supervision of Dr C.C.Rundle. The Ar-Ar extraction and isotopic analysis are performed in the University of Leeds, supervised by Drs. D.C.Rex and P.G.Guise.

D.2.1 Rock Crushing and Mineral Separation
A sub-sample of the rock is crushed using the jaw crusher several times, then milled with a "BICO Pulverizer" to the size required to liberate the mineral for analysis. The crushed rock is further sieved through the 1mm, 500, 250 and 125 μm openings. Jäger et al. (1985) reported that the best grain size range for analysis is 60-80 mesh fractions (250-180 μm). This is further supported by Chen et al. (1984) where they found that 50-80 mesh fractions correspond well with the mean value for both K and $^{40}$Ar$_{sm}$. These sizes provide a balance between sample heterogeneity in large fractions and increased atmospheric argon contamination in the finer fractions.

Mineral separation technique can be accomplished using either magnetic separation or heavy liquid separation. Two types of magnetic separator are used, the "Davies" rotational magnetic separator and "Frantz Isodynamic" barrier magnetic separator. While the former relies on a rotating magnet to pick the magnetic material and deposit it in separate cups, the latter relies on gravity to drop the non-magnetic material while the magnet holds the magnetic ones. A hand magnet is utilised.
to separate the highly magnetic fractions (i.e. magnetite) from the rock. Heavy liquids are used to separate minerals based on their specific gravity. Diiodomethane (S.G. 3.315-3.25 g/ml @ 20°C) and Bromoform (S.G. 2.880-2.910 g/ml @ 20°C) are the most useful liquids, the former to separate hornblende, while the latter to separate feldspars. To achieve final purity, non-composite grains were hand picked under a microscope.

The sample is washed to eliminate the finer size particles, which have larger surface area and therefore are prone to atmospheric argon contamination. This is done by putting the powdered sample in a beaker, filled with distilled water, stirring it until the finer particles are suspended in water and then disposing of the water. When the water becomes clear after stirring, the sample is washed with de-ionised water at least a couple of times to rinse away any foreign ions and then dried in an oven. The resulting powder at this stage should weigh at least 15 grams.

D.2.2 Potassium Content Analysis
Powdered sample is weighed using "Sartorius" digital scales with 0.00001 g precision ( ~0.2-0.3 g for amphibole, ~0.08-0.10 g for biotite). The sample is then placed in a 25 ml platinum crucible and dissolved using ~2 ml concentrated perchloric acid and 8 ml 40% hydrofluoric acid. The crucibles are handled using platinum-tipped tongs only, to prevent K contamination from organic sources. The mixture is then evaporated overnight on a 150°C hot plate in a fume hood. HF during this time breaks down the silicate lattice to form silicon tetrafluoride, which will vaporise, and fluorides. Perchloric acid converts the insoluble fluorides to highly soluble perchlorates and the solution eventually boils to dryness.

The residue is then diluted with de-ionised water, acidified with 6M HCl to aid dissolution. This solution is transferred to a clean volumetric flask (100, 250, 500 ml depends on the expected K content, as the flame photometry works best for 5-25 ppm concentration). After thorough mixing, a 5 ml aliquot is mixed with 5 ml of standard lithium solution (~200 ppm Li).

The potassium content of the mixture of the sample and lithium solution can be determined using "Instrumentation Laboratory Flame Photometer 543," for Na, K and Li concentration with direct digital reading. Dalrymple & Lanphere (1969) provided a detailed explanation of flame photometry principles. The reading is relative to Li internal standard.

It is best to aspirate the cleaning fluid for a few minutes, followed by deionised water for about 15 minutes to warm up and stabilise the flame photometer. For every 3 samples, the instrument
is calibrated against the standard Li solution and de-ionised water blank. Duplicate readings are taken for each sample, more if the readings do not agree within 1%, with the average taken as the result. In house standard of either Biotite Mo 40 (high K content) or Teneriffe 1725 Basalt (low K content) are also analysed in duplicate for comparison and instrumentation check.

The resulting read out can be transformed to percentage of potassium in sample using the following equation:

\[
\frac{\text{Reading} \times 2 \text{ (for } 1 \text{ lithium : 1 solution mixture)} \times \text{volume of mixture in flask (in ml)}}{\text{Sample weight (in milligrams)} / \text{volume of aliquot tested (in ml)}} = \% \text{ K in sample}
\]

### D.2.3 Argon Isotopes Determination

Argon concentrations are determined by means of Isotope Dilution Analysis. External Radio-Frequency Induction heating is used to fuse the sample, mounted in a molybdenum crucible, inside a vacuum extraction line. After a known amount of \(^{38}\text{Ar}\) spike is introduced to the system, non-noble gases are trapped by a combination of the cooling and heating of the titanium sponge and a cold bath provided by liquid nitrogen. The resultant clean inert gases (mainly argon) are analysed with a "Micromass 1200" mass spectrometer linked to an IBM PC 1.1 with VG 5400 Mass Spectrometer software. Detailed discussion of the mass spectrometer methodology can be found in Dalrymple & Lanphere (1969) and Faure (1986). The mass spectrometer utilised is a magnetic deflection mass spectrometer with 60° magnetic sector, sensitive for \(3 \times 10^{-4}\) Amps/Torr. Fig.D.1 is a schematic diagram of the mass spectrometer and argon extraction line set up.

The \(^{40}\text{Ar}/^{38}\text{Ar}\) and \(^{36}\text{Ar}/^{38}\text{Ar}\) ratios are calculated based on 20-50 points time-variation measurement. A discrimination correction ratio (300.4) is applied to both \(^{36}\text{Ar}/^{38}\text{Ar}\) and \(^{40}\text{Ar}/^{38}\text{Ar}\) values, to standardise the equipment read out with an international standard, based on current atmospheric argon ratios. Age calculation is done using constants provided by Steiger & Jäger (1977).

Quality assurance is obtained by periodically running in house NIGL sample biotite Mo40 and occasionally with an International Standard GL-0 glauconite (24.8 nl/g Ar). Duplicate analyses on selected samples were performed.

### D.2.4 Ar Step Heating Procedure

Purified and weighed samples (~0.1 g for amphibole, ~0.05 g for mica) are wrapped in high purity
Fig.D1 Schematic diagram of Micromass 1200 Mass Spectrometer
Isotopic Dating

aluminum foil and loaded into Spectrosil phials. They are then irradiated in the Ford Reactor, Ann Arbor Michigan, USA. International standard MMHb-1, HB-3GR and Leeds University internal standard FY12aHnb are used as the flux monitors. The samples are loaded into the arms of a glass storage tree above the furnace and the entire system is baked overnight at 125°C. For analysis, the sample is dropped into the furnace and the step heating begins. The temperature is monitored using an infrared optical pyrometer. Heating steps are of 30 minutes duration and the furnace allowed to cool for 10 minutes between steps. The liberated gas is cleaned using two successive getters (a mixture of Ti-Zr metal shavings and Ti sponge) heated to 800°C and allowed to cool. The clean gas is collected in a small volume inlet section, by absorption on charcoal at liquid nitrogen temperature, before being introduced to the mass spectrometer. Argon isotope analyses are performed using a MAP 215 mass spectrometer with Nier type source, Faraday collector and magnet peak jumping under computer control. Ion beams are detected by a Cary 401M electrometer. The $^{36}\text{Ar}/^{38}\text{Ar}$ ratio is calculated on 11 cycles time-variation measurements. $^{37}\text{Ar}$ and $^{38}\text{Ar}$ are also measured to correct the Ca and Cl reactions respectively. The argon ratio, age and errors for each gas fraction are calculated utilising formulae after Dalrymple & Lanphere (1971) using a computer programme written by P.G.Guise. Constants are after Steiger & Jäger (1977).

Details of the method can be obtained from Dalrymple & Lanphere (1971), Dallmeyer (1979) and McDougall & Harrison (1988).

D.2.5 Isochrons

Two types of isochrons can be plotted for K-Ar data to gain a weighted average of a group of rocks which are either cogenetic or are affected by a thermal event which completely reset their K-Ar clocks. These are the $^{40}\text{K}/^{36}\text{Ar}$ v.s. $^{40}\text{Ar}/^{38}\text{Ar}$ (Hayatsu & Carmichael, 1970; Roddick & Farrar, 1971; Hayatsu & Palmer, 1975; Fitch et al., 1976; Hayatsu & Carmichael, 1977) and the $^{40}\text{K}$ v.s. $^{40}\text{Ar}^*$ (radiogenic $^{40}\text{Ar}$) (Harper, 1970) isochrons. An isochron age can be obtained from the slope of the best fit linear regression line (here the regression line is calculated using QuattroPro spread sheet software). More emphasis is put on data points with higher $^{40}\text{Ar}/^{36}\text{Ar}$ and $^{40}\text{Ar}^*$ (lower atmospheric argon component). The intercept on the $^{40}\text{Ar}/^{36}\text{Ar}$ axis should be close to the present $^{40}\text{Ar}/^{38}\text{Ar}$ ratio (295.5) used to correct the $^{40}\text{Ar}_{\text{atm}}$ (atmospheric $^{40}\text{Ar}$) component. Significant deviation from this value indicates incorrect individual ages, although the isochron age may still be correct.

The regression line on the $^{40}\text{K}$ v.s. $^{40}\text{Ar}^*$ isochron should pass through the origin. A negative $^{40}\text{Ar}^*$ intercept indicates argon loss, whereas a positive $^{40}\text{Ar}^*$ intercept indicates the presence of excess Ar. These two isochrons are plotted for suites rocks with of three or more K-Ar analyses.
D.3 RESULTS AND DISCUSSION OF ERROR

D.3.1 Introduction

All K-Ar ages are within 95% confidence level (2σ). For the potassium analysis, two sources of error are known: [1] geological causes: i.e. leaching of K, K fractionation and K heterogeneity in the different minerals in the sample (specifically for whole rock analysis), [2] analytical and instrumentation error. The error in the potassium result is calculated from the standard error of the mean (1σ), reported with a minimum of 1% for accuracy purposes. When compared with K content obtained either from XRF (for whole rock) or from microprobe (mineral separate), the flame photometry results tend to be lower by ~20%. The discrepancy in the whole rock result may be due to the different fraction used in analysis; 125-500 μm for flame photometry, whereas rocks are powdered to a clay size fraction for XRF analysis. Potassium variation with sample size has been noted before (e.g. Chen et al., 1984; Jäger et al., 1985) and may be attributed to fractionation during size separation. Differences with mineral chemistry results may be due to different phases of amphiboles, zonation within a mineral phase and impure separation. In all results, however, there are very good reproducibility in the flame photometry result (mostly < 1%, all results < 9% error). Further correction is, therefore, not necessary.

Argon analysis will have 3 major sources of error:

1. Geological causes: i.e. high atmospheric argon contamination which is problematical in young samples, however this will be minimised with the $^{36}\text{Ar}/^{38}\text{Ar}$ correction. Sample heterogeneity, especially in whole rock analysis, should be taken into account in interpreting the data. Since argon is a noble gas, thus does not bond with other atoms in the lattice, argon loss is possible in the following settings (Faure, 1986):
   a. Where crystal lattices cannot retain argon, even at low temperature and pressure.
   b. Where rocks have been partially or completely melted, followed by crystallisation of new minerals from the melt.
   c. Where rocks have undergone subsequent heating (metamorphism and/or metasomatism).
   d. Chemical weathering and alteration by aqueous fluids, changing both the K and Ar content in the rocks.
   e. Solution and redeposition of water soluble minerals.
   f. Mechanical breakdown of minerals, radiation damage and shock waves.
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2. Physical causes: e.g. excessive grinding (increases the atmospheric argon content due to large surface area and/or loss of argon from the lattice (Faure, 1986); grains spilling out of crucible during heating (normally can be avoided by heating the sample slowly) and the presence of atmospheric argon, due to leakage, in the mass spectrometer system.

3. Analytical and instrumentation error.

When the $^{40}\text{Ar}^*$ is below the detection limit of the mass spectrometer, the maximum age provided is calculated on the assumption that $^{40}\text{Ar}^*$ is 1% of the $^{40}\text{Ar}_{\text{tot}}$ (in most cases when $^{40}\text{Ar}^*$ is low, it is <1% of $^{40}\text{Ar}_{\text{tot}}$). This assumption is usually reliable, except when the $^{40}\text{Ar}_{\text{tot}}$ is extremely high, which normally indicates heterogeneity (mixture with low K phases) or leakage in the extraction line.

Errors in the isochron diagrams are difficult to assess, as they will depend on the error in the $^{40}\text{K}/^{40}\text{Ar}$, $^{40}\text{Ar}/^{39}\text{Ar}$, and the $^{40}\text{Ar}^*$ results; these errors will give a maximum and a minimum steepness of slope that can be drawn about the data points, with the isochron age represented by the best fit line. As the isochron diagrams serve as a weighted average, an average of the two isochron ages and the difference between each isochron age with the average, can be used as a mean isochron age and the error.

All samples are *in situ* unless stated. Fig.D.2 is a map of the sample localities. Table D.1 summarises the K-Ar results from the Bacan region. Ar-Ar step heating results have been discussed in Chapter Three.

**SIBELA METAMORPHIC COMPLEX**

**CONTINENTAL AFFINITY**

**Sample:** B102  
**Location:** S. Ra, Central Bacan.  
**Lithology:** Float of epidote, biotite and sphene bearing quartz-feldspathic schist.  
**K-Ar Age:** A mica concentrate (~80% pure) is analysed. The $^{40}\text{Ar}^*$ was below the limit of detection of the machine. The maximum age calculated is 0.21 Ma.  
**Comments:** This rock is part of the continental basement interpreted to be Palaeozoic in age (Hamilton, 1979). The radiometric age is considerably lower than the expected age, attributed to the low $^{40}\text{Ar}^*$ component. Two possible causes are: recent resetting (loss of $^{40}\text{Ar}^*$) or contamination by hydrothermal fluid rich in $^{36}\text{Ar}$.  

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Fig. D.2 Location of Samples for Isotopic Dating from Bacan.
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Sample: B22
Location: S. Ganda Suli, Central Bacan.
Lithology: Float of staurolite, garnet mica schist.
Ar-Ar Age: -0.3 ± -0.8 Ma.
Comments: The $^{40}$Ar* is negative, giving a negative age, which is further evidence for $^{36}$Ar enrichment.

OPHIOLITIC AFFINITY
Sample: BM572
Location: S. Ra.
Lithology: Amphibole cumulate float containing amphibole, plagioclase (An$_{95-100}$), opaque, epidote, chlorite, muscovite and sphene.
K-Ar Age: The age derived from the (90%) amphibole separate is 46.4 ± 3.5 Ma (Middle Eocene).
Comments: This age is younger than the expected age (Palaeozoic or Cretaceous). It is interpreted to be the metamorphic age, which corresponds to the volcanism in the region.

Sample: B32
Location: S. Ganda Suli.
Lithology: Amphibolite float, next to in situ amphibolite. The rock show is mylonitized and consists of amphiboles, plagioclase (An$_{38-67}$), opaque, chlorite and sphene.
K-Ar Age: The age of the amphibole concentrate (~95% pure) is 90.6 ± 10.9 Ma (Late Cretaceous).
Ar-Ar Age: The Ar-Ar spectrum shows two possible plateaux, one at 37.6 ± 9 Ma (51.2% $^{39}$Ar) and the second at 99.5 ± 3.7 Ma (33.6% $^{39}$Ar).
Comments: This is the only sample analysed by both methods. The U-shaped spectrum is indicative of excess argon from the mantle, which means that the minima is slightly older than the age during closure temperature (Lanphere & Dalrymple, 1976). Theoretically there is only one Ar-Ar plateau at 38 Ma (>50% $^{39}$Ar released), however the high temperature record is consistent with the K-Ar age and the age derived from B97.

The high temperature age is interpreted to be possible crystallisation age, due to similarity with ages derived from the Eastern Halmahera ophiolite. The low
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temperature age is interpreted to be related to the Late Oligocene volcanism and extensional phase.

Sample: B97
Location: S. Ra.
Lithology: Metamorphosed peridotite cumulate with amphibole, minor opaque, epidote, chlorite and sphene.
Ar-Ar Age: The U-shaped Ar-Ar spectrum of the amphibole concentrate (>95% pure) forms a minima at 98.0 ± 12.0 Ma.
Comments: This age is consistent with sample B32 and the Eastern Halmahera Ophiolite and therefore interpreted to be a high temperature, possibly crystallisation age. The U-shape spectrum is indicative of excess argon, and the age calculated is therefore slightly older than time of closure (Lanphere & Dalrymple, 1976).

Sample: B103
Location: S. Ra.
Lithology: Metaperidotite cumulate containing clinopyroxene, green amphibole and opaque.
Ar-Ar Age: The Ar-Ar spectrum of the amphibole concentrate (>95% pure) is U-shaped with a minimum at 46.0 ± 12.0 Ma (59.4% ⁴⁰Ar).
Comments: The age is interpreted as metamorphic age related to either the Late Oligocene volcanism and extensional phase or the 45 Ma Pacific Plate reorganisation.

Sample: BM571
Location: S. Ra.
Lithology: Cumulate hornblendite with minor intercumulate plagioclase and opaque.
Ar-Ar Age: The Ar-ar spectrum is U-shaped with a minimum at 56.0 ± 8.0 Ma.
Comments: The age is interpreted as metamorphic age related to either the Late Oligocene volcanism and extensional phase or the 45 Ma Pacific Plate reorganisation.

SALEH METAMORPHIC COMPLEX
Sample: SM12
Location: P. Saleh Besar.
Lithology: Metabasite collected from a large angular boulder, near in situ. The rock includes plagioclase (An₉₂), clinopyroxene, opaque, prehnite, pumpellyite. Chlorite and Smectite replaces the matrix.
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K-Ar Age: The whole rock sample yields an age of 12.0 ± 0.6 and 10.9 ± 0.8 Ma (Middle Miocene).

Comments: Based on stratigraphic arguments this sample is predicted to be older than Miocene. This age is interpreted to be the age of metamorphism, possibly related to dioritic intrusion (SR2 and SR6).

BACAN FORMATION

Sample: BR73
Location: Northern coast of Bacan.
Lithology: A feldspar-phyric basalt containing plagioclase (An$_{10-18}$), clinopyroxene, opaque, epidote, pumpellyite, prehnite and stilbite. The matrix is replaced by smectite.
K-Ar Age: The age of the whole rock sample is 11.6 ± 8.3 Ma (Middle Miocene). A duplicate analysis yielded 14.94 ± 2.26 Ma.
Comments: The large error in the first analysis is attributed to the large $^40$Ar$_{em}$ component. The second analysis used larger grain size, thus reducing the $^40$Ar$_{em}$ component. Two stages of heating were employed for the second analysis, and the age derived is a combination of the two heatings. Error calculated is based on the formulae: \[ \text{Sum of Error} = \left( \frac{\text{Error}_1^2 + \text{Error}_2^2}{2} \right)^{1/2} \] The age of this formation is Late Eocene based upon micropalaeontological evidence. The younger radiometric age is ascribed to a resetting event, during the sub-greenschist facies metamorphism. This is consistent with the fact that stilbite is the K-bearing phase in this rock.

Sample: BM433
Location: Air Kabur near S. Kusu-Bibi.
Lithology: Feldspar-phyric basaltic-andesite containing plagioclase (An$_{2-16}$), clinopyroxene, opaque, secondary carbonate and zeolite. The matrix is glass and smectite.
K-Ar Age: The whole rock age is 14.8 ± 1.2 Ma (Middle Miocene).
Comments: This age is interpreted to be metamorphism age, consistent with that of BR73. This is supported by the fact that the K-bearing phases (smectite and zeolite) are both metamorphic minerals.

SOUTH BACAN FORMATION

Sample: BM250
Location: S. Bibinoi.
Lithology: Float sample of amphibole-rich andesite which includes pyroxene, amphibole,
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opaque and apatite phenocrysts. Biotite replaces amphibole, matrix is glass.

K-Ar Age: The amphibole mineral separate age is 20.8 ± 2.0 Ma.
Comments: This age is consistent with the biostratigraphic age (Early Miocene) therefore is interpreted to be formation age.

Sample: BM188
Location: S. Silang.
K-Ar Age: The age of the whole rock sample is 7.50 ± 1.12 Ma (Late Miocene).
Comments: This age is younger than the biostratigraphic age. It is interpreted as a partially or completely reset age, associated with the Neogene volcanism.

KAPUTUSAN FORMATION
TWO PYROXENE ANDESITE (TPAN)
Sample: BR58
Location: Northeastern corner of Bacan.
Lithology: Fresh TPAN clast in the volcanic breccia. The rock comprises plagioclase (An


K-Ar Age: The first analysis used a pyroxene concentrate. The $^{40}$Ar* was below the detection limit of the mass spectrometer. The maximum age calculated is 6.59 Ma. A duplicate analysis was run using the whole rock fraction. The age derived is 6.70 ± 0.28 Ma (Late Miocene).
Comments: The pyroxene separate is poor in K, therefore has virtually no $^{40}$Ar*. The second analysis used the whole rock sample, which has higher K content, indicating that the K-bearing phase is the glass. The age is interpreted to be formation age and is consistent with the stratigraphic age.

Sample: BR16
Location: Tg. Sepi, eastern coast of North Bacan.
Lithology: Fresh TPAN clast in the pyroclastic unit. The rock consist of complex zoned and twinned plagioclase (An$_{55-76}$), clino- and orthopyroxene and opaque phenocrysts in a glassy matrix.
K-Ar Age: A large amount of $^{40}$Ar$_{sat}$ in the whole rock analysis, causes a large error in the $^{40}$Ar* so that $^{40}$Ar* is unreliable. The maximum age of the sample is 8.7 Ma. A
duplicate analysis was run, using the same grain size, with a larger amount of sample and without quartz wool. This effort was in vain, as the error in the $^{40}\text{Ar}^*$ was still too large, yielding a maximum age of 3 Ma. A third analyses used a larger grain size, to reduce the $^{40}\text{Ar}_\text{adm}$, and resulted in an age of $5.85 \pm 0.79$ Ma.

Comments: The three runs of this sample show that the larger grain size does contribute to lowering the $^{40}\text{Ar}_\text{adm}$ component. This has the effect of increasing the relative $^{40}\text{Ar}^*$ component, thus improving the error of $^{40}\text{Ar}^*$. The larger grain size used should have little effect on heterogeneity because most of the K is held in the fresh glass which cooled rapidly having effectively a single blocking temperature. The other phases are all K-poor, therefore will have little bearing on the analysis. This age is interpreted to be formation age and is consistent with the stratigraphic age.

Sample: BM361
Location: S. Goro-goro, inland from where BR16 was collected.
Lithology: Fresh TPAN clast in the pyroclastic deposit. The rock contains plagioclase ($\text{An}_{50}$, $\text{An}_{80}$), clino- and orthopyroxene and opaque phenocrysts in a glassy matrix.
K-Ar Age: The whole rock sample yields an age of $5.52 \pm 0.76$ Ma (Late Miocene-Early Pliocene).
Comments: This age is interpreted to be the formation age and is consistent with stratigraphic evidence. This age is within error of that from BR16.

Sample: BT36A
Location: P. Waring, south of P. Kasiruta.
Lithology: Fresh basaltic clast in the pyroclastic deposit. The rock consist of plagioclase ($\text{An}_{54-76}$), clinopyroxene and opaque phenocrysts in a glassy matrix.
K-Ar Age: The first analysis was done using the whole rock sample. $^{40}\text{Ar}^*$ was below the limit of detection of the machine, with a maximum age of 10.37 Ma. The second analysis used a larger amount of the whole rock sample. The age derived is $4.91 \pm 1.37$ Ma (Late Miocene-Early Pliocene).
Comments: These two analyses show that increasing sample weight can help the analysis. The larger sample means larger $^{40}\text{Ar}^*$, which pushes the $^{40}\text{Ar}^*$ above the limit of detection of the machine. This age is interpreted to be the formation age and is consistent with the stratigraphic age.
HORNBLENDE PYROXENE ANDESITE (HPAN)

Sample: BR49
Lithology: Clast of HPAN from a boulder of volcanic breccio-conglomerate. The rock is composed of plagioclase (An_{40-42}), orthopyroxene, hornblende, apatite and opaque phenocrysts in a glassy matrix.

K-Ar Age: The whole rock age is 7.47 ± 0.97 Ma (Late Miocene). Duplicate analysis of a hornblende and pyroxene separate was conducted. $^{40}\text{Ar}^*$ was below the detection limit of the machine. The maximum age derived is 8.81 Ma, consistent with the whole rock result. An attempt to lower the $^{40}\text{Ar}_{\text{en}}$ content, thus increase the $^{40}\text{Ar}^*/^{39}\text{Ar}_{\text{en}}$ ratio, was made by heating the sample twice. Both analyses resulted in $^{40}\text{Ar}^*$ below the limit of detection, suggesting that some of the $^{40}\text{Ar}^*$ was released at the first heating which is at ~650°C. The maximum age of this sample is calculated to be 10.64 Ma, which is consistent with the whole rock result.

Comments: These analyses show that the whole rock result is more reliable in the case of fresh, young volcanic. This might be due to the hydrous nature of amphibole, which is prone to alteration and Ar loss. The straight fit result, in both $^{40}\text{Ar}/^{38}\text{Ar}$ and $^{36}\text{Ar}/^{38}\text{Ar}$ ratios plot against time, are discarded due to incompatibility with stratigraphic age, and the maximum ages from the duplicate and triplicate analyses. The analyses also show that multiple heating procedure, without an ability to produce low temperature, is not very useful in reducing $^{40}\text{Ar}_{\text{en}}$. This age is interpreted to be the formation age and is consistent with biostratigraphic dates.

Sample: BM26
Location: Northern P. Mandioli.
Lithology: HPAN clast, collected from a volcanic breccia deposit. The rock contains plagioclase (An_{40-49}), clinopyroxene, amphibole, apatite and opaque phenocrysts in a glassy matrix.

K-Ar Age: A separate of 90% hornblende and pyroxene was analysed. The calculated age is 4.97 ± 0.46 Ma (Late Miocene-Early Pliocene).

Comments: This age is consistent with BT36A, BR203 and the biostratigraphic ages, and therefore is interpreted to be a formation age.

Sample: BR203
Location: P. Waring.
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Lithology: HPAN clast in the pyroclastic deposit, stratigraphically above BT36A. The rock consists of plagioclase (An_{48-66}), clinopyroxene, pargasite, apatite and opaque in a glassy matrix.

K-Ar Age: The pargasite and clinopyroxene concentrate (~90% pure) yields an age of 4.23 ± 1.13 Ma (Late Miocene-Early Pliocene).

Comments: This age is consistent with the ages from BT36A and BM26 (south of P. Waring). The age is interpreted to be the formation age and is compatible with biostratigraphic dates.

Sample: BM453
Location: P. Mamalayu, west of P. Waring.
Lithology: HPAN clast in the pyroclastic deposit. The rock consist of plagioclase (An_{49-89}), amphibole, clinopyroxene, apatite and opaque phenocrysts in a glassy matrix.
K-Ar Age: The hornblende concentrate was analysed. The ^{40}Ar* was below the detection limit of the machine. The maximum age is 2.78 Ma. The whole rock sample was analysed as a duplicate. The calculated age is 2.71 ± 0.21 Ma.
Comments: Although the age is considerably lower than the rocks from P. Waring (~5Ma), it is still consistent with the stratigraphic age and therefore it is considered to be the formation age.

Sample: BM131
Location: S. Kaputusan in northern Bacan.
Lithology: Massive HPAN, possibly lava flow, containing plagioclase (An_{46-63}), clinopyroxene, amphibole and opaque in a glassy matrix. Heulandite is present.
K-Ar Age: The hornblende and pyroxene concentrate was analysed. The error in the ^{40}Ar* is too large to be reliable. The maximum age of this sample is 1.13 Ma. A duplicate analysis using the whole rock sample yields a 2.54 ± 0.32 Ma age.
Comments: The whole rock analysis gave a better result than the mineral separate. This is mainly due to the fact that the whole rock is more K-rich than the mineral separate and therefore contains more ^{40}Ar*.

Sample: BP30
Location: South of S. Kaputusan, across P. Obit.
Lithology: HPAN boulder within a poorly consolidated, medium grained sandstone (?pyroclastic). The rock contains plagioclase (An_{44-77}), clinopyroxene, amphibole,
Isotopic Dating

and opaque phenocrysts in a glassy and smectite matrix.

**K-Ar Age:** A whole rock analysis of this sample shows that the \(^{40}\text{Ar}^*\) is below the detection limit of the machine. The maximum age derived is 2.17 Ma. A duplicate analysis was run using the whole rock sample, yielding a 2.53 ± 0.37 Ma age (Pliocene).

**HORNBLENDE ANDESITE (HBAN)**

Sample: BR174
Location: P. Waring (as BT36A & BR203).
Lithology: HBAN float consisting of plagioclase (An\(_{46.6}\)), pargasite and pargasitic hornblende and apatite in a glassy matrix.

**K-Ar Age:** The age derived using the amphibole concentrate (~90% pure) is 5.06 ± 0.47 Ma (Late Miocene–Early Pliocene).

**Comments:** This age is within error with BR203 and BT36A and is consistent with the stratigraphic age and therefore is considered to be the age of formation.

Sample: BR82
Location: Northwestern corner Bacan, west of BR58.
Lithology: HBAN clast in the pyroclastic deposit. It includes plagioclase (An\(_{47.7}\)), tschermakitic and edinitic hornblende and opaque phenocrysts in a glassy matrix.

**K-Ar Ages:** A hornblende concentrate was analysed. The large error of the 2.01 ± 1.83 Ma age is due to the low \(^{40}\text{Ar}^*\) and its large error. A duplicate analysis is run using the whole rock sample, giving 2.23 ± 0.13 Ma.

**Comments:** This shows that the whole rock sample is more reliable than the mineral separate, mainly because it is more K-rich (thus \(^{40}\text{Ar}^*\)-rich). Although the age is considerably different from BR58, it is consistent with stratigraphic age and is interpreted to be the age of formation.

Sample: BM135
Location: S. Kaputusan.
Lithology: HBAN (green hornblende) with amphibole xenocrysts. The rock contains plagioclase (An\(_{50.7}\)), amphibole, apatite and opaque phenocrysts in a glassy groundmass. Stilbite and calcite replace plagioclase, while quartz is a product of devitrified glass.

**K-Ar Age:** A concentrate of the green hornblende was analysed. The \(^{40}\text{Ar}^*\) was below the detection limit of the machine. The maximum age of this sample is 1.75 Ma. A
duplicate analysis using whole rock yields an age of 0.904 ± 0.323 Ma.

Comments: This again shows that whole rock result is better than the mineral separate. The age of this rock could represent the age of alteration as the glass has been affected by low-temperature metamorphism.

HORNBLENDE BIOTITE PYROXENE ANDESITE (HBIA)

Sample: BR197
Location: North of Kaputusan.
Lithology: HBIA clast in a pyroclastic deposit. It includes plagioclase (An_{45-70}), amphibole, biotite and opaque phenocrysts in a glassy matrix. Quartz is present as a byproduct of devitrified glass.
K-Ar Age: The age of the whole rock sample is 5.53 ± 2.68 Ma (Late Miocene-Early Pliocene). For duplication, a hornblende separate was analysed. The error in the "Ar* is too large which makes the age unreliable. The maximum age derived from this sample is 0.36 Ma. This age is discarded, due to its high "Ar* component. This age is therefore interpreted with caution.

Comments: This sample again shows that the whole rock result is better than the mineral separate. The large error in the age is a result of large "Ar* component. This age is therefore interpreted with caution.

Sample: BM256
Location: S. Gofio, south Bacan.
Lithology: Float of HBIA containing amphibole, biotite, pyroxene, apatite and opaque phenocrysts in a glassy matrix.
K-Ar Age: The whole rock analysis yield a 4.78 ± 0.30 Ma.
Comments: This sample is the only one with large discrepancy in the K content analyses between the flame photometry and the XRF results (1.632 and 4.691 %K respectively). Although the resulting age is consistent with the stratigraphic age, it should be taken with caution.

Sample: BR53
Location: P. Gandaha (across from BR49).
Lithology: Massive HBIA lava flow, consisting of plagioclase (An_{46-62}), amphibole, biotite, orthopyroxene and opaque phenocrysts in a glassy matrix.
K-Ar Age: The first whole rock analysis shows that the "Ar* was below the limit of detection of the machine. The maximum age calculated is 7.026 Ma. Duplicate
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analysis was performed using the hornblende concentrate. The $^{40}\text{Ar}^*$ was still below the limit of detection of the machine. The maximum age calculated is 1.644 Ma. A triplicate using a larger size sample was performed resulting in a $0.457 \pm 0.139$ Ma age.

Comments: This sample shows that for a young sample, the amount of $^{40}\text{Ar}_\text{adm}$ is critical in determining the age, as the amount of $^{40}\text{Ar}^*$ is low. This age is considerably lower than the BR49 age ($7.47 \pm 0.97$ Ma) collected across the island and therefore may be a reset age due to later volcanism.

INTRUSIVE ROCKS

Sample: BR80
Location: P. Nusa Babi.
Lithology: Quartz monzodiorite containing plagioclase (An$_{55}$,g), clino- and orthopyroxene, amphiboles, biotite, opaques and zircon.
K-Ar Age: Two analyses were performed on this sample, the first using a hornblende and biotite concentrate yielded $28.6 \pm 9.5$ Ma. The second analysis used ~100% biotite separate. The age of the sample is $19.8 \pm 1.6$ Ma (Early Miocene).
Comments: The large error in the first analysis is attributed to sample heterogeneity. The duplicate analysis yielded an age which is within error of the previous analysis, and consistent with field observation (pre-Middle Miocene Ruta Fm.).

Sample: BM285
Location: S. Rain.
Lithology: Hornblende bearing quartz monzodiorite (Nusa Babi Intrusive), intruding the South Bacan Formation.
K-Ar Age: The hornblende concentrate (~90% pure) was analysed. The $^{40}\text{Ar}^*$ is below the detection limit and the calculated maximum age is 6.16 Ma.
Comments: The large $^{40}\text{Ar}_\text{adm}$ component resulted in the large error in the $^{40}\text{Ar}^*$, which makes it unreliable, hence the maximum age. The large error in the $^{40}\text{Ar}^*$ indicates either the sample is altered or there was a leakage in the Ar extraction line.

Sample: BR267
Location: Northern foothill of Sibela Mountain (Babang area), Central Bacan.
Lithology: Biotite-epidote bearing quartz diorite float.
K-Ar Age: The age derived from the ~100% biotite separate is $2.24 \pm 0.89$ Ma (Pleistocene-
Isotopic Dating

Pliocene).

Comments: The mineralogical similarity with the Nusa Babi Intrusive suggests that this is part of the same intrusion (~20Ma). The younger age is interpreted to be the age of metamorphism, related to the Quaternary volcanism in the area. This is supported by the presence of metamorphic minerals (albite, epidote).

Sample: SR2
Location: P. Saleh Kecil.
Lithology: Hornblende bearing diorite dyke cutting through the Saleh Metamorphic Complex. The rock comprises plagioclase (An\textsubscript{82-91}), amphiboles, opaques, muscovite, epidote and zeolite.
K-Ar Age: The amphibole separate (~90% pure) was analysed. The age derived is 10.8 ± 8.7 Ma (Late Miocene).
Comments: The large error in the age is attributable to the low \(^{40}\text{Ar}\)\(^*\) and its associated large error.

Sample: SR6
Location: P. Saleh Kecil.
Lithology: Diorite, intruding SR2, with a strong fabric showing hornblende and epidote segregation zones. The rock includes plagioclase (An\textsubscript{82-91}), pyroxene, amphibole and opaque. Chlorite and muscovite replace amphiboles.
K-Ar Age: The hornblende separate (~90% pure) was analysed. The age derived is 9.29 ± 1.58 Ma (Middle-Late Miocene).
Comments: This age is consistent with field observation (younger than SR2). Although the rock has a strong fabric, its petrography suggests magmatic origin. This age is therefore interpreted to be crystallisation age.

Sample: B66
Location: Southwest corner of the Sibela Mountains.
Lithology: A microdiorite dominated by plagioclase and green amphiboles.
K-Ar Age: The whole rock age is 15.1 ± 1.6 Ma (Middle Miocene).
Comments: This age is interpreted to be a resetting age associated with Kaputusan volcanism.
### Table I. Summary result of the K-Ar radiometric dating in the Bacan Region.

<table>
<thead>
<tr>
<th>Fm</th>
<th>Sample#</th>
<th>Mtrl</th>
<th>Grain size (microns)</th>
<th>% K (1σ error in %)</th>
<th>Wt for Ar (g)</th>
<th>$^4\text{Ar}_{in}$ (ng)</th>
<th>$^4\text{Ar}^*$ (ng/g) (1σ error in %)</th>
<th>$^4\text{Ar}_{ex}$ (in %)</th>
<th>Calculated Age (Ma, error 2σ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiMC</td>
<td>B102</td>
<td>mica</td>
<td>125-250</td>
<td>4.466 ± 1.00</td>
<td>0.0821</td>
<td>0.30</td>
<td>-0.1285 ± 0.15 %</td>
<td>&lt; lod</td>
<td>&lt; 0.209</td>
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<tr>
<td>SiMO</td>
<td>BM572</td>
<td>amph</td>
<td>125-250</td>
<td>0.231 ± 3.46</td>
<td>1.1672</td>
<td>0.60</td>
<td>0.4218 ± 1.59 %</td>
<td>54.95</td>
<td>46.40 ± 3.50</td>
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<tr>
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<td>B32</td>
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<td>125-250</td>
<td>0.467 ± 3.43</td>
<td>0.2079</td>
<td>1.75</td>
<td>1.6857 ± 5.16 %</td>
<td>83.34</td>
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<tr>
<td>SaMB</td>
<td>SM12</td>
<td>wr</td>
<td>125-250</td>
<td>0.842 ± 1.00</td>
<td>0.9867</td>
<td>0.75</td>
<td>0.3938 ± 2.20 %</td>
<td>65.91</td>
<td>12.00 ± 0.60</td>
</tr>
<tr>
<td>SaMB</td>
<td>SM12</td>
<td>wr</td>
<td>125-250</td>
<td>0.934 ± 1.00</td>
<td>1.9346</td>
<td>2.51</td>
<td>0.3977 ± 3.46 %</td>
<td>76.53</td>
<td>19.00 ± 0.80</td>
</tr>
<tr>
<td>BFm</td>
<td>BR73</td>
<td>wr</td>
<td>125-250</td>
<td>0.213 ± 1.00</td>
<td>1.0016</td>
<td>3.14</td>
<td>0.0966 ± 35.68 %</td>
<td>97.01</td>
<td>11.60 ± 8.30</td>
</tr>
<tr>
<td>BFm</td>
<td>BR73dp</td>
<td>wr</td>
<td>250-425</td>
<td>0.212 ± 4.30</td>
<td>1.0198</td>
<td>0.69</td>
<td>0.0832 ± 8.35 %</td>
<td>88.99</td>
<td>10.00 ± 3.50</td>
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<tr>
<td>BFm</td>
<td>BM433</td>
<td>wr</td>
<td>250-500</td>
<td>2.100 ± 2.68</td>
<td>1.0262</td>
<td>3.39</td>
<td>1.2124 ± 2.93 %</td>
<td>73.13</td>
<td>14.80 ± 1.20</td>
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<tr>
<td>BFm</td>
<td>BM250</td>
<td>hb</td>
<td>125-250</td>
<td>0.218 ± 3.45</td>
<td>1.0275</td>
<td>0.50</td>
<td>0.1768 ± 3.39 %</td>
<td>73.34</td>
<td>20.80 ± 2.00</td>
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<tr>
<td>BFm</td>
<td>BM188</td>
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<td>250-500</td>
<td>1.289 ± 2.27</td>
<td>1.0236</td>
<td>0.43</td>
<td>0.3767 ± 7.13 %</td>
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<td>7.50 ± 1.12</td>
</tr>
<tr>
<td>TPAN</td>
<td>BR58</td>
<td>px</td>
<td>125-250</td>
<td>0.062 ± 6.99</td>
<td>0.8589</td>
<td>1.37</td>
<td>-0.051 ± 30.89 %</td>
<td>&lt; lod</td>
<td>&lt; 6.593</td>
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<tr>
<td>TPAN</td>
<td>BR58dp</td>
<td>wr</td>
<td>250-425</td>
<td>1.166 ± 1.08</td>
<td>2.2149</td>
<td>0.94</td>
<td>0.3044 ± 1.78 %</td>
<td>58.28</td>
<td>6.70 ± 0.28</td>
</tr>
<tr>
<td>TPAN</td>
<td>BR10</td>
<td>wr</td>
<td>125-250</td>
<td>1.209 ± 2.23</td>
<td>0.1179</td>
<td>4.85</td>
<td>0.5794 ± 98.85 %</td>
<td>98.61</td>
<td>&lt; 8.174</td>
</tr>
<tr>
<td>TPAN</td>
<td>BR16</td>
<td>wr</td>
<td>250-425</td>
<td>1.498 ± 5.31</td>
<td>1.0052</td>
<td>1.38</td>
<td>0.3414 ± 4.19 %</td>
<td>80.11</td>
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<tr>
<td>TPAN</td>
<td>BR16</td>
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<td>250-425</td>
<td>0.966 ± 1.29</td>
<td>0.3237</td>
<td>0.43</td>
<td>0.2073 ± 6.75 %</td>
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<td>BM361</td>
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<td>125-250</td>
<td>1.461 ± 1.37</td>
<td>0.1856</td>
<td>10.96</td>
<td>-0.2582 ± 77.33 %</td>
<td>&lt; lod</td>
<td>&lt; 10.370</td>
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<tr>
<td>TPAN</td>
<td>BR36A</td>
<td>wr</td>
<td>125-250</td>
<td>1.461 ± 1.37</td>
<td>0.6465</td>
<td>2.42</td>
<td>0.2792 ± 13.89 %</td>
<td>93.06</td>
<td>4.91 ± 1.37</td>
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<tr>
<td>TPAN</td>
<td>BM26</td>
<td>hb</td>
<td>125-250</td>
<td>1.098 ± 2.69</td>
<td>0.6240</td>
<td>0.46</td>
<td>0.2121 ± 3.73 %</td>
<td>77.58</td>
<td>4.97 ± 0.46</td>
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<tr>
<td>TPAN</td>
<td>BR203</td>
<td>px</td>
<td>125-250</td>
<td>1.890 ± 2.49</td>
<td>1.0420</td>
<td>4.05</td>
<td>0.3114 ± 13.13 %</td>
<td>92.59</td>
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<tr>
<td>TPAN</td>
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<td>125-250</td>
<td>0.449 ± 1.45</td>
<td>0.5609</td>
<td>2.72</td>
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<td>&lt; lod</td>
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<td>250-425</td>
<td>1.382 ± 1.56</td>
<td>2.2700</td>
<td>1.20</td>
<td>0.1166 ± 4.69 %</td>
<td>81.91</td>
<td>2.71 ± 0.21</td>
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<td>BM131</td>
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<td>125-250</td>
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<td>0.9990</td>
<td>2.30</td>
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<td>98.04</td>
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<td>1.544 ± 2.75</td>
<td>2.4686</td>
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<td>BP30</td>
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<td>0.387 ± 1.30</td>
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<td>1.088 ± 1.00</td>
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<td>0.662 ± 1.00</td>
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<td>2.19</td>
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<td>1.358 ± 1.56</td>
<td>1.0238</td>
<td>0.24</td>
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<td>0.390 ± 1.73</td>
<td>0.5014</td>
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<td>0.0321 ± 92.11 %</td>
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<tr>
<td>Fm</td>
<td>Sample#</td>
<td>Mtrl</td>
<td>Grain size (microns)</td>
<td>% K (1σ error in %)</td>
<td>Wt for Ar (g)</td>
<td>$^{39}\text{Ar}^{*}$ (nl/g) (1σ error in %)</td>
<td>$^{39}\text{Ar}_{\text{ex}}$ (in %)</td>
<td>Calculated Age (Ma, error 2σ)</td>
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<td>wr</td>
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<td>0.4664 ± 24.21 %</td>
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<td>BR197dp</td>
<td>hb</td>
<td>125-250</td>
<td>0.785 ± 1.21</td>
<td>0.2168</td>
<td>0.0150 ± 88.71 %</td>
<td>98.66</td>
<td>&lt; 0.363</td>
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<td>BM256</td>
<td>wr</td>
<td>250-500</td>
<td>1.632 ± 1.00</td>
<td>0.3493</td>
<td>0.3035 ± 2.98 %</td>
<td>72.16</td>
<td>4.78 ± 0.30</td>
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<td>BR53</td>
<td>wr</td>
<td>125-250</td>
<td>1.812 ± 1.00</td>
<td>0.1603</td>
<td>-0.7134 ± 48.08 %</td>
<td>&lt; lod</td>
<td>&lt; 7.026</td>
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<td>hb</td>
<td>125-250</td>
<td>1.418 ± 8.08</td>
<td>0.2390</td>
<td>-0.0517 ± 133.84 %</td>
<td>&lt; lod</td>
<td>&lt; 1.644</td>
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<td>BR53tp</td>
<td>wr</td>
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<td>1.318 ± 5.23</td>
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<td>0.0234 ± 14.28 %</td>
<td>93.13</td>
<td>0.457 ± 0.139</td>
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<td>Nbl</td>
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<td>hb+bt</td>
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<td>28.60 ± 9.50</td>
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<tr>
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<td>BR80dp</td>
<td>bt</td>
<td>125-250</td>
<td>4.291 ± 1.00</td>
<td>0.1024</td>
<td>3.3251 ± 3.87 %</td>
<td>68.74</td>
<td>19.80 ± 1.60</td>
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<td>125-250</td>
<td>0.407 ± 5.40</td>
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<td>0.1020 ± 1159.23 %</td>
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<td>500-1000</td>
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<td>SR2</td>
<td>amph</td>
<td>250-500</td>
<td>0.184 ± 1.00</td>
<td>0.2027</td>
<td>0.0774 ± 40.24 %</td>
<td>97.36</td>
<td>10.80 ± 8.70</td>
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<td>SR6</td>
<td>amph</td>
<td>125-250</td>
<td>0.364 ± 1.00</td>
<td>1.0611</td>
<td>0.1318 ± 8.49 %</td>
<td>89.20</td>
<td>9.29 ± 1.58</td>
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<tr>
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<td>wr</td>
<td>125-250</td>
<td>0.656 ± 1.00</td>
<td>2.2110</td>
<td>0.1306 ± 3.33 %</td>
<td>75.85</td>
<td>15.10 ± 1.60</td>
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</table>

**SI** - Sibela Metamorphic Complex, Continental  
**SIMO** - Sibela Metamorphic Complex, Ophiolite  
**SaMB** - Saleh Metamorphic Complex, Basite  
**BFm** - Bacan Formation  
**SiBFm** - South Bacan Formation  
**Nbl** - Nusa Babi Intrusive  
**SI** - Saleh Intrusive  
**TPAN** - Kaputusan Fm, two pyroxene andesite  
**HPAN** - Kaputusan Fm, hornblende pyroxene andesite  
**HBAN** - Kaputusan Fm, hornblende andesite  
**HBIAN** - Kaputusan Fm, hornblende, biotite, pyroxene andesite  

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Constants after Steiger & Jäger (1977)
APPENDIX E
CLASSIFICATION OF ROCKS FROM METAMORPHIC COMPLEXES

[1] Continental Metamorphic Rocks

A. Calc-quartzo-feldspathic gneisses
Quartzose feldspathic gneiss with strong foliation defined by micas (biotite ± muscovite) in a matrix of plagioclase and quartz associated with epidote and iron ore.

- B8A-8B augen sphene + apatite (+ retrograde chlorite); mylonitic texture
- B14 augen feldspar - muscovite (+ retrograde chlorite + calcite)
- B18 + amphibole
- B19 augen plagioclase - epidote + kyanite
- B23 + sphene porphyroblast
- B25 - epidote + amphibole (+ retrograde chlorite)
- B70 augen feldspar - muscovite - epidote
- B71-73
- B79 - muscovite + sphene + apatite (+ retrograde chlorite).
- B87 - muscovite
- B100 - biotite
- B102 augen plagioclase + apatite (+ retrograde chlorite)
- B106 - biotite
- BA14 - muscovite; mylonitic texture
- BA16 - muscovite + calcite (+ retrograde chlorite)
- BR269 + sphene porphyroblasts
- BT40A,D,E as BR269 (+ calcite + retrograde chlorite)

B. Graphitic garnet schists
Garnet and staurolite porphyroblasts in a quartz, biotite, muscovite, graphite and iron ore matrix. Garnet contains many inclusions (including biotite, muscovite and quartz), intimately associated with biotite and graphite. Staurolite normally streaked with graphite and/or iron ore. Staurolite porphyroblasts are subparallel while garnet porphyroblasts are oblique to fabric.

- B2-3 snow ball texture in garnet
- B15
- B17
- B20-21
- B22 + pyrophyllite (after kyanite) (+ retrograde chlorite)
- B74 + kyanite + plagioclase
APPENDIX E

B75-76
B78 + plagioclase; quartz-rich areas with subordinate micas
B82-83 rotated garnet
B84 + kyanite
B85-86
B105 zoned garnet + plagioclase; quartz-rich layers alternating with pelitic layers

C. Non-graphitic garnet schists
   B4 skeletal garnet + quartz + plagioclase + biotite + muscovite + epidote + zircon
   B7 metabasites: garnet (with amphibole and plagioclase inclusions) + amphibole + plagioclase + quartz + iron ore

D. Non-garnet schists
   B5 muscovite + biotite + quartz + plagioclase + tourmaline + iron ore
   B13 augen plagioclase + sphene + quartz + biotite + epidote + iron ore + amphibole + (+ calcite + chlorite in veins)
   B24 as B13
   BA15 metatuff: epidote + quartz + biotite + iron ore (+ retrograde chlorite)

E. Impure carbonates
   B1 calcite ± chondrodite (with talc, opaque and carbonate inclusions) porphyroblasts + talc + iron ore; foliation 270/25, defined by flattened minerals
   BT40B foliated quartzo-feldspathic biotite with marble bands: calcite + quartz + plagioclase + iron ore

A. Basic & ultrabasic rocks
   B26-27 serpentinised lherzolite: serpentinite (after olivine) + amphibole (tremolite after clinopyroxene) + talc + iron ore (+ calcite in veins); 315/78 foliation
   B34 serpentinised harzburgite with relict olivine + plagioclase
   B37 metagabbro: amphibole + plagioclase + iron ore + sphene
   B40 dunite: serpentinite + iron ore + calcite
   B41 contact between metagabbro and amphibolite + epidote
   B44 serpentinised gabbro: orthopyroxene + clinopyroxene + plagioclase + iron ore
   B45 metagabbro: amphibole + plagioclase + quartz + iron ore
APPENDIX E

B47 gabbroic: clinopyroxene (some chloritised and serpentinised) + epidote + plagioclase + quartz
B49 dunite: serpentinite + graphite/iron ore
B54 metagabbro: relict clinopyroxene + amphibole + plagioclase
B56 metadolerite: amphibole + plagioclase + quartz (+ chlorite in veins)
B77 serpentinised harzburgite with relict orthopyroxene
B81 metadolerite (fine grained amphibolite): quartz + hornblende
B91 metaperidotite: amphibole (after pyroxene) in a talc and serpentine matrix
B93 metagabbro: clinopyroxene + plagioclase
B94 cumulate gabbroic rock: green amphibole + plagioclase + epidote and opaque
B95 as B94
B97 metamorphosed peridotite (?wehrlite): possibly primary clinopyroxene + amphibole + plagioclase + iron ore (+ sphene + epidote segregations + chlorite)
B103 metamorphosed peridotite (?wehrlite) with high T granoblastic texture, possibly primary clinopyroxene + amphibole + iron ore
B104 harzburgite: olivine + orthopyroxene + altered plagioclase
BA17A metagabbro: amphibole + plagioclase + iron ore + retrograde chlorite + epidote + sericite (both after plagioclase) + calcite
BA35 layered metagabbro: pale amphibole + plagioclase
BR45 metagabbro: epidote (after clinopyroxene) + plagioclase + quartz + iron ore
BM569 as B26-27

B. Amphibolites

Interlocking amphibole and plagioclase crystals with subordinate iron ore and rarely biotite. Crystals are smaller than metagabbro which also has sutured contact between grains.

B29
B35 + sphene
B38-39 + sphene (+ sericite + epidote after plagioclase)
B46 contains areas of diorite ?vein: plagioclase + quartz rich
B48
B98
B99 relict orthopyroxene and metagabbroic layer
BA9 + epidote + quartz/albite
BA28
BM571 (?magmatic) hornblendite: - plagioclase

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APPENDIX E

BM572 (cumulate) + sphene + epidote + mica + chlorite

C. Sheared rocks

B9-12 mylonitised gneissose rocks: quartzo-feldspathic biotite gneiss with chlorite veining and possible pseudotachylite
B16 foliated metabasite: quartz + plagioclase + epidote + amphibole + iron ore
B28 mylonitic schist: amphibole + plagioclase + quartz + muscovite + sphene + iron ore + graphite (+ chlorite + calcite)
B31-33 mylonitic schist: amphibole + plagioclase + iron ore + sphene (+ chlorite + prehnite + sericite)
B36 as B31-B33
B42-43 shear zone, contact between coarse grained amphibolite (+ prehnite after pyroxene) and fine grained amphibolite (+ sphene)
B50-53 sheared amphibolites: amphibole + plagioclase + iron ore + biotite + epidote + sphene (+ prehnite in veins)
B55 sheared metagabbro: relict clinopyroxene + amphibole + plagioclase
B101 sheared amphibolite: amphibole + plagioclase + relict orthopyroxene and clinopyroxene in a serpentinite
B107 cataclastic texture in metagabbro: amphibole + plagioclase + quartz
BA17 mylonitic metagabbro: amphibole + plagioclase + quartz + epidote
BA38 foliated metagabbro with epidote veins
BR269 as B50-53
BM563 as B50-53 - muscovite
BM564 foliated amphibolite + epidote; alternating bands of quartzo-feldspathic and hornblende schist
BM565 as BM563
BM567-568 quartzo-feldspathic amphibole gneiss: amphibole + quartz + plagioclase + biotite + epidote + iron ore (+ retrograde chlorite)
BM570 as BM567-568

[3] Saleh Metasedimentary Rocks

[A] Phyllites

SM3 foliated amphibole quartz phyllite (metatuff): composed primarily of angular and poorly sorted green amphibole + quartz + plagioclase + iron ore + epidote (in veins). Quartz appears to be both of detrital and authigenic origin, locally filling the veins.
APPENDIX E

BA37 metatuff: epidote + quartz + brown biotite + iron ore (retrograde chlorite)
BR39-40 calc-quartzo-feldspathic phyllite: calcite + quartz + plagioclase + smectite + iron ore
BT49 as BR39-40
BT52 as BR39-40
BR28 foliated calcareous phyllite: calcite + quartz + amphibole + iron ore porphyroblasts in a micritic matrix

[B] Schist:
BA30 augen plagioclase + quartz + brown biotite + iron ore (+ calcite + retrograde chlorite)

[C] Calc-quartzo-feldspathic gneiss
BA27 alternating layers of epidote-rich quartzo-feldspathic gneiss, with distinct quartz horizons, and amphibole-rich quartzo-feldspathic gneiss. Strong foliation defined by alignment of brown biotite and muscovite in matrix of plagioclase and quartz
BA33 as BA27 - plagioclase

C. Non-garnet schists.
as BA15

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