COMMONWEALTH OF AUSTRALIA
DEPARTMENT OF NATIONAL DEVELOPMENT
BUREAU OF MINERAL RESOURCES, GEOLOGY AND GEOPHYSICS

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Peridotite-Gabbro-Basalt Complex in Eastern Papua:

An Overthrust Plate of Oceanic Mantle and Crust

BY

H. L. DAVIES

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CORRIGENDUM

Page 22—line 39 should read: 148°35′E; Smith & Green, 1961) and hyaloclastite (8°40′S, 147°35′E). Higher

Page 35—caption should read: Figs 8 and 9. Variation of MgO with Al_2O_3 and 'FeO'.

Page 42—line 6 should read: Basal Group Pillow lavas of altered basalt with many dykes

Page 44—line 3 should read: the plutonic rocks has not been established (e.g. Bortolotti et al., 1969).

Page 44—line 25 should read: much more complex structure 'probably formed by nearly solid penetrative flow'.

COMMONWEALTH OF AUSTRALIA

DEPARTMENT OF NATIONAL DEVELOPMENT

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SUMMARY

The Papuan Ultramafic Belt is a peridotite-gabbro-basalt complex which crops out over a length of 400 km (NW-SE) and a width of 40 km, on the northeastern side of the Owen Stanley Range in eastern Papua (7°-10°S, 147°-149°E). From top to bottom the complex consists of

Basalt zone	Basalt and spilite, massive and as pillow lavas, some dacite	4-6 km
Gabbro zone	High-level gabbro (ophitic), to 1 km Granular gabbro, some cumulates	4 km
Ultramafic zone	Cumulates, up to 0.5 km Noncumulates: harzburgite, etc., with metamorphic textures	4-8 km

The basalts have tholeiitic affinities but contain more SiO_2 (52%) and less K_2O (.07%) than typical oceanic tholeiites. The gabbros are rich in CaO (14.5%) and MgO (12.2%) and poor in alkalis (Na_2O .07%, K_2O .04%); Mg/Fe ratio is high. The cumulus ultramafics have variable Mg/Fe in contrast with the noncumulus ultramafics which consist mainly of olivine Fo_{93} and enstatite En_{93}.

The complex may be part of an overthrust sheet of oceanic crust and mantle. The noncumulus ultramafics might represent pre-existing (convecting?) mantle, while the basalt, gabbro, and cumulus ultramafics are the products of intrusive-extrusive activity at a tensional zone in the ocean floor (e.g., a midocean ridge). If the basalt and gabbro are related, the parent magma must have been richer in Mg and poorer in alkalis than supposed typical tholeiite melts, and the basalts must have crystallized from residual liquids produced by crystal fractionation in the crust.

Basalt, gabbro, and cumulus ultramafics crystallized probably in the Cretaceous and were intruded by tonalite in the Eocene. The complex was thrust over Cretaceous sediments in Eocene or Oligocene time, presumably as a result of interaction between Australian and Pacific plates.

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The work described in this Bulletin was presented in partial fulfilment of the requirements for the degree of Doctor of Philosophy at Stanford University.

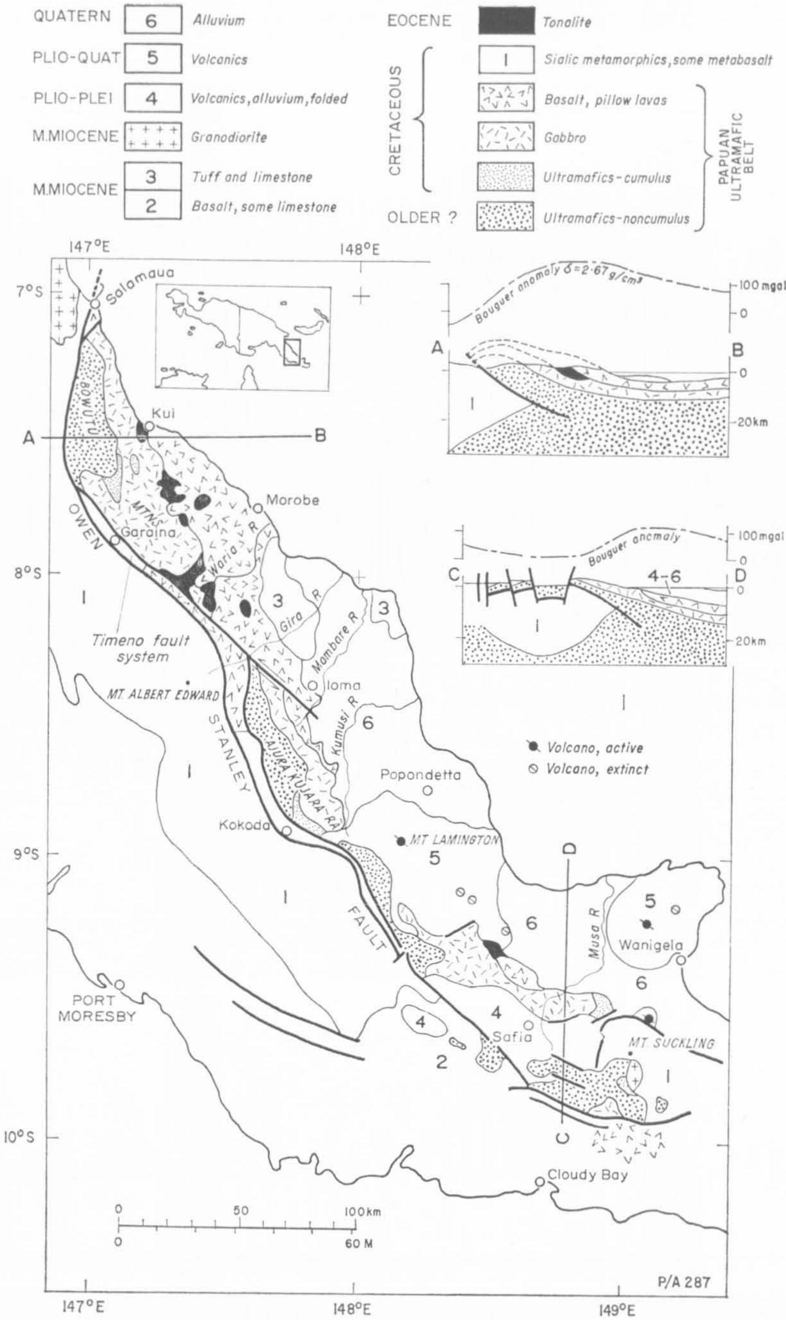


Fig. 1. Papuan Ultramafic Belt: location, map, and sections.

INTRODUCTION

This Bulletin presents preliminary results of an investigation of the Papuan Ultramafic Belt, a peridotite-gabbro-basalt complex which is thought to be part of a plate of Cretaceous oceanic mantle and crust. A summary of results up to 1967, which includes a brief review of previous work, has already been published (Davies, 1968). R. N. England is currently making a detailed mineralogical study of the peridotites and gabbros.

Location and access

The Ultramafic Belt is exposed over a length of 400 km and a width of 40 km along the northern flank of the Owen Stanley Range in eastern Papua (Fig. 1). For ease of discussion the Belt may be divided along its length into northern (7°S-8°10′S), central (8°10′S-9°15′S), and southeastern (9°15′S-10°S) sectors, the whole area is bounded by longitudes 146°55′E and 149°10′E. The northern sector is made up of the Bowutu Mountains; the central sector of the Otavia, Ajura Kujara, and Guava Ranges; and the southeastern sector of the Sibium, Didana, Keman, and Awariobo Ranges. These ranges have average elevations of 1200-1500 m and peaks to 2500 m above sea level; peaks in the Owen Stanley Range reach 4000 m. Major rivers are the Waria (northern sector), Gira-Aikora, Mambare, and Kumusi (central sector), and Musa (southeastern sector). Main settlements are Garaina, Kui, and Morobe (northern sector), Ioma, Kokoda, and Popondetta (central sector), and Afore and Wanigela in the southeast. Access is by boat in the north and by air elsewhere; motor vehicle roads connect Kokoda and Popondetta to the coast, and short logging roads radiate from Kui.

This account of the geology is based on foot traverses across the Belt at 6 to 12 km intervals.

Rock specimens and thin sections are stored at Bureau of Mineral Resources, Canberra. All specimen numbers have the project prefix 6552.

TECTONIC SETTING

J. E. Thompson (pers. comm., 1957; Thompson & Fisher, 1965) was the first to propose that the Papuan Ultramafic Belt might be a plate of oceanic crust and upper mantle thrust southwestward over the sialic core of eastern Papua. Although it is not yet conclusively proved, various lines of evidence support Thompson's concept. The evidence is examined in the following paragraphs.

*Evidence that the Belt is a plate dipping oceanward**

Several writers have described ultramafic sheets in central and western Papua-New Guinea that dip southwards towards the Australian continent (Dow et al., 1968; Visser & Hermes, 1962). The interpretation that the Papuan Ultramafic Belt dips away from the Australian continent and towards the ocean is based on (1) the distribution of rock types within the Ultramafic Belt and (2) the pattern of Bouguer anomalies.

The distribution of peridotite, gabbro, and basalt within the complex suggests

* The term oceanward is used because the direction of dip is variable from northeast to southeast. For instance, in the northern part of the Bowutu Mountains dip is easterly at perhaps 30°; in the southern part of the Bowutu Mountains dip is southeasterly, and the greater widths of gabbro and basalt that are exposed indicate that the angle of dip is much lower, perhaps around 10°.

a crude layering with peridotite inland (toward the Australian continent), gabbro centre, and basalt oceanward. The classic interpretation of such layered sequences is that peridotite is at the base and basalt at the top; as a consequence, the sequence must dip oceanward (sections, Fig. 1). Direct evidence, however, cannot be obtained from contacts and layered structures within the complex. Attitude of the rare layering within the peridotite and gabbro is not consistently in one direction, and has presumably been disturbed by minor faulting. Sequences of interlayered massive lava and pillow lava in the basalt of the Bowutu Mountains are sub-horizontal or have northerly and easterly dips between 20° and 40°.

The gravity evidence for easterly dip was first recorded by St John (1967, pp. 110-4) and has been amplified by Milsom (pers. comm., 1968). It has been discussed elsewhere (Davies, 1968) and will be recapitulated here only briefly. Bouguer lows coincide with the outcrop of the sialic metamorphics; Bouguer highs are associated with ultramafic outcrop areas but are consistently offset oceanwards from the outcrop by 10-20 km. This is taken to indicate oceanward dip of the ultramafic bodies (sections, Fig. 1). Angles of dip computed by St John are between 20° and 30°.

The dipping plate is thought to be continuous with *in situ* oceanic mantle and crust, a suggestion that could be tested by seismic refraction studies and gravity measurements at sea. On the basis of onshore gravity St John (*loc. cit.*) has postulated thick sialic crust seaward of the Belt. In view of the regional geology this seems unlikely; perhaps instead of sialic crust there are Upper Tertiary sediments of greater thickness and lower density than St John has allowed.

Evidence of thrusting

Overthrusting is evidenced by (1) direct surface mapping of the fault zone which bounds the Ultramafic Belt, and by (2) the metamorphic facies of rocks immediately below, or west of, this fault zone.

The fault system

The fault system which bounds the Ultramafic Belt on the southwest side can be traced for a straightline distance of 300 km and is marked by a sinuous valley with steep southwestern face. For most of its length it actually consists of two or more faults (Fig. 1; Pl. 1): a major fault which bounds the sialic metamorphics (the Owen Stanley Fault) and a series of subsidiary faults which bound the ultramafics (the Timeno Fault system). In the northern Bowutu Mountains the two faults merge, but elsewhere they are separated by a fault slice, 5-10 km wide, of basalt, minor marl, and gabbro which is probably Ultramafic Belt material caught up in the thrusting.

The Owen Stanley and Timeno Faults are marked by shear zones between 100 and 500 m wide. The Timeno shear zone is sheared serpentinite at 7°55′S-8°00′S; in some other areas it includes zones of metamorphics of amphibolite and granulite facies (e.g. at 8°18′S, 9°08′S, 9°14′S, 9°47′S, and 9°50′S; Fig. 2 and Pl. 2, fig. 1). In other localities the Timeno fault system is marked by greenschist-facies rocks which grade westwards to prehnite-pumpellyite facies and virtually unmetamorphosed basalts of the fault slice. A zone of greenschist-facies mylonite up to 1.5 km wide is developed where the Timeno fault system coincides with a major strike-slip fault (Gira Fault); displacement is left-lateral and is of the order of 90 km.

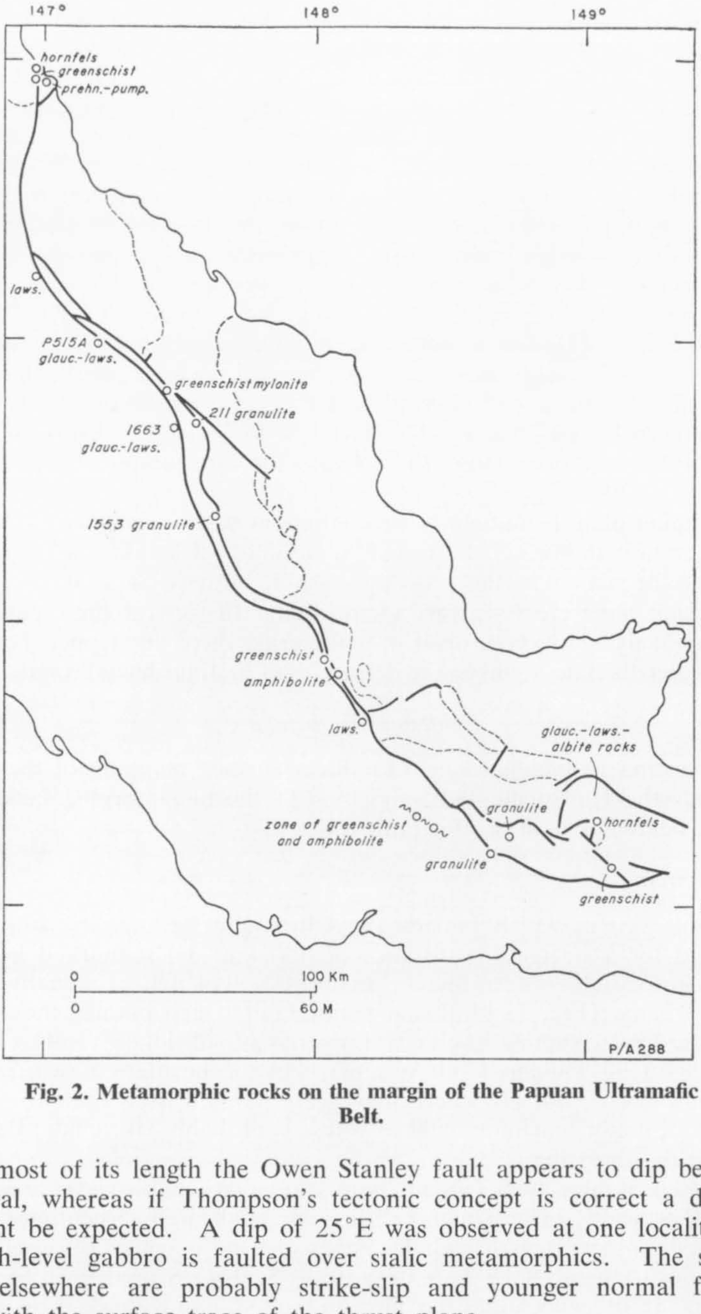


Fig. 2. Metamorphic rocks on the margin of the Papuan Ultramafic Belt.

For most of its length the Owen Stanley fault appears to dip between 70°E and vertical, whereas if Thompson's tectonic concept is correct a dip of about 30°E might be expected. A dip of 25°E was observed at one locality (9°20′S) where high-level gabbro is faulted over sialic metamorphics. The steeper dips observed elsewhere are probably strike-slip and younger normal faults which coincide with the surface trace of the thrust plane.

The metamorphics of amphibolite and granulite facies along the Timeno fault system are interpreted as the products of high temperatures and pressures generated during thrusting (cf. Scott & Drever, 1953; Hsu, 1969). The granulites consist of brown hornblende, plagioclase, pyroxene, and in one case olivine (Pl. 2, fig. 1). The composition of one analysed granulite (specimen 97) is intermediate between typical gabbro and basalt of the Ultramafic Belt. A sample of greenschist metamorphics from the Timeno fault system (No. 94) has a composition near that of average Ultramafic Belt basalt, except for higher K_2O.

Metamorphism below the thrust plane

The generally sialic rocks west and southwest of the Owen Stanley Fault, i.e. immediately below the supposed thrust plane, have metamorphic mineral assemblages intermediate between blueschist and greenschist facies. These are characterized by lawsonite and glaucophane but rarely the two together. A typical lawsonite-bearing schist consists of quartz-calcite-lawsonite-graphite-muscovite-albite with, less commonly, actinolite-glaucophane-chlorite-stilpnomelane-epidote-pumpellyite-sphene-garnet. Typical glaucophane-bearing rocks are metabasalts and metagabbros, some of which are not schistose. Glaucophane is commonly associated with pumpellyite-epidote-chlorite-albite-quartz-sphene and relict clinopyroxene, and, less commonly, with stilpnomelane-muscovite-lawsonite-garnet. Textures in most of these rocks indicate a second stage of deformation after that in which the metamorphic minerals developed.

These mineral assemblages indicate high-pressure low-temperature conditions. Similar assemblages have been found below major thrusts in California and Oregon (Blake et al., 1967) and in New Caledonia (Coleman, 1967; Black, 1969). The conditions may be due to overpressure related to thrusting (Blake et al., 1967; Coleman, 1967) or to increased hydrostatic pressure in the downthrust plate (Ernst, 1965, 1969). In the case of the Papuan rocks overpressure related to thrusting seems more likely.

The Papuan rocks have not been mapped in any detail, and may well prove to be much more varied in metamorphic grade than is suggested in this brief summary. High-grade blueschists are known from one area (8°01′S); these are coarsely crystalline (graphite)-quartz-muscovite-lawsonite-glaucophane gneisses. Greenschist metamorphics occur immediately below the thrust in another area: this is southeast of Mount Suckling (at 9°42′S, 149°E), where rapid erosion by the Ibinambo River has exposed the contact in a steep landslide scar (Pl. 2, fig. 2). Mineral assemblages in two rock specimens are sphene-quartz-albite-epidote-actinolite (1808) and epidote-garnet-chlorite-muscovite-quartz (1809). The higher temperatures indicated by these assemblages may be related to emplacement of the nearby granodiorite pluton.

Evidence that the Papuan Ultramafic Belt is oceanic crust and mantle

The Papuan Ultramafic Belt may be:

(a) part of a faulted plate of oceanic crust and mantle (Thompson & Fisher, 1965; Davies, 1968),

(b) a differentiated lopolith of the Bushveld type (Dow & Davies, 1964), or

(c) a thick differentiated seafloor extrusion formed in the manner sometimes proposed for Mediterranean ophiolites (Dubertret, 1955; Brunn, 1956, 1960). The arguments that favour hypothesis (a) are:

1. It provides the simplest explanation in view of the regional setting of sialic crust to the southwest and oceanic crust to the northeast, and accords with the evidence of thrusting and oceanward dip of the thrust plate.

2. The 8-10 km thickness of the gabbro and basalt zones matches crustal thicknesses of 7 and 8 km in the Solomon Sea (Rose et al., 1968; Khan & Woollard, 1968).

3. The rock types of the Belt can be duplicated completely by rocks dredged from the present-day ocean floors.

4. The Belt corresponds quite well with other proposed examples of elevated oceanic crust and mantle, such as the Mediterranean ophiolites and the Troodos complex of Cyprus.

5. The ultramafic zone of the Belt is similar to alpine-type peridotites (Thayer, 1967; Loney, et al., in press) in regard to its mineralogy, rock types and relations between them, and degree of deformation. Peridotites of this type are commonly held to be faulted fragments of the upper mantle (Hiessleitner, 1952; Hess, 1955; Roever, 1957).

Points 3-5 are elaborated in pages 40-43 and following.

Plate tectonics and emplacement of the Ultramafic Belt

The thrust emplacement and subsequent strike-slip faulting of the Papuan Ultramafic Belt can be explained in terms of plate tectonics (Le Pichon, 1968; Isacks et al., 1968). Papua-New Guinea is on the northern edge of the Australian plate and straddles the contact with the Pacific plate (Denham, 1969). Relative to a fixed Antarctic plate, the Australian plate is currently moving north and the Pacific plate moving west (Fig. 3). These relative motions have probably continued since upper Eocene time. The northward movement of the Australian plate caused north-south compression in the Papua-New Guinea region. The main result was the development of the New Britain trench as a sink for the Australian plate; a subsidiary effect may well have been the thrusting of a plate of oceanic mantle and crust (the Papuan Ultramafic Belt), southwards over the Cretaceous geosynclinal sediments of what is now eastern Papua. Westward movement of the Pacific plate undoubtedly has caused east-west compression and sinistral torque in the Papua-New Guinea region. The main result has probably been left-lateral strike-slip faulting along the Solomon Islands-New Ireland outer margin; a subsidiary effect may well have been left-lateral strike-slip faulting of the Papuan Ultramafic Belt thrust sheet. The present rate and azimuth of differential movement between the Australian and Pacific plates in the New Guinea region is 11 cm/year compression on azimuth 069° (computed for 3°S, 142°E; Le Pichon, 1968).

Time of overthrusting

The marine magnetic record (Le Pichon & Heirtzler, 1968) and paleomagnetic measurements on land (McElhinny & Wellman, 1969) indicate that Australia began to move north away from Antarctica in the upper Eocene, about 43 m.y. ago (Anomaly 17). The New Britain Trench probably developed in the Eocene, for the oldest known rocks on New Britain are Eocene (R. J. Ryburn, pers. comm., 1969); the island was probably constructed by island-arc volcanic activity after the trench developed. Overthrusting of the Papuan Ultramafic Belt plate also took place in early Cainozoic time. Overthrust and underthrust plates include Creta-

Fig. 3. Probable movements of lithosphere plates in the New Guinea region.

ceous rocks, and thrusting preceded deposition of lower Miocene sediments. The most likely time of overthrusting would seem to be upper Eocene or Oligocene.

The east-west compression caused by movement of the Pacific plate was also probably operative in early Cainozoic time, and perhaps played some part in the overthrusting. The strike-slip faulting caused by Pacific plate movement appears to postdate the overthrust, and probably continued into the late Cainozoic.

Post middle Miocene uplift

Vigorous vertical movements since mid-Miocene time have been superimposed on the thrust and strike-slip faults. Judging from gravity data, these are not simple isostatic adjustments (J. S. Milsom, pers. comm., 1968) and are probably best explained as responses to horizontal compressive forces such as those discussed above.

The Mid-Miocene to Recent vertical movements have caused continuing elevation of the Owen Stanley Range sialic core and of the inland edge of the Ultramafic Belt. The most spectacular evidence of uplift is seen in the southeast, where fault blocks of unconsolidated gravels in the Musa Valley have been recently elevated 300 m or more, and fault scarps and tilted river gravels are common. Despite the field evidence of continued vertical movements the area is seismically quiet.

PAPUAN ULTRAMAFIC BELT

The Papuan Ultramafic Belt may be subdivided into an Ultramafic zone, a Gabbro zone, and a Basalt zone (Table 1); the Basalt zone overlies the Gabbro zone, which in turn overlies the Ultramafic zone, and tonalite (quartz diorite) intrudes at the gabbro-basalt interface. The Gabbro and Basalt zones are presumed to be oceanic crust, and the Ultramafic zone oceanic mantle. Assuming a dip of 30° for cross-sections at 7°30′S and 8°35′S, the Ultramafic zone has a minimum thickness of about 4 km, the Gabbro zone about 4 km, and the Basalt zone 4-6 km.

TABLE 1. ROCK UNITS OF THE PAPUAN ULTRAMAFIC BELT

Cretaceous	*Basalt zone*	Basalt, spilite, lavas, pillow lavas, dacite in one area.	4-6 km	Interpreted as oceanic crust
Eocene	*Tonalite stocks*			
Cretaceous?	*Gabbro zone*	High-level Gabbro: ophitic, may grade upwards into Basalt zone; up to 1 km thick. Granular Gabbro: allotriomorphic granular, includes cumulates; grainsize 1-2 mm.	4 km	
Cretaceous? and older	*Ultramafic zone*	Cumulus ultramafics: probably grade upwards into cumulus gabbros; up to .5 km thick. Noncumulus ultramafics: harzburgite, dunite, enstatite pyroxenite; deformed metamorphic textures; grains 4-20 mm.	4-8 km	Interpreted as oceanic mantle

The gabbro, basalt, and uppermost ultramafics may be cogenetic, in the sense that the gabbros and uppermost ultramafics may represent magma reservoirs for the basalts. This is suggested by cumulus* textures within the gabbros and uppermost ultramafics, and by the high proportions of magnesium and calcium in the gabbros. The underlying and greater part of the Ultramafic zone is probably not a part of this magma cycle. Most probably it represents pre-existing mantle, above which the gabbros and basalts were intruded and extruded.

Ultramafic zone

Ultramafic rocks are exposed along the inland side of the Ultramafic Belt for most of its length. In the northern Bowutu Mountains they form a pod about 20 km wide and 50 km long, and in the Otavia/Ajura Kujara Ranges they form an elongate lens about 10 km wide and 50 km long. If an average dip of 30° is assumed, the approximate thickness of ultramafics in the Bowutu Mountains is 8 km and in the Ajura Kujara Range is 4 km. If dips are flatter near the basal fault, because of drag along the fault, these estimates of thickness may be too high by perhaps a factor of 2. In the extreme southeast the ultramafics are probably a relatively thin subhorizontal sheet: they cover a large area but there is no associated positive gravity anomaly. Exposures in the Keman and Awariobo Ranges indicate that the thrust sheet is at least 2 km thick.

The Ultramafic zone is made up of olivine, pyroxene, and chromite; what little plagioclase it includes is confined to gabbro pegmatite dykes and tongues of intrusive gabbro and tonalite near the contact with the Gabbro zone. The complete absence of plagioclase in the ultramafic rocks themselves is noteworthy as it sets the Ultramafic zone apart from the ultramafic zones of the great layered intrusions such as Stillwater, Bushveld, and Great Dyke (Jackson, 1967) and Muskox (Irvine & Smith, 1967).

Near the contact with the overlying Gabbro zone some of the ultramafic rocks have cumulus textures, and are included in the Cumulus member of the Ultramafic zone. The greater part of the Ultramafic zone has no recognizable cumulus textures and will be called the Noncumulus member. The Cumulus member may well be the basal part of the great pile of gabbroic cumulates which makes up much of the Gabbro zone. The Noncumulus member probably originated separately as primitive mantle material which formed the country-rock floor for the gabbroic intrusions. If this is the case, one would expect to find a contact where cumulates rest on the Noncumulus member, or perhaps even a basaltic chilled margin, but no exposure of such a contact has yet been found.

Noncumulus Member

The Noncumulus member makes up more than 90 percent of the Ultramafic zone. It consists of harzburgite, dunite, and orthopyroxenite; constituent minerals are: olivine Fo_{93}, enstatite En_{93}, and chromite; clinopyroxene occurs only as exsolution lamellae in enstatite. Harzburgite is the most common rock type; it consists of olivine (60-80%), enstatite, and accessory chromite. Most harzburgite is homogeneous rock, with enstatite distributed randomly throughout

* I.e., textures which indicate accumulation of early-formed crystals on the floor of a magma chamber.

(Pl. 3, fig. 1); but some is layered with alternating enstatite-rich and enstatite-poor horizons (Pl. 4). Dunite is closely associated with harzburgite, either in layers as described above, in irregular interfingering masses, or, less commonly, as dunite dykes and veins; it consists of olivine and accessory chromite. Enstatite pyroxenite forms veins, dykes, and irregular bodies up to 5 m thick, always intrusive into the other rocks.

In hand specimen dunite and harzburgite are pale green and vitreous, darker green with a waxy appearance if partly serpentinized, and brown on weathered surface. The distribution of enstatite is best judged on weathered surfaces, where the pyroxene stands out by reason of its cleavage and greater resistance to erosion. Grainsizes in harzburgite and dunite average about 4 mm, with individual grains up to 1 or 2 cm and a few even 10 cm. Grains are interlocking anhedra with varying degrees of deformation (Pl. 3, fig. 2; Pl. 5).

Enstatite pyroxenite is pale greyishyellow in hand specimen. It has two common forms of occurrence: (1) as narrow planar veins, 1-3 cm across, which criss-cross most ultramafic outcrops, and (2) as coarse-grained irregular dykes up to 5 m thick. The coarse-grained pyroxenites have been found in only a few areas, but within these areas they may account for 20-30 percent of the total exposed rock. Individual grains are up to 10 cm across and are moderately to strongly deformed. Most of the coarse-grained pyroxenites consist solely of enstatite and accessory chromite, but one sample (26, Saia River; 7°23′S, 147°06′E) includes some olivine, both intergranular and as irregular grains which penetrate the larger pyroxene grains. The accessory chromite in these rocks is also coarse-grained, commonly averaging 8-10 mm instead of the 0.1 or 0.2 mm characteristic of the dunite and harzburgite. The coarse-grained enstatite pyroxenites appear to be pegmatitic intrusions. Origin by silica metasomatism of olivine-rich rock seems unlikely because metasomatism would not explain the coarser accessory chromite, nor would one expect the sharp contacts which characterize the dykes.

The best exposures of coarse-grained pyroxenite are in the central Ajura Kujara Range (the Okawu and Opi River headwaters at around 8°44′S) in rocks which are probably less than 1 km below the Gabbro zone contact. Cliffs on the Opi River show irregularly shaped bodies of coarse-grained pyroxenite intruding relatively coarse-grained (2-3 cm) dunite.

Deformation

Deformation in rocks of the Noncumulus member is ubiquitous but variable in intensity. Most rocks show moderate deformation with kink-bands in both olivine and enstatite (Pl. 3, fig. 2). Within any one olivine grain, rotation of extinction angles by kinking is usually about 10° or 15°. In any one thin section there are usually a number of grains in near optical continuity; these have probably re-oriented under stress so that crystallographic planes (slip planes) are parallel to the strain direction, or have recrystallized in the strain field. Commonly kinking has progressed to the point of rupture; thus figures on average grainsize quoted are probably minima. The likelihood that large grains have broken down into smaller presents problems to fabric analysis, as does the coarse grain of the rocks.

More strongly deformed rocks are found near the Owen Stanley and Timeno Faults, where textures range from moderately cataclastic to mylonitic. Cataclasites persist across a zone about 1 km wide immediately east of the basal fault in the Bitoi River, Aikora River, and Log Creek sections (7°13′S, 8°17′S, and 8°33′S respectively; Pl. 5, fig. 1). In the northern Guava Range (9°00′-9°05′S) the strong parallel northwesterly lineaments which partly control drainage are probably strain lineaments, rather than faults, marked by subcataclasis comparable to textures near the basal fault.

In the extreme southeast (Keman and Awariobo Ranges) deformation is generally more severe than elsewhere in the Ultramafic Belt, and mylonites and cataclasites are relatively common. One mylonite specimen (No. 624 from 9°54′S, 149°04′E) consists of olivine and enstatite fragments up to 1 or 2 mm long embedded in a finely granulated (0.01 mm) olivine and enstatite matrix. The overall higher degree of deformation in the southeast accords with the interpretation that the thrust sheet here is relatively thin.

The least deformed Noncumulus ultramafics are in the thick lenses of ultramafics in the Bowutu and Ajura Kujara mountain ranges, and particularly towards the eastern edge or top of the lenses. These lenses or pods appear to have acted as cohesive structural units, with deformation concentrated at their margins and, in the case of the Ajura Kujara Range, along scattered northwesterly shears. The least deformed ultramafic sample is a coarse-grained harzburgite (Pl. 5, fig. 2) collected 3 km west of the gabbro contact on the Paiawa River; grains 10 and 20 mm across are barely kinked and maximum rotation of extinction in any one grain is 2°. However, the many triple-grain junctions in the thin section suggest that the texture may be due to annealing and grain growth from a previously deformed rock.

Evidence for separate origin of the Noncumulus ultramafics

1. Complete lack of cumulus textures sets the Noncumulus ultramafics apart from the overlying cumulus ultramafics and gabbros.

2. Although cumulus textures are not easy to recognize in olivine-rich olivine-enstatite rocks, the complete lack of plagioclase or any minerals other than olivine, enstatite, and chromite, makes cumulus origin especially unlikely. Plagioclase-free cumulates are known from the great layered intrusives, but not with thicknesses anywhere near 4 km.

3. At the present stage of mineralogical investigation there seems to be a break in chemical composition between Noncumulus and Cumulus olivine and enstatite (R. N. England, pers. comm., 1969), but this is not yet definitely established.

4. The grainsize of Noncumulus rocks is considerably coarser than that of the overlying cumulates and gabbros, 4-10 mm as against 1-2 mm.

5. The ratio of ultramafic rocks to gabbro and basalt is too high, at 1:2, for the ultramafics to have developed by crystal settling from gabbroic or basaltic liquids. Ratios of ultramafics to gabbro in the great layered intrusions are between 1:4 and 1:6 (Jackson, 1967, pp. 25-6).

6. There is not nearly enough felsic enrichment in the gabbros to balance the mafic enrichment of the Ultramafic zone.

7. The presence of dunite dykes and coarse-grained pyroxenite dykes indicates that the Noncumulus ultramafics have been held at temperatures near 1500° C at some stage in their evolution (cf Loney et al., in press). Temperatures of formation of layered intrusives of the Stillwater-type are near 1200° C (Hess, 1960; Jackson, 1961).

8. Noncumulus ultramafics are generally more severely deformed than the Cumulus ultramafics. Although this might be explained by the proximity of the Noncumulus member to the basal fault, it is taken to imply that a period of deformation separated the two members.

Noncumulus member as primitive upper mantle

If the Noncumulus ultramafics have no genetic link with the overlying rocks, it is most likely that they represent part of a pre-existing, i.e. pre-Cretaceous, upper mantle. They are similar in all regards to typical alpine peridotites (Hess, 1955; Thayer, 1960, 1967; Loney et al., in press), and alpine peridotites have long been supposed to be dismembered parts of the upper mantle (Hiessleitner, 1952; Hess, 1955; Roever, 1957). Green & Ringwood (1967) suggest that ultramafics of this type (low Ca, Al, etc.) represent refractory mantle from which basaltic magma has been generated by partial melting. Partial melting of these rocks to generate the Cretaceous lavas is unlikely, for at that time they were only 10-15 km below the ocean floor, and it is improbable that temperature at this depth would be high enough for partial fusion. If they are a refractory residue they probably yielded their basalt component at greater depth, and well before Cretaceous time.

The Cumulus member

The Cumulus member forms the uppermost part of the Ultramafic zone and is approximately 100-500 m thick. It consists of ultramafic cumulates with no plagioclase; the component minerals are olivine, orthopyroxene, clinopyroxene, and chromite. Each occurs as a cumulus phase in some rock, separately or with the other minerals. The member is distinguished from the Noncumulus member by the presence of cumulus textures, by its generally lower degree of deformation, by its generally, but not invariably, finer grainsize, and by the presence of clinopyroxene as a major mineral phase. It is distinguished from overlying Gabbro zone cumulates by its complete lack of plagioclase.

The extent and thickness of the Cumulus member are not known. Rocks of this type have been found near the contact between the Ultramafic and Gabbro zones at scattered localities throughout the Ultramafic Belt. On present evidence it seems unlikely that the Cumulus member is continuous along this contact but rather that it is well developed in some areas and absent in others. One of the problems in defining the extent of the cumulates is the general difficulty in recognizing olivine-rich cumulates, especially in the field. Even in thin section good crystal shapes of settled olivine grains are readily disguised by postcumulus olivine overgrowth or by reaction with postcumulus enstatite. Olivine cumulates can be recognized by fabric analysis (Jackson, 1961) and shape studies may ultimately be the best way to define the Cumulus member. Alternatively, the composition of the olivine may prove to be diagnostic.

The best section of ultramafic cumulates (Pl. 6) is in the south arm of the Wele River in the Bowutu Mountains (7°42′S, 147°07′E). Rock types include

olivine-hypersthene*-augite cumulate	layered cumulate
olivine-chromite cumulate	(No. 1548)
olivine-hypersthene-augite cumulate	(1554)
olivine-augite cumulate?	(1556)
hypersthene-augite cumulate?	(1558)

A layered cumulate from the west side of the Guava Range (93; 9°13′S, 148°06′E) consists of layers about 1 cm thick of the following settled phases: olivine, chromite, olivine-hypersthene, olivine. The cumulus assemblages olivine-augite and hypersthene-augite suggest affinities to the Muskox paragenesis, rather than that of the Stillwater-Bushveld (Jackson, in press).

Serpentinization

Most of the Ultramafic zone rocks are serpentinized to some extent; an estimated average figure for the entire zone is about 20 percent. Serpentinization is related to shearing, but some shear zones are free of serpentine. For example, the Timeno fault system at the Aikora River (8°18′S) includes unserpentinized olivine mylonite. Serpentinization of ultramafic rocks appears to increase near contacts with gabbroic rocks, and definitely increases near tonalite intrusions. Some serpentinization may also be related to groundwater movement, as has been suggested in western North America (Barnes et al., 1967; Barnes & O'Neil, 1969). A relationship with groundwater movement or weathering is suggested for the Papuan ultramafics because the best exposures of unserpentinized ultramafic rock are found in rapidly eroded gorges, e.g. Paiawa (7°35′S), Aikora (8°17′S), upper Opi (8°43′S), Nia (9°05′S), Duuba (9°48′S, 148°55′E), and Adau (9°50′S, 148°44′E). An exception is the Mambare Gorge (8°37′S), where rocks are moderately serpentinized. Thus, although overall serpentinization is estimated at 20 percent at the present surface, it is probably considerably less at depth, where one might expect shears to be less common, and water to be generally scarce. There is certainly no evidence of the serpentinite layer proposed by Hess (1955) in a model of oceanic mantle and crust.

The serpentine minerals appear to be mostly or entirely chrysotile and lizardite except in the Mount Suckling area (about 9°40′S, 149°E), where antigorite is common (optical identification). Green (1961) noted antigorite in the ultramafic breccias of the Musa valley, 15 km northwest of Mount Suckling.

* Hypersthene = orthopyroxene En_{70-90}.

Gabbro zone*

The Gabbro zone crops out between peridotite to the west and basalt to the east for most of the length of the Ultramafic Belt. The zone has an average thickness of about 4 km. Mappable units are:

Rock unit	*Thickness*	*Description*	*Grainsize*
High-level Gabbro member	1 km appr.	Ophitic or subophitic texture, and/or zoned plagioclase laths	1-2 mm.
Granular Gabbro member	3 km appr.	Granular allotriomorphic or hypidiomorphic texture. Rock types include (1) cumulus gabbro, (2) homogeneous intrusive gabbro, (3) streaky gabbro, and, less commonly, (4) very fine-grained gabbro, and (5) gabbro pegmatite	1-2 mm.
Cumulus Gabbro submember	up to 3 km	Where cumulus gabbros could be easily recognized in the field, and are present in sufficient quantity, they have been mapped as a separate submember of the Granular Gabbro member	1-2 mm, some coarser

Granular Gabbro member

The Granular Gabbro member includes a wide variety of gabbroic rocks which are characterized by allotriomorphic or hypidiomorphic granular texture, with or without some degree of planar or linear fabric. Minerals are augite, hypersthene, bytownite, and in some rocks olivine. Primary hornblende is rare. Opaque iron oxides and sulphides are lacking in most of the granular gabbros, but make up as much as 10 percent by volume of some of the very fine-grained gabbros. Chromite and rutile are rare accessories in some cumulus gabbros.

Cumulus Gabbro submember

Cumulus gabbros are much more common than was previously reported (Davies, 1968): as many as 30 percent of the thin sections of rocks of the Gabbro zone show diagnostic cumulus textures (Pl. 9; Pl. 10; Pl. 11, fig. 2). Other gabbros with oriented fabric may or may not be cumulates; some owe their fabric to flow during crystallization, but others may be 3-phase cumulates (e.g. augite-hypersthene-bytownite cumulates) in which the diagnostic crystal shapes and sedimentary texture have been disguised by overgrowth on all three settled phases. Where the cumulus gabbros can be readily recognized in the field and crop out over a sufficiently large area, they have been differentiated from the other granular gabbros as a Cumulus Gabbro submember.

The cumulates which are most easily recognized in the field are those which are layered, especially where the layering includes sedimentary structures such as graded bedding (Pl. 7, fig. 2; Pl. 8, fig. 1) and scour-and-fill structures (Pl. 8, fig. 2). The graded beds consist of mafic minerals at the bottom and plagioclase at the top of each bed; this has been called rhythmic layering (Wager & Deer,

* Gabbro is used in a broad sense to include norite, picrite, troctolite, and other gabbroic rocks.

PLATE 1

Waria valley looking northwest. Owen Stanley Range (sialic metamorphics) is on the left, Bowutu Mountains (gabbro and some ultramafics) on the right. The valley floor is bounded by the Owen Stanley and Timeno Faults (on left and right sides respectively) and includes some outcrop of Cretaceous(?) basalt and marl. Oblique air-photograph taken from about 8°S, 147°15′E, and 7200 m altitude.

PLATE 2

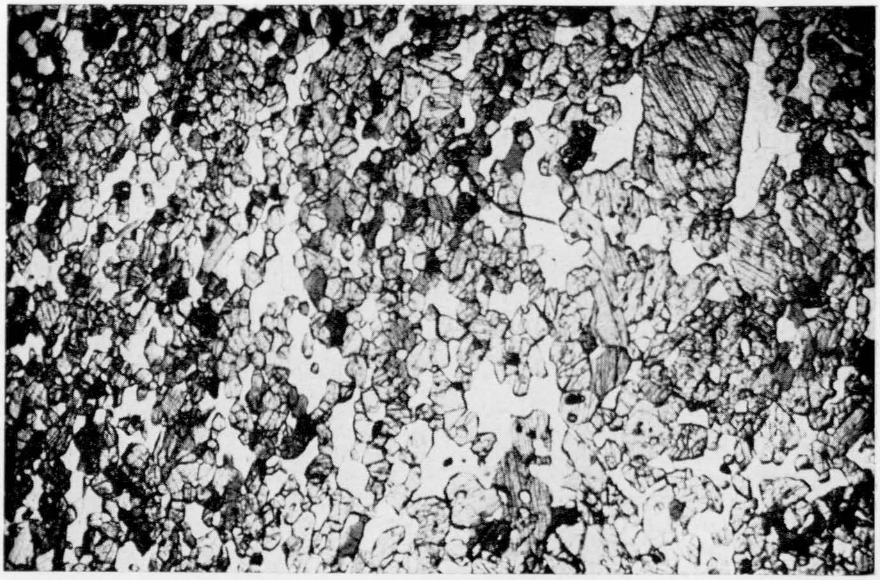


Figure 1. Olivine-hornblende-pyroxene-labradorite granulitic gneiss from south of Mt Avuru (9°47′S, 148°42′E). Photomicrograph with plane light shows equigranular (0.1 mm) red-brown hornblende (dark), clino- and orthopyroxene, several larger pyroxene oikocrysts, and plagioclase.

Figure 2. Thrust contact between ultramafics above (lighter colour) and basic greenschist facies metamorphics (darker), in the head of the Ibinambo River at 9°42′S, 149°E. The contact and the schistosity in the metamorphics dip 60°SE, away from the viewer.

PLATE 3

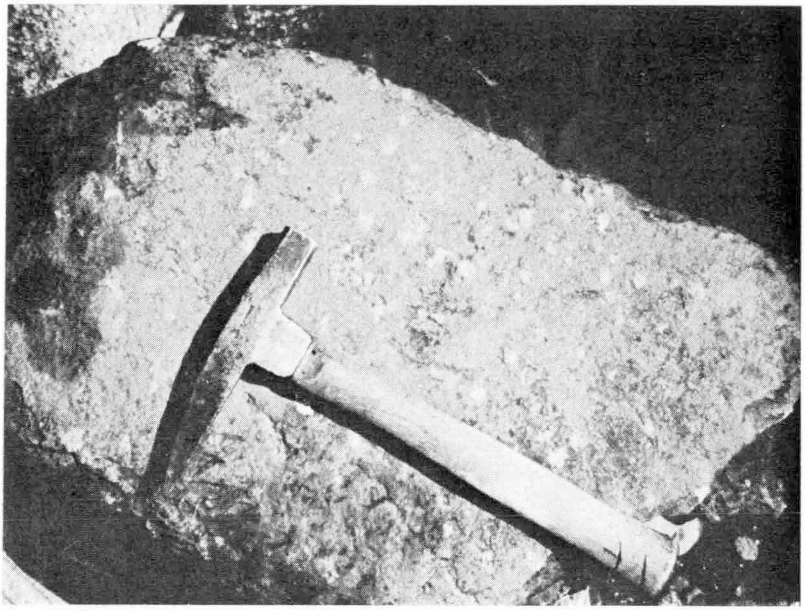


Figure 1. Typical harzburgite of Noncumulus member ultramafics; enstatite grains (lighter) are randomly distributed in olivine-rich rock.

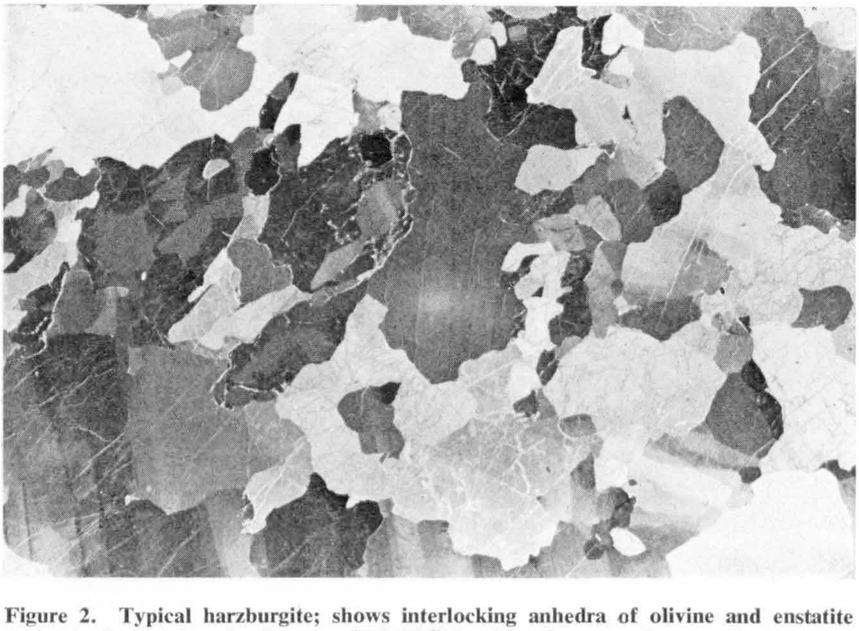
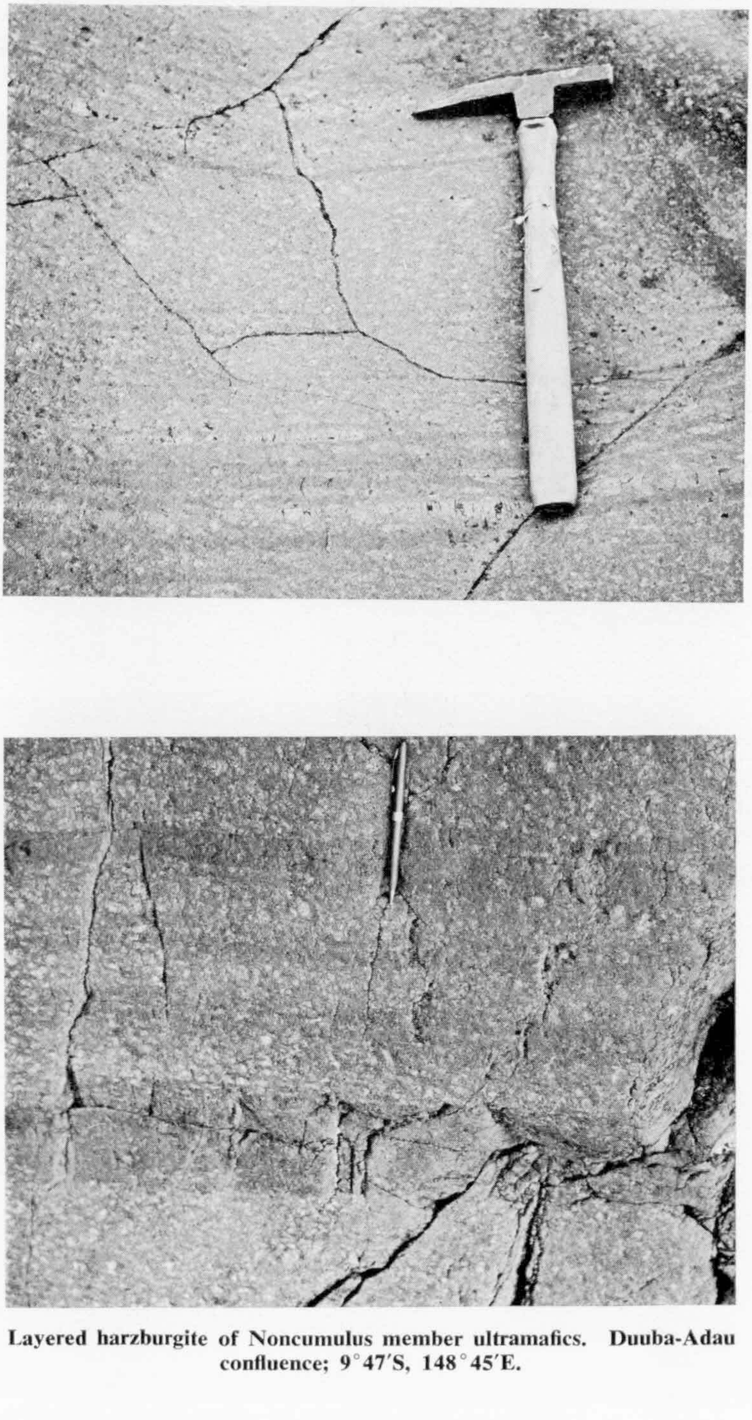


Figure 2. Typical harzburgite; shows interlocking anhedra of olivine and enstatite (minor); both minerals show strain lamellae. The larger grains are about 4 mm across. Crossed nicols.

PLATE 4



Layered harzburgite of Noncumulus member ultramafics. Duuba-Adau confluence; 9° 47′S, 148° 45′E.

Figure 1. Protoclastic harzburgite from outcrop 850 m east of the Timeno fault system of the Otavia Range. Specimen 1053; Log Creek; 8°33′S, 147°36′E. Field of view is 2 x 3 cm. Crossed nicols.

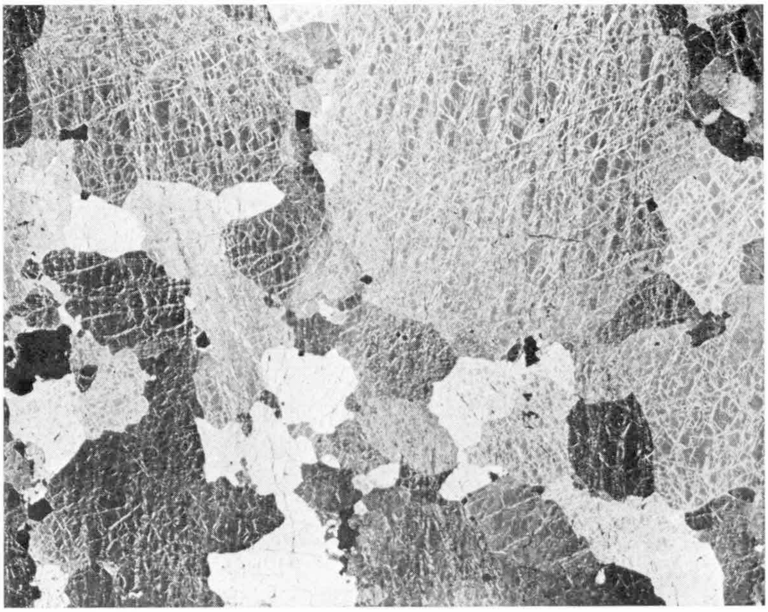
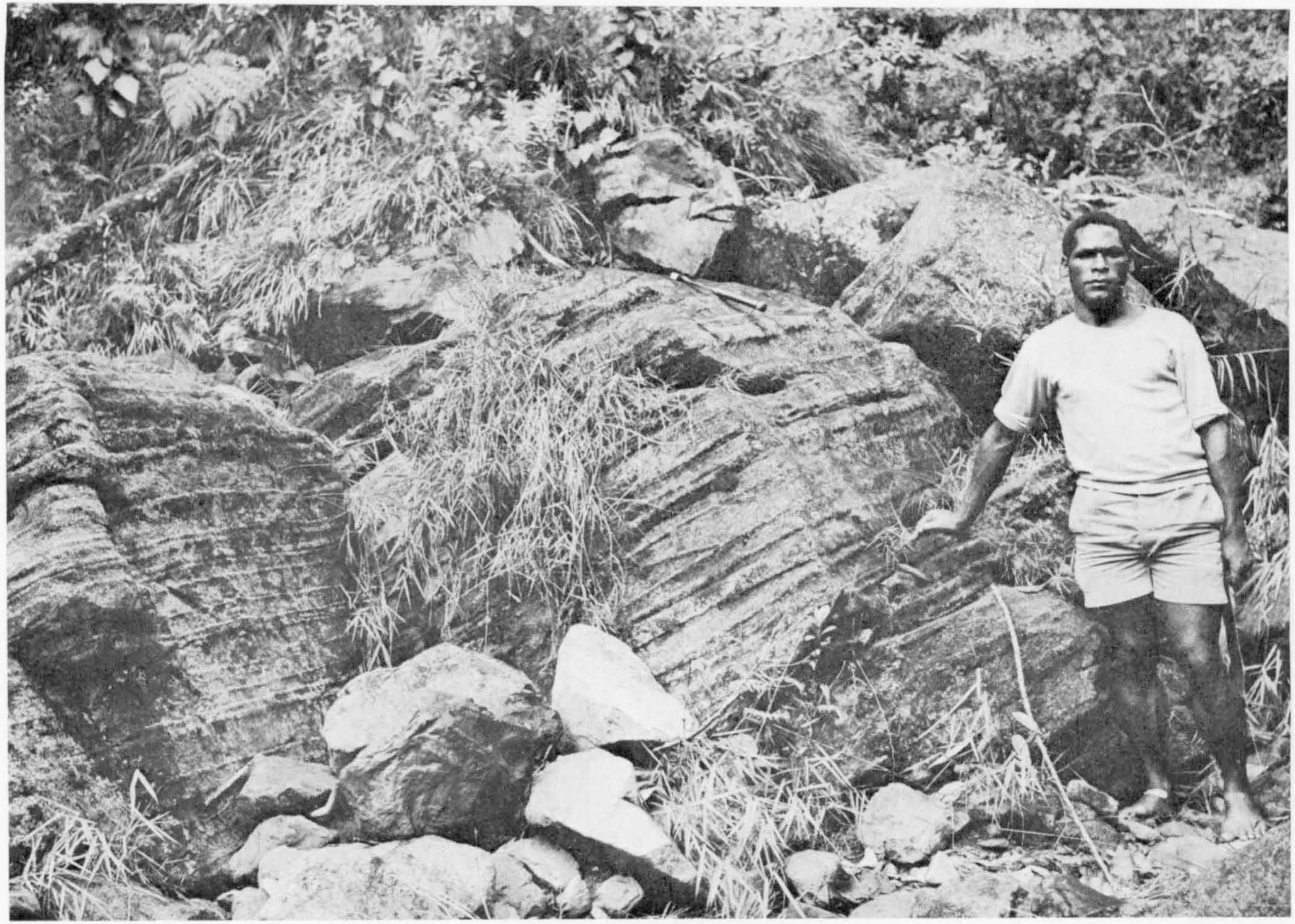


Figure 2. Harzburgite, No. 507 from Paiawa River (7°35′S, 147°06′E). Unlike most other peridotite samples, the olivine grains are not kinked. However, the many triple-grain junctions suggest that the present texture is due to annealing and grain growth from a previously deformed rock. The large grain is about 15 mm across. Crossed nicols.

PLATE 6



Layered ultramafic cumulate, Wele River South, 7°42′S, 147°07′E; cumulus phases are olivine, chromite, hypersthene, and augite; postcumulus phases are hypersthene and augite. The sequence contains no plagioclase. Pyroxene-rich layers are more resistant to erosion.

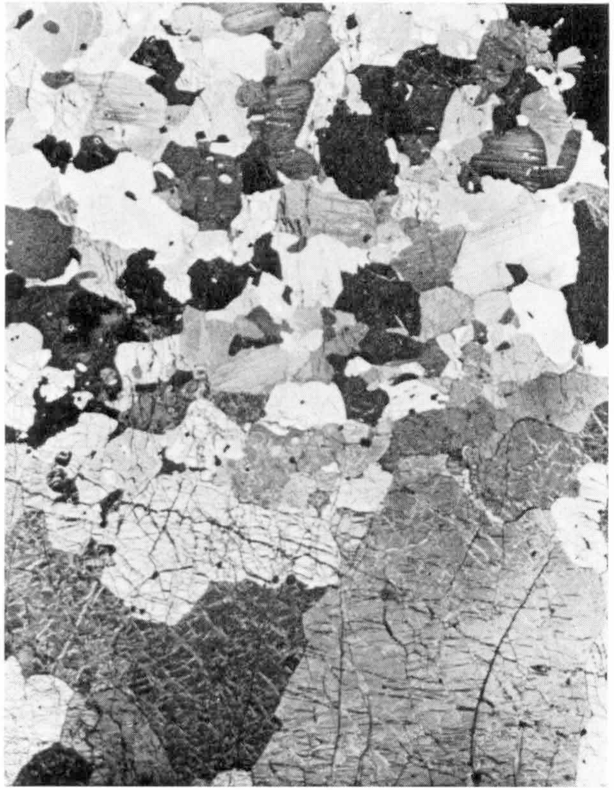


Figure 1. Contact between olivine-chromite cumulate (below) and hypersthene-olivine-augite cumulate; same outcrop as Plate 6. Overgrowth by reaction with intercumulus liquid has disguised the cumulus grains; cumulus origin is suggested by the orientation of the *c*-axis of the orthopyroxene parallel to the phase contact. Average grainsize 1 mm at top and 5-7 mm at bottom. Specimen 1548. Crossed nicols.

Figure 2. Mineral graded bedding (rhythmic layering) in a hypersthene-bytownite-augite cumulate; specimen 764, Duuba River (9°48′S, 148°46′E).

PLATE 8

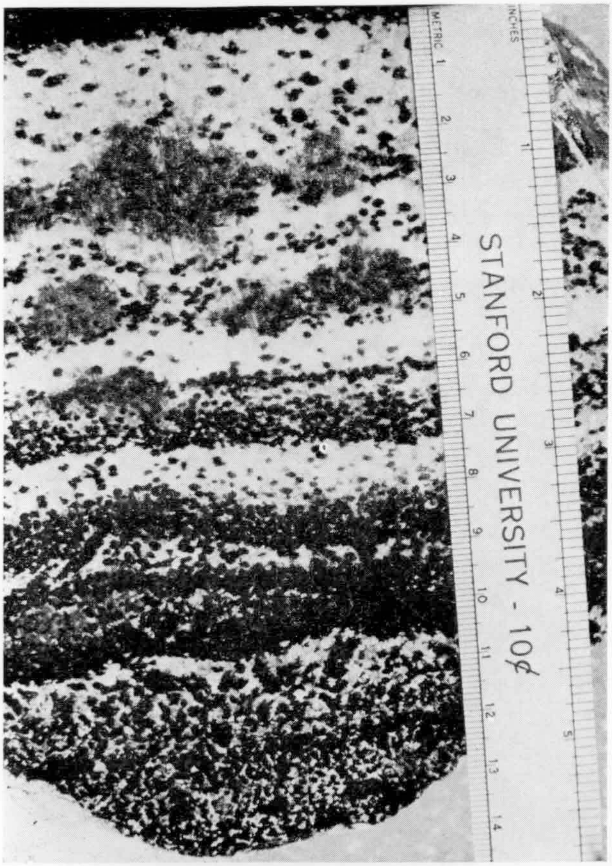


Figure 1. Mineral graded bedding in a plagioclase-olivine cumulate with postcumulus clinopyroxene oikocrysts. Eight or nine mineral graded beds, 1-3 cm thick, make up a larger mineral graded unit 10 or 14 cm thick (see text); specimen 1979, Sivai Creek; 9°34′S, 148°52′E.

Figure 2. Scour-and-fill structures in mineral graded hypersthene-bytownite-augite cumulate; Wele River South; 7°42′S, 147°13′E.

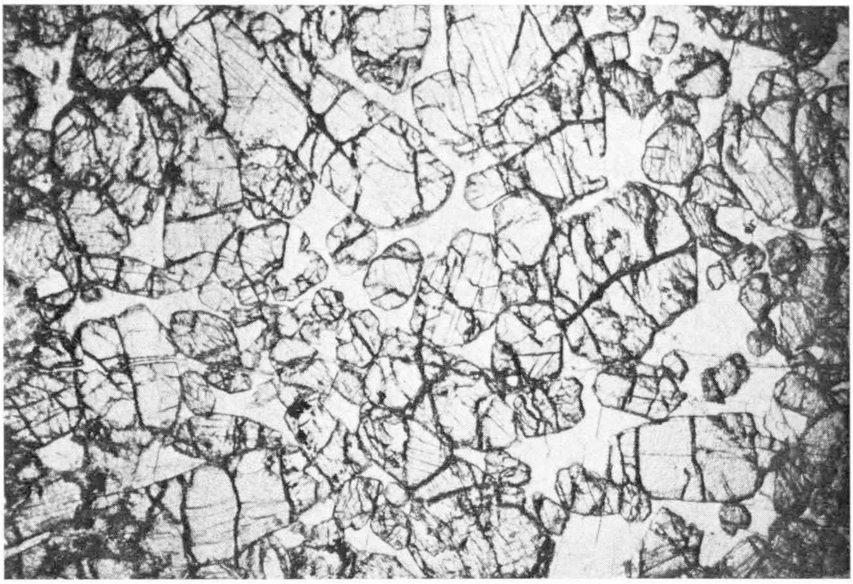


Figure 1. Hypersthene cumulate—not obviously layered in hand specimen. Hypersthene grains are mostly 0.2-0.3 mm across and up to 1.5 mm long; postcumulus bytownite oikocrysts are 5-10 mm across. Specimen 560; Waki Creek; 9°53′S, 148°55′E. Plane light.

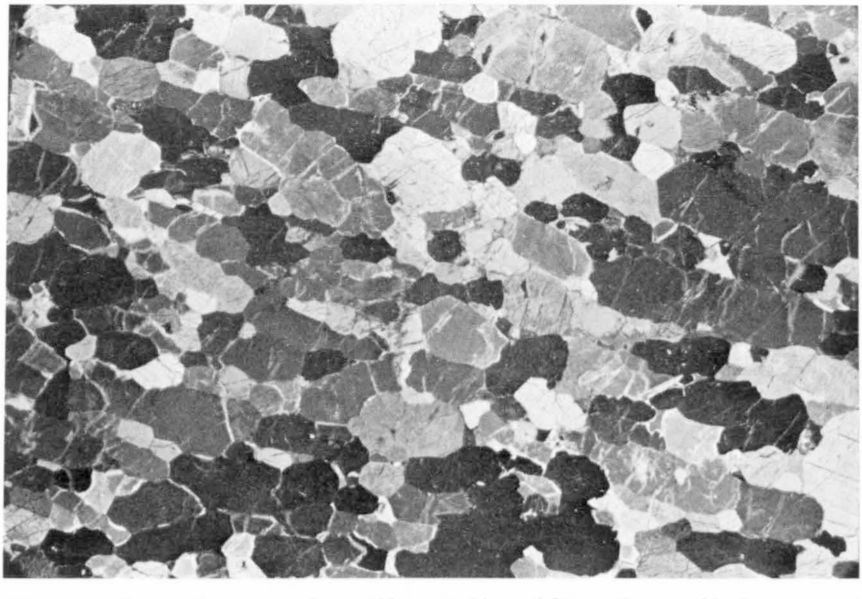


Figure 2. Hypersthene cumulate. Close packing of hypersthene subhedra suggests that intercumulus fluid was squeezed off; there is a little postcumulus clinopyroxene and plagioclase. Specimen 528; Sou River; 7°35′S, 147°13′E. Grain lengths are 1-2 mm. Crossed nicols.

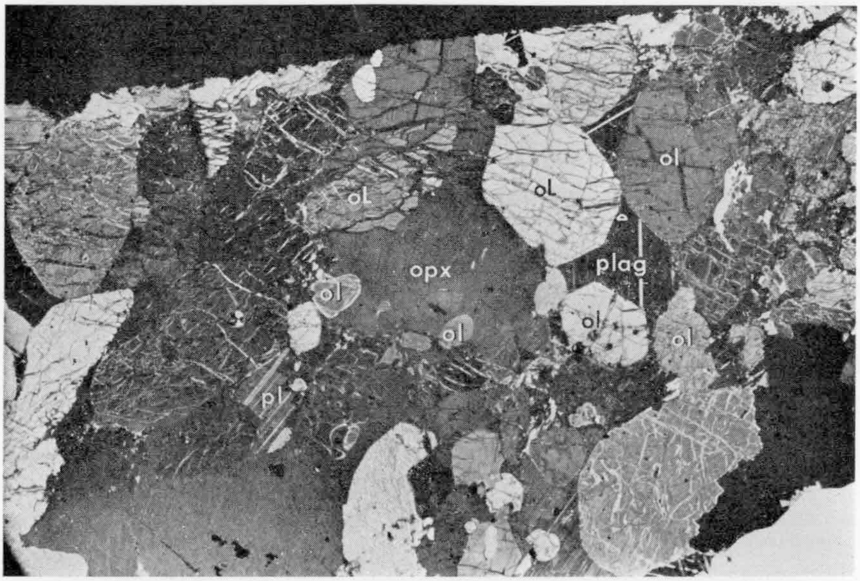


Figure 1. Reaction between cumulus olivine and postcumulus hypersthene. Olivine grains enclosed in hypersthene have irregular embayed shapes; those enclosed in plagioclase are subhedral, e.g. upper and centre right. Olivine grains are 1 mm across. Specimens 1150; 14715 Creek; 7°38′S, 147°15′E. Crossed nicols.

Figure 2. Reaction between cumulus olivine and postcumulus clinopyroxene (e.g. upper right), and lack of reaction between cumulus plagioclase and postcumulus clinopyroxene. Cumulus grains are 1 mm across; clinopyroxene oikocrysts are up to 10 mm. Specimen 429; coast near Lake Salus; 7°10′S, 147°10′E. Crossed nicols.

PLATE 11

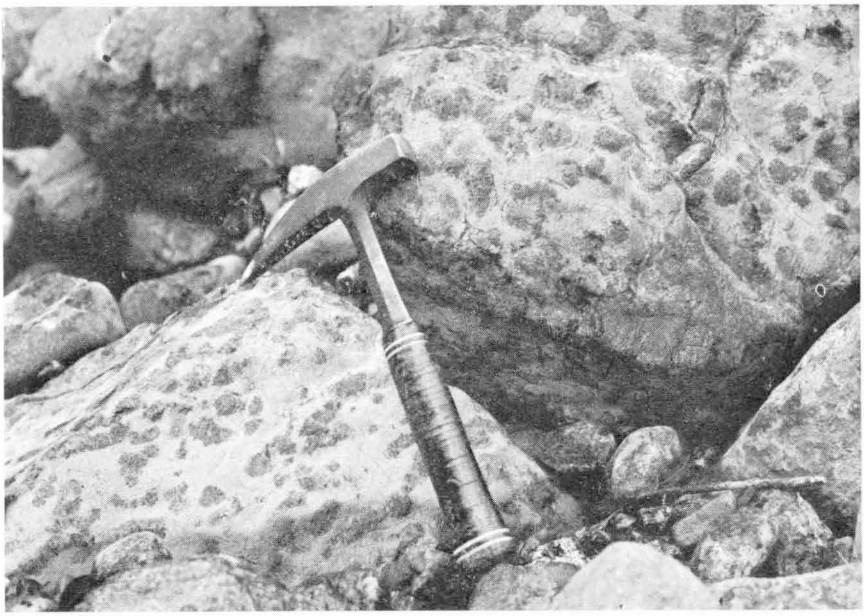


Figure 1. Olivine cumulate with postcumulus plagioclase and poikilitic orthopyroxene. This is texturally similar to the poikilitic harzburgite (olivine cumulate) of the Ultramafic zone of the Stillwater Complex. Wele River West; 7°42′S, 147°08′E.

Figure 2. Bytownite cumulate with postcumulus green hornblende. The largest plagioclase grains are 3 mm across. Specimen 1137; Sisa tributary; 8°37′S, 147°49′E. Crossed nicols.

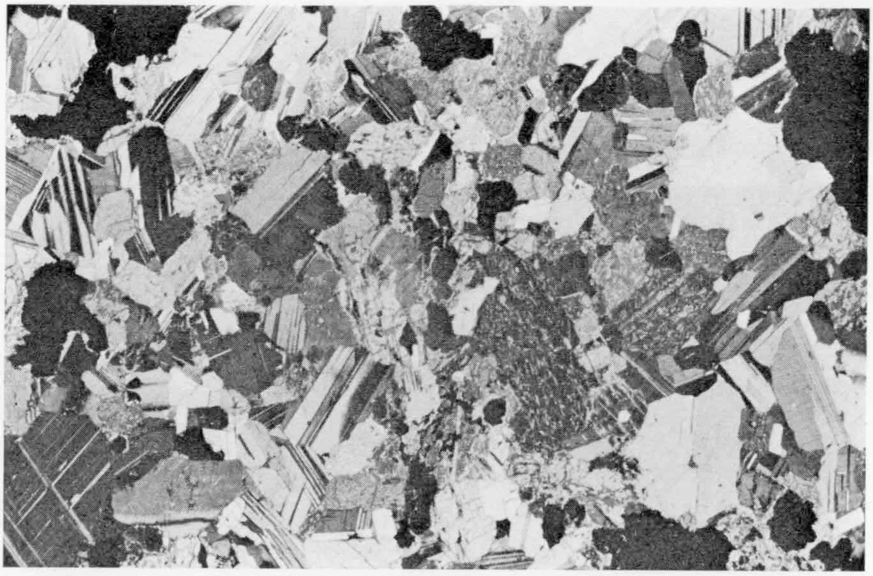


Figure 1. Intrusive homogeneous gabbro. Minerals are bytownite, hypersthene, and clinopyroxene; texture is hypidiomorphic granular, and grainsize 1-2 mm. Specimen 532; Sou River; 7°37′S, 147°13′E. Crossed nicols.

Figure 2. Streaky gabbro. Melagabbro, probably a cumulate, with irregular subparallel leucogabbro veins. Waki Creek; 9°53′S, 148°55′E.

Figure 1. Streaky gabbro in this case may be a deformed layered cumulate. Wele River West; 7°42′S, 147°08′E.

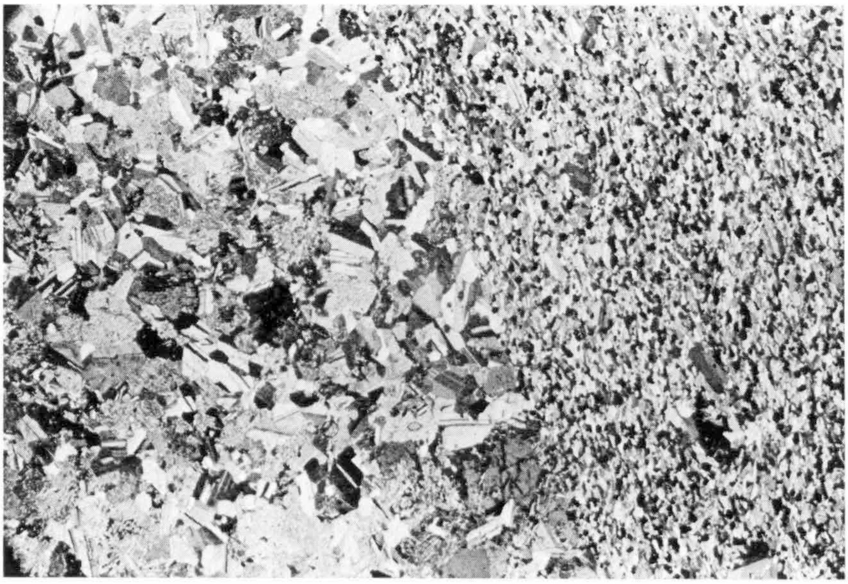
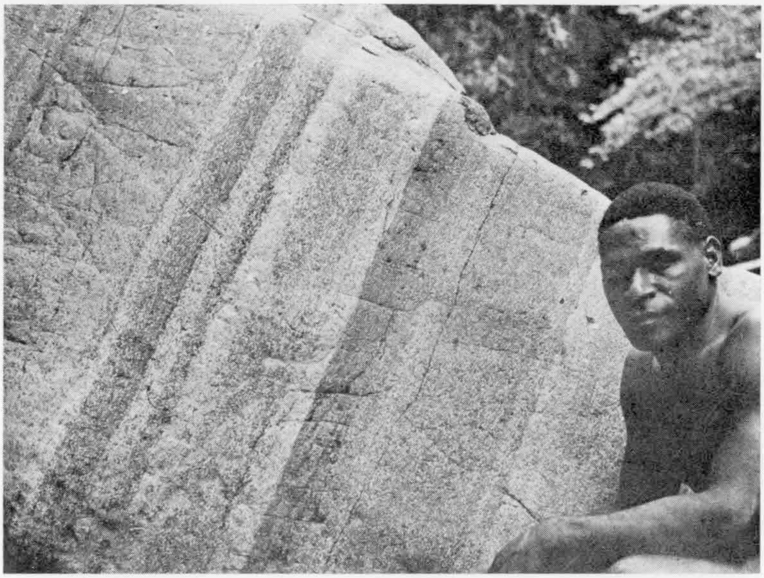


Figure 2. Very fine-grained gabbro dyke (grainsize 0.2 mm) in coarser (0.5-1.0 mm) gabbro. The dyke rock has oriented fabric due to flow; opaque iron oxides make up 10-15 percent of the mode. Specimen 38; Paiawa River; 7°34′S. 147°13′E. Crossed nicols.

PLATE 14



A deformed layered olivine gabbro which may or may not be a cumulate. Specimen 844; Musia Creek; 9°25′S, 148°25′E.

PLATE 15

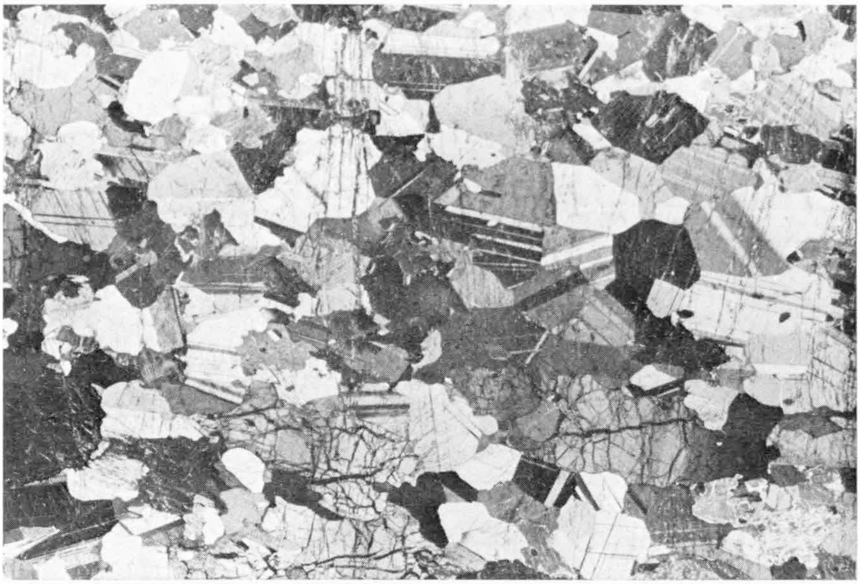


Figure 1. Deformed layered olivine gabbro (sp. 844, Pl. 14). The section was cut normal to the layering. Plagioclase subhedra and larger kink-banded olivine oriented with their long axes in the layering plane. Another thin section cut parallel to the layering shows that the fabric is planar rather than linear. If this was a cumulate the original olivine grain shapes have been modified by plastic flow under stress. Plagioclase grainsize is around 1 mm. Crossed nicols.

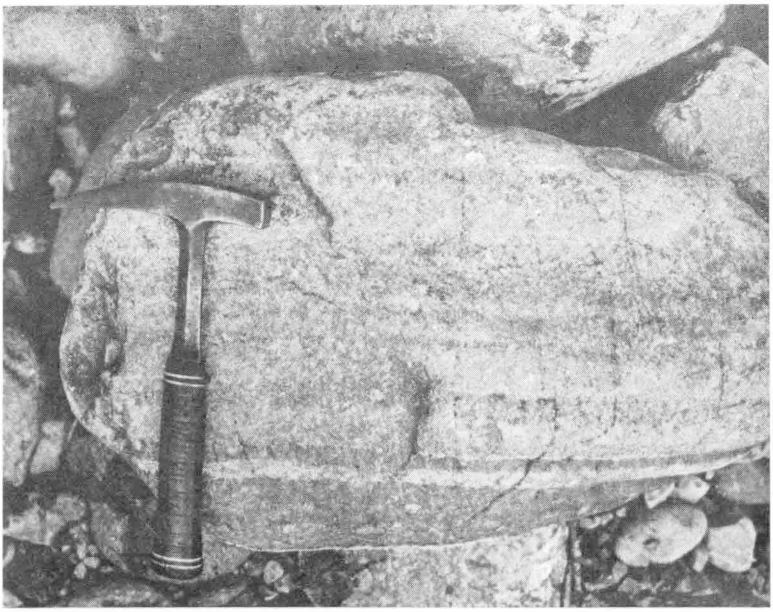


Figure 2. Layered gabbro, probably a cumulate. Olivine-rich layers are serpentinized and subsequently sheared. Athoro River; 8°50′S, 148°20′E.

PLATE 16

The eastern Bowutu Mountains, looking northwest. The ridge in the left foreground has the distinctive concave slopes of the Basalt zone.

1939) or mineral graded bedding (Jackson, 1967, table 1). Each bed supposedly represents a shower of crystals. Rates of settling are proportional to the density of the grain and to the square of its size, and thus the mafic grains settle out first, provided that they are as large as, or larger than, the plagioclase grains. Layers in gabbro 1979 (Pl. 8, fig. 1) are successively poorer in mafic minerals from the bottom to the top of an eight-layer sequence. This suggests that there are two mechanisms controlling the development of layers, one with a long period and the other a short one. Perhaps the long-period event is the precipitation of a shower of crystals, and the short-period events are pulses of movement in the magma. The scour-and-fill structures seen elsewhere (Pl. 8, fig. 2) indicate that currents disturbed the floor of the magma chamber from time to time.

Broader scale layering may juxtapose two completely different rock types; for instance, a 40-cm homogeneous layer of olivine cumulate may be overlain with sharp contact by a clinopyroxene-plagioclase cumulate perhaps 30 cm thick. This is a phase contact (Jackson, 1967, table 1) and indicates a crystallization hiatus followed by a sudden and unexplained change in the nature of the crystallization products. Another type of layering involves sudden changes in grainsize at what Jackson would term form contacts. Thus in a sample from the Awariobo Range (sample 713, Doriri Creek, 9°51′S, 148°45′E) hypersthene cumulate is overlain by very fine-grained hypersthene-bytownite cumulate which becomes progressively coarser and richer in plagioclase upwards.

The best exposures of the layered cumulates are in the southern Sibium and the Didana Ranges (9°20′-9°35′S) and were initially mapped and described by Smith & Green (1961) and Green (1961). The layered cumulates are exposed in areas up to 4 km across and 15 km long. By plotting average width and average dip of the layers we can estimate a thickness of 1-3 km, but these figures should be accepted with caution, for dips are rather variable and there is some evidence of faulting. Individual layers have not been traced beyond the limits of outcrop in any one stream bed. With careful mapping it might be possible to establish whether the same stratigraphic succession extends for distances of up to 15 km. My impression is that it does not.

Localities of three common types of rhythmic layering are summarized in the following table:

Cumulus phases	*Localities*	
Olivine, clinopyroxene, plagioclase	Sibium and Didana Ranges	
	Inaf Creek	9°25′S, 148°20′E
	Musia Creek	9°25′S, 148°25′E
	Nr Mamama Creek	9°30′S, 148°43′E
	Ajura Kujara Range	
	Athoro River	8°50′S, 147°48′E
	Bowutu Mountains	
	Morobe River	7°48′S, 147°17′E
Orthopyroxene, clinopyroxene, plagioclase	Keman-Awariobo Ranges	
	Duuba River	9°47′S, 148°46′E
	Bowutu Mountains	
	Wele River	7°42′S, 147°13′E
Orthopyroxene, plagioclase	Bowutu Mountains	
	Saru River	7°56′S, 147°14′E

Cumulates which show no obvious layering in outcrop or hand specimen (Pls 9 and 10) are at least as common as the layered cumulates. A good illustration is in the Sibium Range (9°20′S, 148°25′E) north of the layered cumulates. About 20 samples of apparently homogeneous gabbro were collected in this area and most of these were subsequently found to have cumulus textures.

Postcumulus mineral phases in different rocks are clinopyroxene, orthopyroxene, and plagioclase, singly or in various combinations; postcumulus hornblende has been found in a few samples from near the top of the Granular Gabbro member, accompanied, in one instance, by biotite, quartz, and andesine (925; 8°25′S, 147°38′E). The general scarcity of postcumulus biotite, quartz, sodic plagioclase, and hornblende indicates either that a very efficient mechanism removed intercumulus fluid, or that the magma was initially poor in soda, potash, and water, and the path of crystallization did not tend to enrich silica in the rest magma. Such a crystallization path would require a balance between the crystallization of silica-rich minerals such as hypersthene, and that of silica-poor minerals such as bytownite and olivine.

Reaction between settled olivine and postcumulus enstatite (Pl. 10) indicates that pressure during crystallization was less than 5-7 kilobars, if, as seems likely from the mineralogy, the immediate environment was anhydrous (Boyd et al., 1964; Kushiro, 1969).

Present knowledge of the distribution of the different types of cumulates is sufficient for only a few general statements. Cumulates which approach ultramafic bulk compositions tend to be at the bottom of the Granular Gabbro member, where they grade (with disappearance of plagioclase) into the cumulates of the Ultramafic zone. A typical rock in this gradational category is olivine cumulate that contains postcumulus hypersthene oikocrysts and plagioclase (Pl. 11, fig. 1). These rocks are texturally similar to the poikilitic harzburgite (olivine cumulate) of the Ultramafic zone of the Stillwater Complex (Jackson, 1961).

The layered cumulates are perhaps best developed in the lower half of the Gabbro zone, but cumulates of all kinds, including olivine and hypersthene cumulates, are found throughout the Granular Gabbro member, right up to the contact with the High-Level Gabbro. No overall upward enrichment in plagioclase has been recognized, but it may be significant that the few pure plagioclase cumulates found are from near the top of the unit (Pl. 11, fig. 2).

Homogeneous intrusive gabbro

Homogeneous gabbro with allotriomorphic or hypidiomorphic granular texture is the other main rock type of the Granular Gabbro member. It is consistently fine-grained (1-2 mm) and usually shows some degree of planar or linear fabric in thin section. In some exposures it is clearly intrusive into cumulus rocks. In others it contains xenoliths of a very fine-grained gabbro that will be discussed later. The homogeneous gabbros are similar to cumulus gabbros in chemistry and mineralogy: both contain bytownite, brown hypersthene, green augite, and, in some cases, olivine. Olivine where present is commonly partly or completely enclosed by hypersthene. Bytownite typically shows Carlsbad and Albite twinning. Opaques are generally absent.

Streaky gabbro

This term is used for a variety of gabbros characterized by discontinuous lenticular layers, commonly cut by gabbro veins. Some streaky gabbros are gabbroic cumulates with subparallel wavy leucogabbro veins (Pl. 12, fig. 2), others may have the same origin but have been modified by later flowage (Pl. 13, fig. 1), and still others may simply be deformed layered cumulates (Pls 14 and 15).

Streaky gabbro produced by concordant intrusions of later gabbro into earlier cumulate is exposed in the Opi River (373-380; 8°41′S, 147°51′E). The host rock is variously olivine, olivine-bytownite, or olivine-bytownite-augite cumulate with postcumulus hypersthene. The same four minerals, in different proportions, make up the intrusive gabbro. Sills of intrusive gabbro are about 1.5 m thick, and are slightly finer at margins (1 mm) than at centres (1-2 mm); textures are hypidiomorphic granular with some degree of flow-induced orientation. The orientation of the intrusive sills was probably governed by layering in the host cumulate, rather than stress.

Very fine-grained gabbro

These gabbros have average grainsize of 0.2-0.3 mm, and occur as both dykes (Pl. 13, fig. 2) and xenoliths in the coarser gabbros. The dykes have hypidiomorphic flow-oriented texture, while the xenoliths are granular and in some cases hornfelsic. Whereas the coarser gabbros contain no opaques, the very fine-grained gabbros commonly contain unusually large amounts of magnetite, up to 10 percent by volume. The xenoliths and other boulders of uncertain origin contain much magnetite, nickel-poor pyrrhotite, pyrite, and in some cases chalcopyrite. In some samples mineralization is associated with quartz-hornblende metasomatism stemming from the intrusive tonalites, but in others it looks as though the xenoliths have been foci for precipitation of metallic sulphides and oxides from the intrusive gabbroic melt.

Gabbro pegmatite

This rock forms irregular coarser phases in some gabbro exposures, and dykes up to 50 cm wide at a few localities in the Ultramafic zone near the contact with the Gabbro zone. Bytownite and pyroxene grains up to 10 cm long have been noted.

High-level Gabbro member

The High-level Gabbro member is characterized by subophitic and ophitic texture and by zoned plagioclase laths; grainsize is 1-2 mm and less commonly 3 mm. Minerals are augite and/or hypersthene, bytownite or labradorite, and, less commonly, hornblende and granophyric quartz and alkali feldspar; titaniferous magnetite is accessory.

Most outcrops of High-level Gabbro are near the contact between Gabbro and Basalt zones; this, together with textural, mineralogical, and chemical evidence, indicates that the High-level Gabbro may be a chilled phase transitional from the Granular Gabbro member to the Basalt zone. Other outcrops of high-level gabbro which are not obviously near the contact between Gabbro and Basalt zones (a) may possibly be immediately below an irregular part of the contact which has since been removed by erosion, or (b) may represent marginal phases of intrusions of granular gabbro which contributed to the development of the Granular Gabbro member.

Alteration

A small proportion of the gabbros are uralitized and saussuritized. Olivine gabbros, especially cumulates, typically show some alteration of olivine to serpentine, and some corresponding desilicification of the associated plagioclase to form hydrogrossular, prehnite, etc. Most gabbro outcrops are criss-crossed by fine veinlets (about 2 mm thick) of hornblende formed by metasomatic reaction on joint faces. Similarly, metasomatic reaction near tonalite intrusions commonly has produced hornblende oikocrysts in the gabbros.

Deformation

Most of the gabbros are not obviously deformed, but the prevalence of strong jointing is proof that they have been subjected to stress. (As a consequence of this strong jointing, gabbro outcrop areas are characterized by landslides and boulder-strewn planar slopes, in areas where rate of erosion exceeds rate of weathering. Peridotite on the other hand is not strongly jointed, and peridotite outcrop areas are characterized by gorges and convex slopes.) The lack of stress effects in the fabric of the typical gabbros probably reflects the inherent strength of these rocks, due to fine grainsize and interlocking granular texture. By way of contrast the coarser mineral grains of the gabbro pegmatites are, in most cases, deformed. Layered gabbros commonly show signs of slip along olivine-rich layers, apparently due to flowage of the olivine under stress (Pl. 14; Pl. 15, fig. 1). The weakness of the olivine-rich layers is increased, in some instances, by serpentinization (Pl. 15, fig. 2). Some streaky gabbros (Pl. 13, fig. 1) are probably deformed layered cumulates.

A few samples of homogeneous gabbro from the northern Bowutu Mountains have textures which suggest partial recrystallization. Plagioclase grains, in particular, tend towards untwinned polyhedra with apparent 120° triple-point junctions (e.g. 458; 7°28′S, 147°09′E).

Gabbro mylonites crop out on the Gira Fault at 8°12′S, and at several localities in the Awariobo Range near 9°50′S.

Pulses of gabbro intrusion

Three intrusive relationships have been observed within the Gabbro zone:

1. Homogeneous gabbro intrudes cumulus gabbro
2. Homogeneous gabbro intrudes very fine-grained gabbro
3. Very fine-grained gabbro intrudes homogeneous gabbro.

Thus, after the formation of the first cumulates there were at least two magmatic events, probably more. The fine grainsize (1-2 mm) of the cumulus and homogeneous gabbros suggests that both may have been emplaced as a succession of small rapidly cooled intrusions. On the other hand the concentration of early crystallizing cumulus phases (olivine, pyroxene) in the lower levels of the Granular Gabbro member, and of late crystallizing postcumulus phases (hornblende and, in one instance, quartz, biotite, and andesine) in the higher levels, suggests crystallization of the cumulates from a single large (3-4 km thick) intrusion.

Contact between Ultramafic and Gabbro zones

Contacts are of three types:

1. intrusive, gabbro intruding ultramafics, or
2. faulted, or
3. transitional from gabbroic cumulate into ultramafic cumulate.

Intrusive contacts have been observed at three localities: Paiawa River (7°34′S, 147°08′E), Opi River (8°43′S, 147°50′E), and Adau River (9°53′S, 148°58′E. In each case metasomatic alteration of the host rock has made it difficult to determine whether the host ultramafics are Cumulus or Noncumulus member. The contact is faulted in the Awariobo Range at 9°54′S, 148°51′E, and is probably faulted in the Ajura Kujara Range and Otavia Range at 8°35′S - 8°45′S. Transitional contacts have not been seen in place, but can be inferred from the similar mineralogy of gabbroic and ultramafic cumulates, and the proximity of outcrops of the two types, for instance, in the Wele River (7°42′S).

Shallow-dipping contacts can be assumed at several localities where creek bed outcrop is ultramafic and boulders from adjacent hillsides are gabbroic; poor outcrop makes it difficult to define contacts better than this. A traverse across the Ajura Kujara Range at 8°43′S touched the contact at two points and, if we assume that the contact is planar, these points indicate a dip of 15°E. Although the overall dip may be correct, it is unlikely that the contact is planar in detail; in plan view it is very irregular (see Map).

Contact between Gabbro and Basalt zones

The boundary is based on (a) the change from basalt and dolerite (typically uralitized and silicified) to ophitic gabbro (fine-grained, but with no microcrystalline groundmass); and (b) the topography of the Basalt zone with its distinctive cliffs, dip-slopes, and sharp ridges with concave slopes (Pl. 16). The boundary can be sharply defined in some areas, but in others it may be gradational. It is commonly obscured by later tonalite intrusions, or alteration associated with these intrusions. The contact commonly dips at an estimated 30°E, but flattens in the southern Bowutu Mountains; for instance, pillow lavas near the contact on the Waria River appear to be almost horizontal. At one point in the Sakia River (7°41′S, 147°17′E), pillow lavas form cliffs at about 1000 m above sea level and gabbro intruded by tonalite crops out in the river bed, 800 m below.

It has been previously reported that gabbro intrudes basalt (Davies, 1968), but this may not be true. Fine-grained xenoliths in gabbro have been found to have granular texture and, though it is possible that they are recrystallized basalt, they may equally well be very fine-grained gabbro.

Basalt zone

The Basalt zone consists predominantly of massive basalt and of basaltic and spilitic submarine lavas. In the Bowutu Mountains it grades upward into dacitic pyroclastics, and in other areas it includes some fine-grained calcareous sediments. These volcanic and sedimentary units are probably Cretaceous in age. The units have formal stratigraphic names (Smith & Green, 1961; Dow & Davies, 1964), but these will be ignored here for the sake of simplicity, and to accent my interpretation that the volcanics are an integral part of the Papuan Ultramafic Belt. The Basalt zone extends for the full length of the Ultramafic Belt but is partly concealed by middle Miocene and younger volcanics and sediments. Some of the middle Miocene volcanics* are difficult to distinguish from Cretaceous Basalt zone rocks, but the latter may be recognized by greater alteration, the presence of Eocene tonalite intrusions, and, in a very few localities, diagnostic foraminifera in associated sediments.

Age

The age of the Basalt zone rocks is probably Cretaceous or Upper Cretaceous. R. W. Page (pers. comm., 1968) has established an age of 116 m.y. (lower mid-Cretaceous) for pyroxenes from a coarse-grained basaltic lava flow in the Waria River (sample 972; 8°01′S, 147°34′E). He regards this as a maximum age because the mineral has low potassium content (0.067%) and possibly contains excess radiogenic argon. The basalts are older than the Eocene tonalite (50-55 m.y.), which intrudes them, and thus are probably Cretaceous. D. J. Belford (pers. comm., 1966) has found possible Upper Cretaceous foraminifera remnants in sheared marl at 8°07′S, 147°20′E and Upper Cretaceous *Globotruncana* in marl associated with pillow lavas in the extreme southeast at 9°58′S, 148°54′E.

Thickness

The average thickness of the Basalt zone is 4-6 km if one assumes an overall easterly dip of 30°. Observed dips of lava surfaces range from horizontal to 40° north and east and probably reflect minor block-faulting superimposed on an overall easterly dip. The computed thickness is in rough agreement with partial sections observed in the field. Cliff sections in the Lower Waria River (8°00′S, 147°28′E) consist of a thickness of about 1 km of pillow and massive lavas, and field observations north and east of these cliffs suggest another 1-2 km of basaltic lavas.

Stratigraphic succession

The lower part of the Basalt zone is massive dolerite and basalt in some areas (e.g. Sisa River, 8°37′S, 147°50′E), dolerite intruded by parallel basalt dykes in one area (8°30′S, 147°47′E), and simply pillow lavas in another (the Lower Waria River). Pillow lavas are the predominant rock type of the Basalt zone and are particularly well exposed in the southeastern Bowutu Mountains. Compositions are basaltic and spilitic. Thickness of flows is difficult to determine except where pillows are interbedded with massive lava, e.g. Mo River (7°52′S, 147°28′E). In that area flows are typically about 6 m thick. Less common rock types are autobreccia (7°41′S, 147°20′E), agglomerate (9°22′S, 148°35′E; Smith & Green, 1961) and hyaloclistite (8°40′S, 147°35′E). Higher

* Jointed lava and pillow lava south and southeast of the Papuan Ultramafic Belt (Mp on Pl. 1, 9°25′-10°S) are now thought to be Eocene rather than Miocene. This follows a revision by D. J. Belford of previous microfossil determinations.

† D. J. Belford has identified probable Eocene planktonic foraminifera in sediments associated with the dacitic volcanics. It follows that some or all of the dacitic volcanics may be Eocene and may be genetically related to the Eocene tonalite intrusions.

in the section one finds rare keratophyre and, in part of the Bowutu Mountains (7°32′-8S), an increasing proportion of dacitic lava, agglomerate, and tuff†. Limited observations in the Aiwa Creek/Maiama area (7°38′S) suggest that dacite may form as much as 50 percent of the section there. In other parts of the Bowutu Mountains and throughout the remainder of the Ultramafic Belt it seems to be completely absent.

Fine-grained calcareous sediments are mainly restricted to the imbricated fault slice between the main body of the Ultramafic Belt and the underlying sialic metamorphics, which suggests they may originally have been deposited in only the western or near-shore part of the Basalt zone.

The volcanics may have erupted from fissures or pipes. Incomplete circular drainage patterns which may outline small eroded volcanic cones have been noted in two areas (7°48′S, 147°28′E and 8°13′S, 147°40′E); they are 2 and 3 km across. On the other hand, parallel basalt dykes such as might accompany a fissure eruption have been found at 8°30′S, 147°47′E. A few basalt dykes cut the Gabbro zone and, in one area, the Ultramafic zone. These may be related to the Cretaceous Basalt zone vulcanism or to middle Miocene and younger volcanics.

Alteration and metamorphism

Most of the Basalt zone rocks are altered to some extent. Most are uralitized, and many are silicified, saussuritized, chloritized, epidotized, or albitized to varying degrees. Much of the alteration may be deuteric or from reaction with sea water. On the other hand the introduction of silica, epidote, pyrite, and chalcopyrite is probably related to the later tonalite intrusions.

Most of the Basalt zone rocks have not been subjected to regional metamorphism higher than zeolite facies, except that the basalts caught up in the fault slice between the Ultramafic Belt and the sialic metamorphics are partly metamorphosed to prehnite-pumpellyite facies, and, along the Timeno fault system, to greenschist, amphibolite, and granulite facies.

Tonalite

Tonalite (hornblende-quartz diorite) intrudes Gabbro and Basalt zone rocks as stocks up to 5 km across in the Bowutu Mountains and as smaller stocks throughout the length of the Ultramafic Belt. The intrusions are commonly localized on the gabbro-basalt contact as though the basalt had acted as an impermeable cap rock for the ascending tonalitic magma.

Some or all of the tonalites are Eocene (50-55 m.y.) in age, and thus are probably not genetically related to the other Ultramafic Belt rocks. The Eocene age is based on four K-Ar determinations (A. W. Webb, pers. comm., 1966; R. W. Page, pers. comm., 1968) on plagioclase and hornblende mineral separates from samples of two stocks 30 km apart (sample 21: 7°29′S, 147°13′E; sample 1228: 7°48′S, 147°24′E). It is possible that the tonalites are related to the dacitic volcanics of the Basalt zone, for the dacites could conceivably be Eocene instead of Cretaceous. It is also possible that some of the smaller tonalite stocks are Cretaceous rather than Eocene, and represent intercumulus fluid sqeezed off from the gabbro cumulates. One area where this is suggested is at 8°25′S, 147°45′E, where norite with postcumulus hornblende-andesine-quartz-biotite-muscovite (sample 925) gives way to tonalite with albitized plagioclase (samples 1086-1090) over a distance of about 1 km.

Typical tonalite consists of zoned andesine or oligoclase (sometimes albitized), quartz, and green hornblende in variable proportions, with plagioclase and hornblende tending to be euhedral. Titaniferous magnetite may form large skeletal anhedra, and a few samples contain zircon and apatite. A little biotite is present in two samples. Several thin sections contain quartz phenocrysts with the high-temperature form; their presence indicates rapid chilling of a crystal-bearing silicic magma, and argues against granophyric or intercumulus origin.

Xenoliths in tonalite are silicified and mafic minerals tend to be made over to hornblende; some xenoliths have hornfelsic texture. A common contact effect in the gabbro and basalt country rock is the growth of hornblende oikocrysts. Quartz-epidote-pyrite-chalcopyrite mineralization in basalt is thought to be related to the tonalites, because the same minerals are seen as late-stage alteration of, and veins in, the tonalite stocks, for example at Kui.

Basic tonalite and diorite.

Basic tonalite is found on Uriwa and Etewara Creeks (8°38′S, 147°23′E; 8°44′S, 147°26′E; sample 1523), where it appears to grade to diorite (sample 1532). The major minerals are euhedral augite and labradorite and matrix minerals are quartz-albite granophyre, chlorite, epidote, and, very rarely, a little potash feldspar. The diorite has chemical affinities with the Cretaceous basalts (Table 2), and is perhaps a granophyric differentiate. The basic tonalite might also be a granophyric differentiate of the basalts, or, alternatively, a hybrid produced when Cretaceous(?) diorite was assimilated by Eocene tonalite.

CHEMISTRY AND PETROGENESIS

In view of the great size of the Papuan Ultramafic Belt the 32 chemical analyses (Tables 2 and 3) should be regarded as a chemical reconnaissance rather than a thorough chemical survey. Plutonic rocks are adequately represented by a variety of rocks from all parts of the complex, but rocks transitional from plutonic to volcanic (the high-level gabbros and the dolerites of the Basalt zone) are represented by only one sample, and 8 out of 9 samples of the Basalt zone are from one general area, the Bowutu Mountains (see Fig. 4). These inadequacies of sampling should be borne in mind in the following discussion of chemical traits and petrogenesis.

The noncumulus ultramafics are chemically typical of alpine peridotites, with high Mg/Fe and low CaO and Al_2O_3 (both less than 1 percent). The cumulus ultramafics have higher CaO (up to 2 percent) and Al_2O_3 (up to 1.6 percent) and distinctly lower but variable Mg/Fe.

The gabbros are rich in Ca and Mg and poor in alkalis, especially K (Tables 2 and 3); in this they resemble gabbros from other alpine-type peridotite-gabbro complexes (Thayer & Himmelberg, 1968; Thayer, 1969b). There are no obvious chemical differences between samples from different parts of the complex, nor between such diverse rock types as homogeneous intrusive gabbro (532), cumulus gabbro (862), partly recrystallized gabbro (458), and pegmatitic gabbro (123). The only gabbros which differ markedly from the average gabbro composition are the melanorite, 27 (possibly a cumulate), and the high-level gabbro, 144. The high-level gabbro has lower Mg/Fe and higher Na content.

TABLE 2. CHEMICAL ANALYSES AND CIPW NORMS

	3A	46	08	110	117	122	15	27	123	144	145
SiO_2	37.8	42.8	55.2	54.3	54.1	55.3	48.7	52.5	49.8	50.5	48.6
TiO_2	—	.02	.02	.07	.05	.04	.07	.11	.09	.13	.11
Al_2O_3	.50	.23	.71	1.59	1.17	.99	19.4	9.7	14.4	18.3	17.2
Fe_2O_3	2.30	.99	3.0	3.05	2.15	1.03	1.32	3.6	.55	.94	4.1
FeO	5.15	6.45	3.9	8.65	9.95	7.15	3.7	3.9	4.75	5.55	.8
MnO	.10	.10	.16	.23	.18	.19	.10	.17	.13	.13	.09
MgO	46.0	48.0	34.1	28.8	30.0	32.1	9.75	19.5	14.7	9.05	10.0
CaO	.68	.55	1.96	2.0	1.10	1.36	14.1	8.95	13.9	12.8	15.4
Na_2O	.07	.07	.05	.07	.02	.04	.81	.55	.5	1.49	.01
K_2O	.03	.01	.02	.11	.01	.01	.02	.02	.09	.03	.08
P_2O_5	.01	—	.01	.01	.01	.01	.01	.01	.01	.01	.01
H_2O+	6.40	.24	1.04	.42	.42	1.07	.96	.29	1.26	.70	2.80
H_2O-	.34	.19	.08	.13	.14	.11	.33	.22	.2	.08	.23
CO_2	.26	.05	.06	.76	.06	.11	.11	.08	—	.04	.57
	99.64	99.70	100.31	100.19	99.36	99.50	100.01	99.60	100.38	99.03	100.00
SiO_2	40.8	43.1	55.7	54.9	54.8	56.3	49.7	53.0	50.3	51.1	50.4
TiO_2	—	.02	.02	.07	.05	.04	.07	.11	.09	.13	.11
Al_2O_3	.54	.23	.72	1.61	1.18	1.01	19.8	9.8	14.6	18.5	17.8
Fe_2O_3	2.48	1.00	3.02	3.08	2.18	1.05	1.35	3.64	.56	.95	4.35
FeO	5.56	6.50	3.93	8.75	10.07	7.28	3.78	3.94	4.80	5.61	.83
MnO	.11	.10	.16	.23	.18	.19	.10	.17	.13	.13	.09
MgO	49.7	48.4	34.4	29.1	30.4	32.7	9.95	19.70	14.86	9.15	10.37
CaO	.73	.55	1.98	2.02	1.11	1.38	14.4	9.04	14.05	12.94	15.98
Na_2O	.08	.07	.05	.07	.02	.04	.83	.56	.51	1.51	.01
K_2O	.03	.01	.02	.11	.01	.01	.02	.02	.10	.03	.08
Q	—	—	—	2.353	—	—	1.219	3.642	—	.967	7.783
c	—	—	—	—	—	—	—	—	—	—	—
or	.187	.059	.118	.651	.060	.060	.120	.119	.537	.177	.487
ab	.627	.593	.424	.593	.171	.344	6.981	4.684	4.276	12.649	.087
an	1.020	.285	1.658	3.710	3.104	2.533	50.153	24.099	37.176	43.299	48.104
wo	.311	.891	3.191	.566	.818	1.483	8.484	8.360	13.551	8.389	11.229
en	2.689	10.725	83.749	71.914	74.596	79.921	24.733	48.888	31.225	22.614	25.681
fs	.168	1.012	4.486	13.889	17.073	12.703	5.828	4.212	7.161	9.473	—
fo	83.214	76.458	.963	—	.613	.922	—	—	4.048	—	—
fa	5.745	7.951	.057	—	.155	.162	—	—	1.023	—	—
mt	3.535	1.439	4.360	4.434	3.149	1.517	1.949	5.254	.806	1.367	2.665
il	—	.038	.038	.133	.096	.077	.135	.210	.173	.248	.215
ap	.025	—	.024	.024	.024	.024	.024	.024	.024	.024	.024

03A Dunite, 30% serpentinized; Buiawim River, 7°04′S, 147°04′E.
46 Harzburgite; Paiawa River, 7°34′S, 147°12′E.
08 Orthopyroxenite, some olivine; Bitoi River, 7°13′S, 147°01′E.
110 Orthopyroxenite; SE of Umwate, 9°17′S, 148°15′E.
117 Orthopyroxenite; Daiuwa River, 9°19′S, 148°16′E.
122 Orthopyroxenite; Daiuwa River, 9°19′S, 148°17′E.
15 Gabbro; Buiasin Creek, 7°10′S, 147°02′E.
27 Gabbro; Saia River, 7°23′S, 147°06′E.
123 Pegmatitic phase in layered gabbro; Daiuwa Creek, 9°19′S, 148°17′E.
144 High-level gabbro; Imo River headwaters, 9°17′S, 148°24′E.
145 Gabbro, some prehnite-calcite alteration; Imo River, 9°17′S, 148°24′E.

TABLE 2. CHEMICAL ANALYSES AND CIPW NORMS (continued)

	353	384	458	532	582	862	1591	1	14	176	1139
SiO_2	50.5	48.1	48.3	50.3	49.5	51.0	49.3	49.6	48.4	50.3	54.7
TiO_2	.14	.10	.13	.11	.13	.19	.06	1.36	1.45	1.44	.70
Al_2O_3	15.5	16.5	16.7	15.0	19.8	13.8	17.0	13.5	12.5	13.5	14.5
Fe_2O_3	.54	.69	1.45	1.95	1.50	1.02	.76	2.15	2.40	2.5	4.15
FeO	3.80	3.35	3.25	4.85	3.50	5.10	3.15	8.85	9.45	7.65	6.40
MnO	.11	.08	.09	.13	.10	.13	.11	.18	.21	.21	.10
MgO	11.5	11.9	11.1	12.4	8.70	13.6	12.7	7.85	7.9	7.55	5.45
CaO	16.6	16.9	17.6	13.7	14.2	13.0	14.6	10.6	11.0	8.9	9.65
Na_2O	.76	.55	.66	.75	1.26	.92	.38	3.0	1.99	3.65	2.80
K_2O	.03	.06	.02	.02	.01	.03	.01	.13	.04	.03	.09
P_2O_5	.01	.02	.01	.01	.01	.01	.01	.13	.12	.12	.08
H_2O+	.15	1.25	.45	.49	.86	.91	1.36	2.20	3.70	3.50	.94
H_2O-	.19	.07	—	—	.14	.15	.16	.57	.34	.30	.10
CO_2	.02	.11	.03	.07	.08	.01	.13	.16	.16	.08	.03
	99.85	99.68	99.79	99.78	99.78	99.87	99.72	100.28	99.66	99.73	99.69

SiO_2	50.8	49.0	48.6	50.7	50.2	51.6	50.3	51.0	50.8	52.5	55.5
TiO_2	.14	.10	.13	.11	.13	.19	.06	1.40	1.52	1.50	.71
Al_2O_3	15.6	16.8	16.8	15.1	20.1	14.0	17.3	13.9	13.1	14.1	14.7
Fe_2O_3	.54	.70	1.46	1.97	1.52	1.03	.77	2.21	2.52	2.6	4.21
FeO	3.80	3.41	3.27	4.88	3.55	5.16	3.21	9.10	9.91	7.99	6.49
MnO	.11	.08	.09	.13	.10	.13	.11	.19	.22	.22	.10
MgO	11.5	12.11	11.18	12.5	8.81	13.8	12.9	8.07	8.3	7.89	5.53
CaO	16.7	17.2	17.7	13.8	14.4	13.2	14.9	10.9	11.5	9.3	9.79
Na_2O	.76	.56	.67	.76	1.28	.93	.39	3.1	2.09	3.81	2.84
K_2O	.03	.06	.02	.02	.01	.03	.01	.13	.04	.03	.09

Q	—	—	—	1.371	1.177	.28	.643	—	2.126	—	10.746
c	—	—	—	—	—	—	—	—	—	—	—
or	.178	.356	.119	.119	.060	.179	.060	.787	.246	.184	.539
ab	6.462	4.672	5.621	6.391	10.801	7.879	3.274	25.993	17.588	32.222	24.024
an	38.984	42.542	42.824	37.772	48.978	33.843	45.465	23.537	26.172	21.245	27.105
wo	18.197	17.035	18.708	12.597	9.351	13.097	11.434	11.860	12.089	10.022	8.730
en	28.034	21.884	20.954	31.103	21.953	34.282	32.206	16.236	20.550	18.092	13.763
fs	6.367	4.109	3.593	7.408	5.227	8.553	5.358	10.348	13.880	9.617	7.457
fo	.524	5.514	4.816	—	—	—	—	2.651	—	1.069	—
fa	.131	1.141	.910	—	—	—	—	1.862	—	.626	—
mt	.787	1.004	2.116	2.847	2.203	1.497	1.122	3.192	3.634	3.782	6.101
il	.267	.190	.248	.210	.250	.365	.116	2.645	2.876	2.853	1.348
ap	.024	.047	.024	.024	—	.024	.024	.315	.297	.296	.192

353 Intrusive gabbro near ultramafic contact; Opi River, 8°43'S, 147°50'E.
384 Olivine gabbro; Opi River, 8°41'S, 147°51'E.
458 Granoblastic gabbro; Bolu Creek, 7°29'S, 147°10'E
532 Intrusive gabbro; Sou River, 7°35'S, 147°13'E.
582 Gabbro, igneous flow texture; 9°52'S, 149°59'E.
862 Gabbro, 3 phase cumulate; Sisiworo River, 9°21'S, 148°27'E.
1591 Gabbro; Opi River headwaters, 8°45'S, 147°48'E.
01 Basalt, some prehnite, pumpellyite; Salamaua, 7°04'S, 147°04'E.
14 Basalt; Buiasin River, 7°10'S, 147°02'E.
176 Basalt, pillow lava(?); Giumu River, 8°08'S, 147°22'E.
1139 Basalt, uralitized, silicified; Sisa River, 8°37'S, 147°50'E.

TABLE 2. CHEMICAL ANALYSES AND CIPW NORMS (continued)

	1254	1578	1586	94	97	21	1523	1576	1585	1532	987
SiO_2	50.5	49.9	49.5	47.7	48.6	60.8	56.7	65.2	58.9	54.3	64.3
TiO_2	.75	1.49	1.24	1.7	.7	.41	.25	.39	.55	.55	.3
Al_2O_3	13.5	13.7	13.6	14.2	14.7	16.2	17.7	14.5	15.0	14.2	14.7
Fe_2O_3	4.6	4.8	5.8	8.55	1.05	1.5	1.45	2.35	3.8	3.0	2.4
FeO	7.9	8.5	6.7	3.4	7.85	5.4	5.75	2.40	5.1	7.25	1.98
MnO	.22	.20	.12	.18	.17	.14	.10	.06	.14	.17	.07
MgO	7.3	6.75	7.8	5.95	9.10	3.35	3.6	3.0	3.25	6.05	3.7
CaO	11.1	10.9	8.4	11.4	13.2	7.65	8.6	5.1	6.9	.88	1.87
Na_2O	1.79	2.10	3.45	2.5	1.65	2.9	2.6	4.2	4.05	3.15	5.25
K_2O	.03	.06	.11	1.33	.11	.34	.42	1.11	.32	.08	1.12
P_2O_5	.05	.12	.11	.16	.05	.07	.03	.11	.20	.05	.06
H_2O+	1.78	1.11	2.75	2.25	2.3	8.2	2.25	1.25	1.22	2.10	2.95
H_2O-	.20	.05	.47	.19	.24	.27	.29	.13	.38	.27	1.18
CO_2	.05	.01	.06	.34	—	.12	.04	.01	.01	.01	.20
	99.77	99.68	100.11	99.85	99.72	99.97	99.78	99.80	99.81	99.97	100.08
SiO_2	51.7	50.7	51.2	49.2	50.0	61.6	58.4	66.3	60.1	55.7	67.2
TiO_2	.77	1.51	1.28	1.8	.7	.42	.26	.40	.56	.56	.31
Al_2O_3	13.8	13.9	14.1	14.7	15.1	16.4	18.2	14.7	15.3	14.6	15.4
Fe_2O_3	4.7	4.9	6.0	8.83	1.08	1.5	1.49	2.39	3.9	3.08	2.5
FeO	8.1	8.6	6.9	3.5	8.08	5.5	5.92	2.44	5.2	7.43	2.07
MnO	.23	.20	.12	.19	.18	.14	.10	.06	.14	.17	.07
MgO	7.5	6.86	8.1	6.14	9.37	3.39	3.7	3.05	3.31	6.20	3.9
CaO	11.4	11.1	8.7	11.8	13.6	7.75	8.9	5.2	7.0	9.0	1.95
Na_2O	1.83	2.13	3.57	2.6	1.70	2.9	2.7	4.3	4.13	3.23	5.49
K_2O	.03	.06	.11	1.37	.11	.34	.43	1.13	.33	.08	1.17
Q	6.724	5.618	1.396	2.648	—	19.674	13.755	22.451	15.186	7.550	20.932
c	—	—	—	—	—	—	—	—	—	—	—
or	.181	.360	.671	8.067	.668	2.029	2.494	6.664	1.925	.484	6.897
ab	15.496	18.036	30.148	21.714	14.365	24.779	22.113	36.110	34.894	27.309	46.298
an	29.377	28.196	21.995	24.221	33.315	30.477	35.566	17.713	22.202	24.970	7.943
wo	11.119	10.813	8.476	12.757	14.083	2.764	2.867	3.033	4.728	8.111	—
en	18.601	17.064	20.062	15.211	18.881	8.425	9.012	7.591	8.242	15.438	9.604
fs	10.105	9.698	5.873	—	10.601	8.209	9.181	1.964	5.679	10.494	1.342
fo	—	—	—	—	3.110	—	—	—	—	—	—
fa	—	—	—	—	1.925	—	—	—	—	—	—
mt	6.824	7.064	8.685	6.825	1.566	2.196	2.113	3.462	5.610	4.457	3.627
il	1.457	2.872	2.432	3.314	1.368	.786	.477	.752	1.064	1.070	.594
ap	.121	.288	.269	.389	.122	.167	.071	.265	.482	.121	.148

1254 Basalt, uralitized, pillows(?); Waria River, 8°02'S, 147°27'E.
1578 Basalt, uralitized, spilitic; Jema village, bldrs, 8°03'S, 147°27'E.
1586 Basalt, uralitized; Eia River, 8°06'S, 147°37'E.
94 Basic greenschist from Timeno Fault System; Labai Creek, 9°13'S, 148°06'E.
97 Amphibolite from Timeno Fault System; Di Creek, 9°14'S, 148°08'E.
21 Tonalite; near Kui, 7°30'S, 147°13'E.
1523 Basic tonalite with augite; Uriwa Creek, 7°38'S, 147°23'E.
1576 Tonalite; Jema village bldr, 8°03'S, 147°27'E.
1585 Tonalite; Eia River, 8°06'S, 147°36'E.
1532 Diorite with augite and labradorite; Etewara Creek, 7°40'S, 147°30'E.
987 Dacite lava; Woiba Island, 7°33'S, 147°24'E.

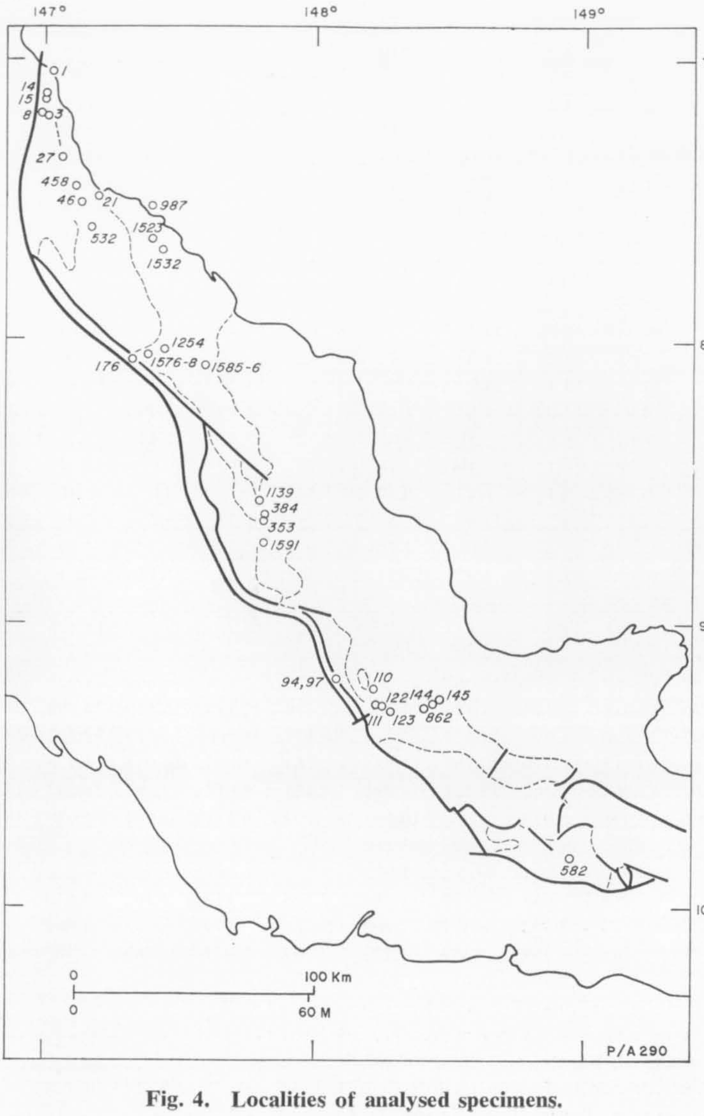


Fig. 4. Localities of analysed specimens.

TABLE 3. AVERAGE ANALYSES

	Gabbro (average of 12)	High-level Gabbro (sample 144)	Basalt (average of 7)	Tonalite (average of 4)
SiO_2	50.48	51.1	51.91	61.6
TiO_2	.11	.13	1.24	.41
Al_2O_3	16.35	18.5	13.94	16.2
Fe_2O_3	1.56	.95	3.88	2.32
FeO	3.85	5.61	8.16	4.77
MnO	.11	.13	.18	.11
MgO	12.24	9.15	7.46	3.36
CaO	14.53	12.94	10.38	7.2
Na_2O	.73	1.51	2.77	3.51
K_2O	.04	.03	.07	.56
	100.00	100.05	99.99	100.04

The basalts have definite tholeiitic affinities, but contain more SiO_2 and less TiO_2 than Manson's (1967, p. 223) average oceanic tholeiite; they are nearer in composition to Manson's average continental tholeiite, especially the average for Australia. Like the gabbros, the basalts have anomalously low K_2O. Similarly the average K_2O content of the tonalites is distinctly low for rocks so silicic.

Petrogenesis

It has already been concluded that the noncumulus ultramafics probably represent pre-existing mantle, and that some or all of the tonalites are much younger than and probably unrelated to the basalts and gabbros. The question remains: are the basalts, gabbros, and cumulus ultramafics related and if so how?

Field relations suggest that they are related. Cumulates are overlain by more rapidly cooled gabbro (High-level Gabbro) which is in turn overlain by dolerite and basalt, and contacts between these rock units appear to be transitional. The chemical analyses, however, show that the gabbros and basalts are not similar except for their low K_2O content. The purposes of the following discussion are to determine

(1) whether these dissimilar rocks might have a common parent,
(2) if so, what the composition of the parent might be, and
(3) how it was differentiated.

AFM diagram

The triangular diagram (Fig. 5) is a weight percent plot of $Na_2O + K_2O$, MgO, and MnO-plus-total-iron-as-FeO. The AFM diagram was introduced by Wager & Deer (1939) and has been generally adopted to illustrate fractionation trends in basaltic magma (e.g. Hess, 1960, pl. 11). As normal fractionation progresses, basaltic melt becomes richer in alkalis and in iron with respect to magnesia, as shown in some typical trends in Figure 6. Wager & Deer, Hess, and others have used FeO rather than total-iron-as-FeO; this is satisfactory for plutonic rocks where most iron is ferrous and the ferrous/ferric ratio remains constant or decreases systemically with fractionation. In the Papuan Ultramafic Belt, however, some of the basalts contain considerable ferric iron and the ferrous/ferric ratio is variable, so total-iron-as-FeO is preferred. This has been calculated as $FeO + O.9Fe_2O_3$.

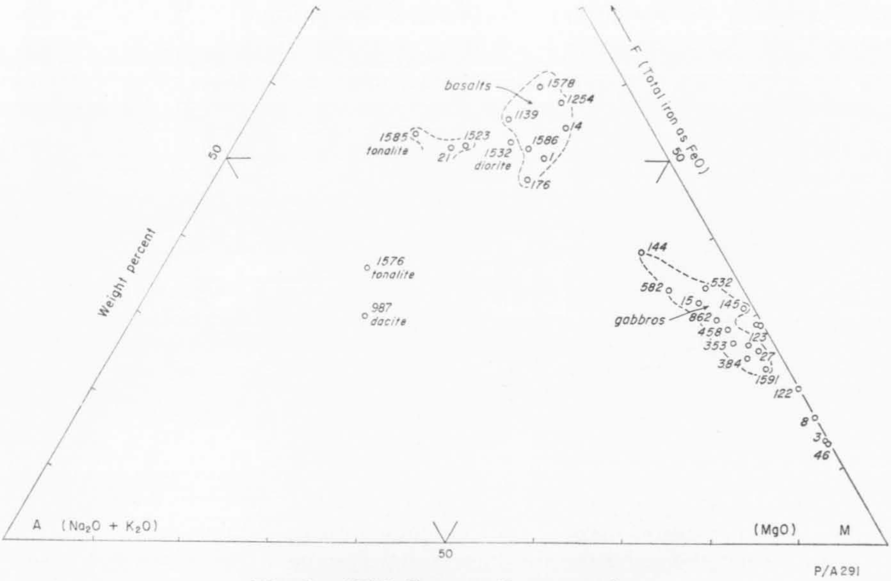


Fig. 5. AFM diagram: Papuan rocks.

Analyses of rocks of the Papuan Ultramafic Belt are plotted on Figure 5. The gabbros plot in a linear group which is aligned between the cluster of basalt analyses and those of the cumulus orthopyroxenites. This linear trend permits one to speculate that there may be a genetic relationship between the gabbros, orthopyroxenites, and basalts. Gabbros which plot nearest the ultramafics would be those most enriched in early-formed crystals such as olivine and orthopyroxene; those which plot nearest the basalts are least enriched. Appropriately, the gabbro which plots nearest the basalts is the one sample of high-level gabbro (No. 144).

If, for the moment, it is assumed that the ultramafic cumulates, gabbros, and basalts are cogenetic, and developed by crystal fractionation from a single parent magma, what is the most likely composition of that parent? Three possibilities are discussed:

1. Because about equal volumes of basalt and gabbro are present in the Papuan Ultramafic Belt, it might be argued that a parent would have a composition about midway between average gabbro and average basalt; this calculates out to a composition near that of the high-level gabbro, No. 144 (Fig. 5). A parent of this composition allows a chemical balance between the granular gabbros, which are richer in Ca and Mg, and the basalts, which are poorer.

2. It could be argued that the undifferentiated parent magma would be represented by the chemistry of the basalt lavas. If the volumes of basalt and gabbro deduced from surface exposure are correct, this would not permit a chemical balance. The only rocks poorer in Ca and Mg than the basalts are the Bowutu Mountains dacites, but (a) these are probably not present in great volume and do not extend along the length of the Ultramafic Belt, and (b) they are not likely end-products of fractionation of gabbro in regard to SiO_2, 'FeO', or TiO_2 (Figs 9, 11, 14).

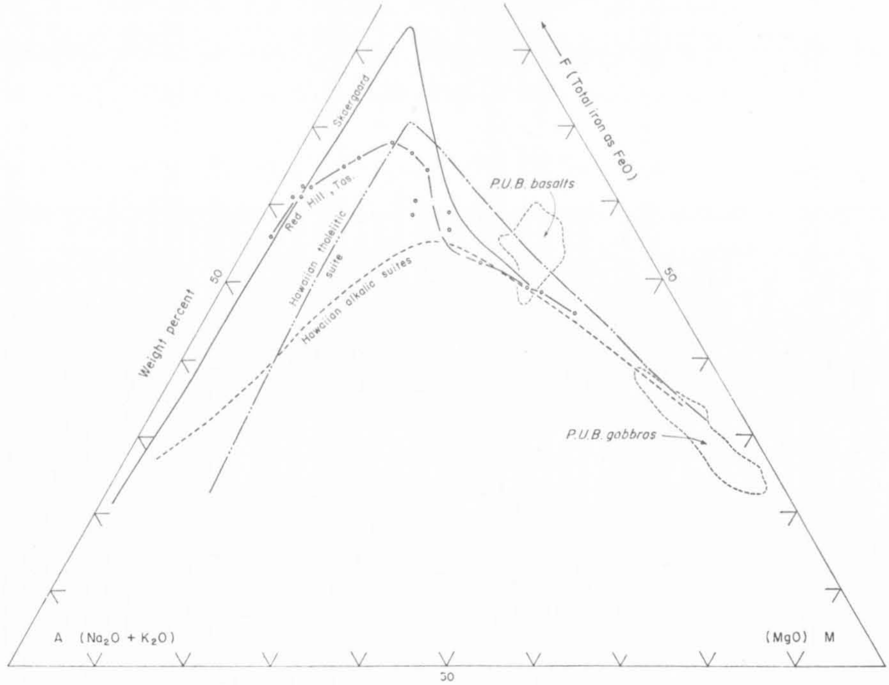


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Fig. 6. Basalt fractionation trends (after MacDonald & Katsura, 1964).

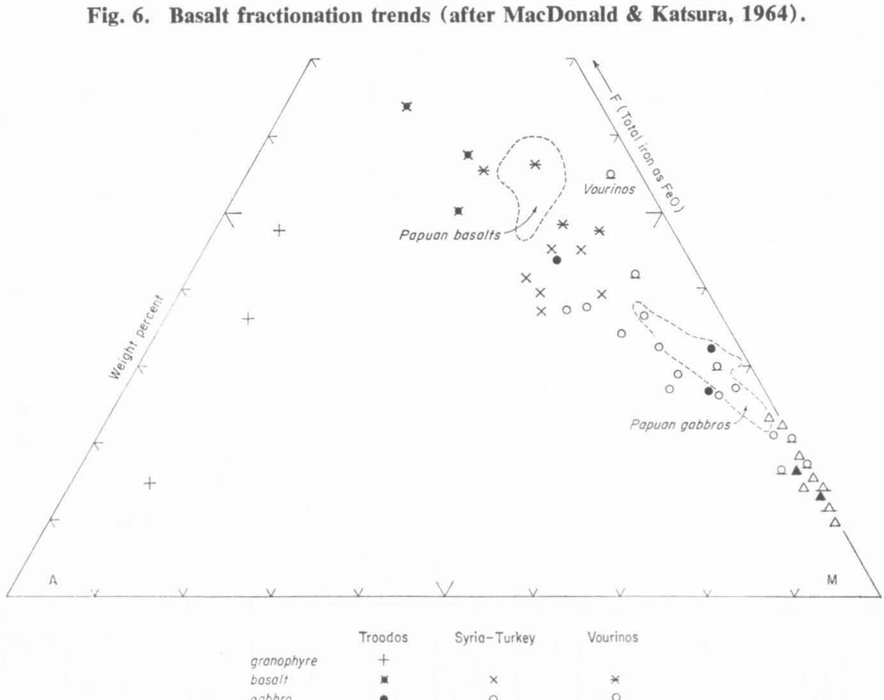

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Fig. 7. Ophiolites (after Thayer, 1967).

3. It is possible that the parent was differentiated at subcrustal levels into basaltic and gabbroic sub-parents, and these moved independently to the crust, where they were emplaced at different levels. However, this would make the simple stratification of the Papuan Ultramafic Belt an improbable coincidence, and would not explain the apparent transition from basalt through dolerite to gabbro.

Comparison with basalt and ophiolite fractionation trends

Figure 6 is an AFM diagram which compares the Papuan rocks with the fractionation trends of basaltic melts in the Skaergaard intrusion, Hawaiian tholeiitic and alkalic lavas, and Tasmanian dolerite. This illustrates the point that the Papuan rocks do not form a typical or complete basalt fractionation series. In Figure 7 the Papuan rocks are compared with two Mediterranean ophiolite sequences and the Troodos Complex of Cyprus. In each of these complexes the basalts are thought to be related to the plutonic rocks, with gradational contacts between basalt and gabbro at least in the case of the two ophiolites. The Papuan trend is similar to that of the two ophiolites, and defines what might be termed an ophiolite trend. The ophiolite trend has higher Mg/Fe and lower alkalis than the typical basaltic trends. This indicates either (a) that the parent magma had higher Mg/Fe and lower alkalis than typical undifferentiated basalt, or (b) that the iron and especially the alkali-rich residua of the ophiolites have been mysteriously removed.

Thayer (1967) has suggested that the ophiolite trend differs from the typical basalt trend because, in the case of the ophiolites, the magma differentiated by crystal settling at mantle, rather than crustal, levels, and the differentiated pluton was then emplaced in the crust as a crystal mush. However, this would not explain the consistent spatial, and possibly genetic, association of basalts with the plutonic rocks. Nor would it explain the absence of iron and alkali-rich residua.

MgO variation diagrams

Figures 8-14 are plots of each of the major rock-forming elements against MgO (after Powers, 1955). These diagrams will be used to evaluate the role of crystal settling of early-formed minerals in the evolution of the Papuan rocks, and, in particular, to assess high-level gabbro 144 as a possible parent for both basalt and gabbro.

Minerals which are known to occur as cumulus phases in the gabbros are bytownite, hypersthene, clinopyroxene, olivine, and chromite. The problem is to ascertain whether it is possible to produce a liquid of typical basalt composition (e.g. the composition of basalt O1) from parent 144 by removal, in reasonable proportions, of bytownite, hypersthene, clinopyroxene, olivine, and chromite. The solution, if there is one, should show the different minerals settling out in proportions commensurate with the mineral proportions observed in the Gabbro zone in the field, namely bytownite predominant, clinopyroxene and orthopyroxene about equal, and olivine minor. Chromite is rare and, within the limits of error of this method, it can be neglected. The solution should also show an approximately equal split of the parent into cumulate (gabbro) and residual liquid (basalt) because gabbro and basalt volumes, as judged from field mapping, are about equal. The problem may be approached graphically as follows.

1. On each of the MgO variation diagrams compositions of the four mineral phases are plotted. These are taken from R. N. England's (pers. comm., 1969) preliminary results of electron microprobe analysis, except in the case of bytownite, which is calculated as 80 percent anorthite 20 percent albite, an approximate average composition based on measurements of extinction angles in Carlsbad-Albite twins.

2. Lines are drawn from each mineral composition (or from a mean composition where more than one analysis is available) through the point 144. These lines may be termed the olivine control, hypersthene control, etc., because they indicate the direction in which the magma composition will move if that particular mineral is removed. If two or more minerals are removed simultaneously, or over a period of time, the new composition of the melt can be found by summing the vectors graphically. The direction of the vector for any mineral is always drawn parallel to the line connecting that mineral and 144, and the length of the vector is proportional to the quantity of that mineral which is removed, and the weight percent of the abscissa and ordinate oxides (MgO and, for example, Al_2O_3) which the mineral contains. Thus, if the removal of hypersthene causes the melt to become x weight percent poorer in MgO, and hypersthene contains 30 weight percent MgO, then the amount of hypersthene removed is $x \cdot \frac{100}{30}$ percent.

3. On Figure 8, lines are drawn from O1 parallel to olivine and hypersthene control. These intersect bytownite control at A and B respectively. This construction shows that, with respect to Al_2O_3 and MgO, composition O1 can be reached from composition 144 by removal of bytownite and olivine or bytownite and hypersthene alone. When the equivalent lines are plotted on the SiO_2 diagram (Fig. 11) it is found that removal of bytownite and hypersthene alone produces a rough fit, but removal of bytownite and olivine alone does not. The CaO diagram (Fig. 10), in turn, cannot be solved unless some clinopyroxene is also removed, and the best fit for the 'FeO' diagram (Fig. 9) is reached when a small amount of olivine is removed as well.

4. Thus by moving from one plot to another, and returning to each several times, a best fit is eventually reached which is represented by the vectors plotted on the diagrams. This may then be checked by a simple calculation based on the similar SiO_2 content of 144 and O1 (and, incidentally, of average gabbro and average basalt). Because both 144 and O1 have silica contents around 51 percent (51.1 and 51.0 respectively), removal from 144 of silica-poor minerals olivine (42 percent silica) and bytownite (48 percent silica) must have been balanced by removal of silica-rich hypersthene (57 percent silica). Removal of clinopyroxene (51.6 percent silica) would have had little effect and is ignored here. In terms of silica content, olivine, bytownite, and hypersthene differ from the parent magma by —9, —3, and +6 weight percent units respectively. To maintain constant silica content, one weight percent unit of hypersthene must settle out for each two weight percent units of bytownite, and three weight percent units of hypersthene for each of olivine.

From the graphical method and this calculation the following proportions of settled minerals can be deduced:

mineral	weight percent
bytownite	30
hypersthene	19
clinopyroxene	6
olivine	2

The figures are only approximate because (a) there are not enough basalt analyses to demonstrate that O1 is a typical basalt, and (b) there are not enough mineral analyses to ensure that those shown in the diagram are representative.

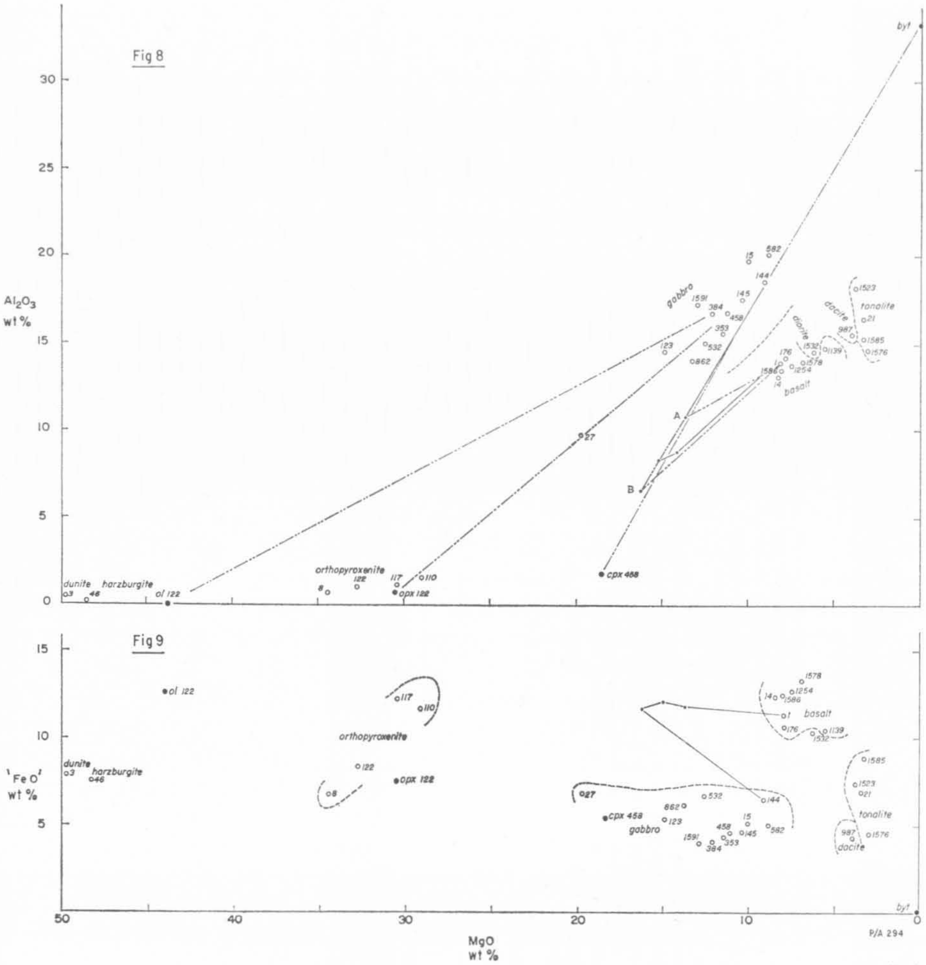
When these proportions are plotted on the Na_2O diagram (Fig. 12) there is a misfit which can be resolved only by removing crystals of anorthite An_{91} instead of bytownite An_{80} from the melt. Possibly some of the early-formed plagioclase was anorthite. (The nature of the plagioclase has been roughly established by estimating the refractive index and determining the optic sign; this method does not distinguish bytownite from anorthite. Where compositions were determined from extinction angles of Carlsbad-Albite twins they were generally around An_{80}.) Alternatively the real parent was richer in Na_2O than is gabbro 144. Misfits in the K_2O and TiO_2 plots suggest that the actual parent was richer in these elements than is gabbro 144, or that some mechanism other than crystal fractionation caused K_2O and TiO_2 to concentrate in the residual liquid. The relative enrichment of TiO_2 in the basalts is particularly striking, although, compared with worldwide averages, the basalts are not rich in titanium.

The exercise shows that 144 is a possible parent for O1, except in regard to K_2O, TiO_2, and possibly Na_2O.

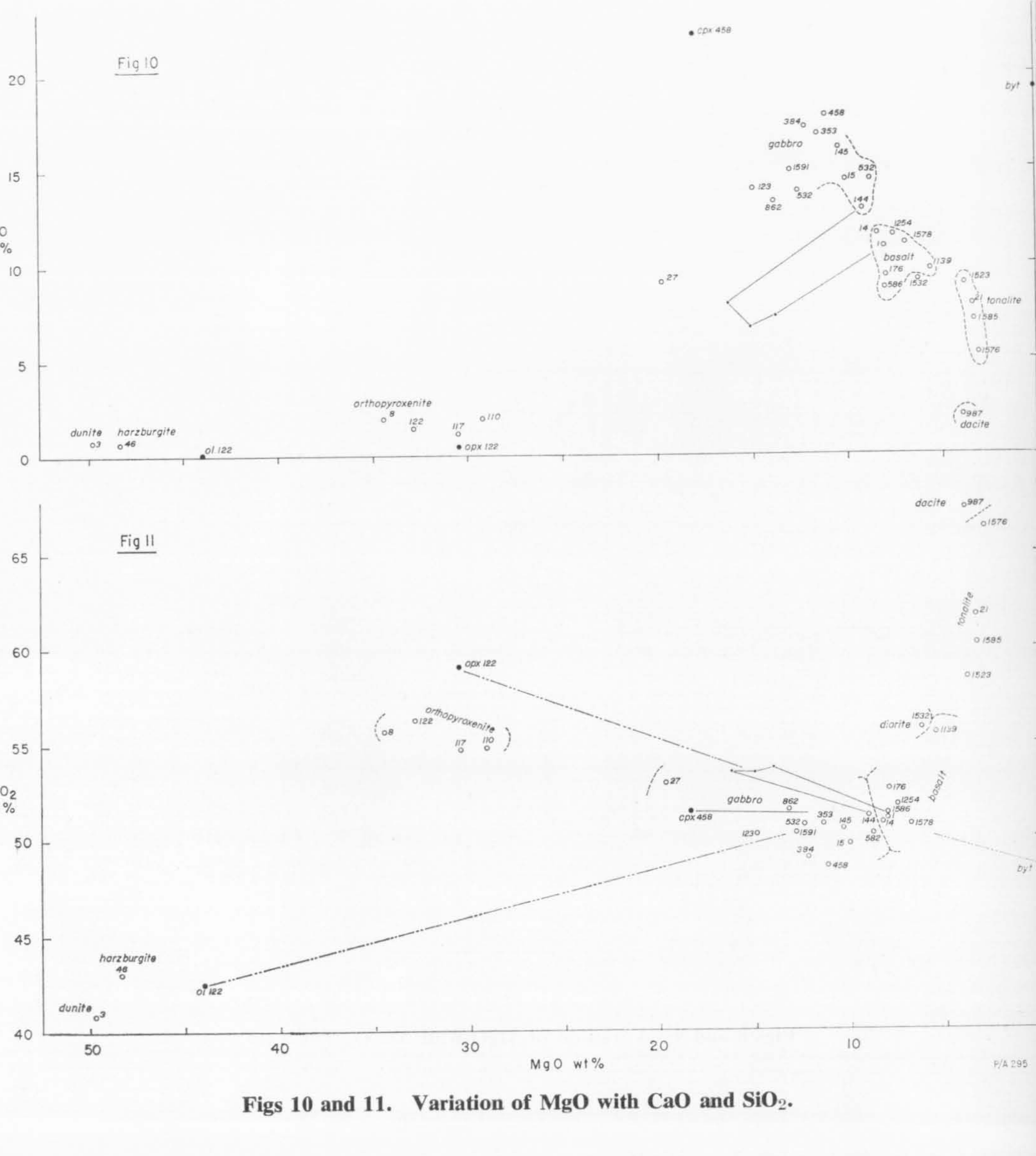
The relative proportions of the different cumulus minerals are in rough agreement with the proportions observed in the field, except for the predominance of orthopyroxene over clinopyroxene. The total weight percent of cumulus crystals, at 57 percent, is slightly higher than should be the case if equal volumes of gabbro and basalt were to be produced. This together with the apparent deficiencies in alkalis and TiO_2 suggests that the actual parent magma is not 144 but has a composition somewhere between 144 and average basalt, and that sample 144 is itself a partly fractionated rock. This must remain a hypothetical solution until further sampling of the high-level gabbros and overlying dolerites establishes whether a completely satisfactory parent exists.

Discussion

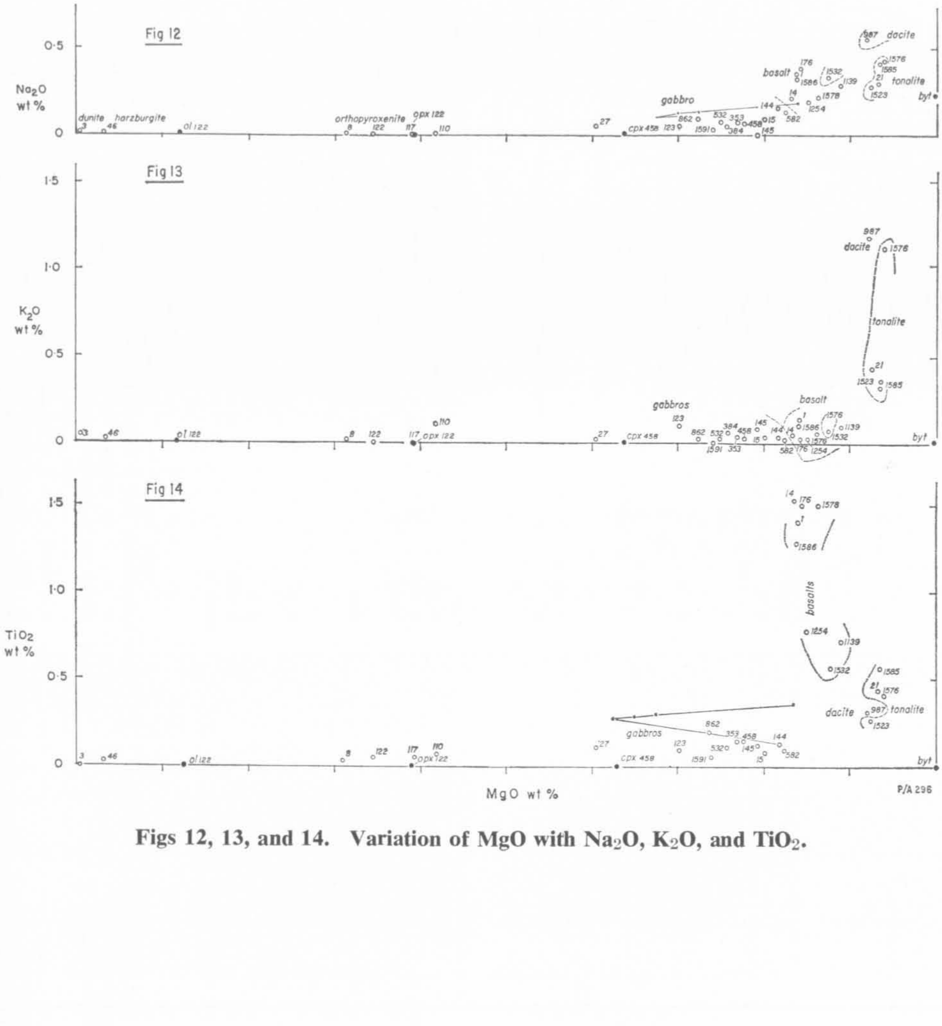
Field relations point towards derivation of ultramafic cumulates, gabbros, and basalts by crystal fractionation of a single parent magma at crustal and uppermost mantle levels (less than 10 km below sea floor). The AFM diagram (Fig. 5) accords with this conclusion. A comparison using the AFM diagram (Figs 6 and 7) shows that the Papuan Ultramafic Belt and two Mediterranean ophiolites do not follow typical basalt fractionation trends, but rather have a distinctive trend with generally higher Mg/Fe and lower alkalis. This indicates either (a)



Figs 8 and 9. Variation of MgO with AC_2O_3 and 'FeO'.



Figs 10 and 11. Variation of MgO with CaO and SiO_2.



Figs 12, 13, and 14. Variation of MgO with Na_2O, K_2O, and TiO_2.

that the parent magma had higher Mg/Fe and lower alkalis than typical undifferentiated basaltic magma, or (b) that the iron and alkali-rich residua in the ophiolite and Papuan Ultramafic Belt systems have been mysteriously removed. Graphical analysis using MgO variation diagrams demonstrates that typical basalt could have been derived from a parent magma a little more 'basaltic' than the high-level gabbro sample 144.

The weight of evidence thus seems to point towards the derivation of the Papuan basalts and gabbros from a basaltic magma, rich in Ca and Mg, by fractionation at crustal levels. There are, however, four objections to this hypothesis:

1. The supposed parent is not known to be represented in the overlying lavas.

2. If the basalts are residual liquids they might be expected to contain phenocrysts or xenoliths of distinctive cumulus crystal phases, such as bytownite and hypersthene. In fact, none has been found.

3. The supposed parent has higher Mg/Fe and lower alkali content than other commonly accepted examples of undifferentiated basaltic magma (Fig. 6).

4. The hypothesis does not explain why gabbro which intruded the earlier-formed cumulates (e.g. No. 532) has a composition very close to that of gabbro with cumulus texture (e.g. No. 862).

A more thorough sampling of the lavas may remove the first objection. A true parental rock need not be especially abundant, for the hypothesis permits most of the lavas to be residual liquids derived by fractionation of parent magma. Others working with ultramafic cumulates have inferred that supposedly associated volcanic rocks must be residual liquids, for example Brown (1956) on the island of Rhum, and Challis (1965) in New Zealand; and O'Hara (1965) has concluded that most basalts have chemical compositions and exhibit crystallization sequences determined by low-pressure crystal fractionation.

The second objection can be accounted for if it is assumed that emptying of the magma chamber during an eruption was not unusually forceful, and permitted any suspended crystals of bytownite, hypersthene, etc., to settle from the melt before eruption. Forbes & Kuno (1965) and Jackson (1968) have pointed out that the tholeiitic magmas are notably poor in xenoliths.

The third objection might be explained as a result of unusual or currently unexpected conditions at the source of the parent magma in the mantle. Perhaps a magma of this composition could be formed by a high degree of partial melting of, for instance, a clinopyroxene-rich mantle rock. Alternatively, unusual conditions might have prevailed at the time of partial melting with regard to pressure, temperature, or availability of water (cf. Green, 1969, p. 409).

The fourth objection might be explained by supposing (a) that the later intrusive gabbro was held liquid at crustal levels long enough to become enriched in Ca and Mg by ingestion of crystals settling from above, or (b) that the same process operated as the magma was rising through the upper mantle and lower crust, or (c) that the intrusive gabbro is a cumulus gabbro which was remobilized by a local increase in temperature. The similar mineralogy and chemistry of the intrusive and cumulus gabbros would suggest that they are closely related.

Evolution of the Papuan Ultramafic Belt

It seems possible, and indeed probable, that the Papuan ultramafic cumulates, gabbros, and basalts are differentiates from a single parent magma. These rocks, together with the underlying noncumulus ultramafics, make up a section of probable oceanic mantle and crust. How did it develop? The model proposed here is that the ultramafic cumulates and gabbros represent former magma reservoirs from which the basaltic volcanics were erupted. The reservoirs were fed by a continuing supply of basaltic magma from a mantle source. Reservoirs were between 4 and 10 km below the sea floor when activity ceased, and may have been less than 4 km below the sea floor when volcanic activity first started. This compares with the depth of 3 km to the reservoir under Kilauea volcano in Hawaii, which Fiske & Kinoshita (1969) have deduced from tilt measurements. The model is similar to that which Challis (1965) has proposed for the New Zealand ultramafics and Permian volcanics, with the difference that in the Papuan complex there is no evidence of continental country rock. The lack of normal basement makes it difficult to envisage the initial stages in the development of the Papuan complex, but no doubt the process is best seen as one of continuing generation of oceanic crust, such as is believed to be taking place at the midocean ridges today (Hess, 1962; Dietz, 1961; Le Pichon, 1968; Isacks et al., 1968). In this case the country rock might best be thought of as earlier-formed oceanic crust which moved away from the ridge at a rate of 2-5 cm a year. The gabbro and basalt would then have filled the tensional zone along the axis of the ridge. The noncumulus peridotite would represent mantle material that moved in the solid state, perhaps by syntectonic recrystallization. This would initially have formed a floor for the gabbroic intrusion, then would have moved with it, as a rigid plate, away from the midocean ridge.

Whether or not the Papuan Ultramafic Belt developed at a midocean ridge, the probable course of events in the development of the complex is:

1. A tensional zone developed in the earth's crust in an oceanic area sufficiently far from any eroding landmass to be free of detrital sediment.

2. Basalt magma was generated in the mantle (presumably at 50-100 km depth), rose to crustal levels, intruded the tensional zone, and reached the ocean floor.

3. The topmost levels cooled rapidly as pillow lavas and massive basalt, grading downwards into dolerite and the 'high-level gabbro'. Lower in the crust magma reservoirs cooled more slowly: early-formed crystals of olivine, orthopyroxene, and clinopyroxene settled to form first ultramafic and then gabbroic cumulate, as plagioclase joined the other cumulus phases.

4. Crystal-settling depleted the residual liquid at the top of the chamber in Mg, Ca, and Al. From time to time this liquid was expelled through the overlying basalt and dolerite to erupt as lava upon the ocean floor.

5. Periodically more magma arrived from the mantle source either to replenish the existing magma chamber, or to form a new chamber by intruding already-solidified crust. These processes continued until the gabbroic and basaltic layers were each built up to a thickness of about 4 km.

The fine grainsize of the gabbros might indicate that the gabbro zone is made up of a great many small intrusions, and this is supported to some extent by the many field observations of gabbro intruding gabbro. If this is so, reservoirs would probably have migrated upwards as the overlying lavas became thicker, as Fiske & Kinoshita (1969) have suggested for Kilauea. On the other hand, the apparent concentration of ultramafic cumulates near the bottom of the zone, and the apparent concentration of water and potash-rich intercumulus material in the top of the zone (in the few instances where such material has been found), would suggest that in any one area the intrusion was a single lopolith up to 4 km thick. More detailed mapping should resolve this question.

DISCUSSION

The Papuan Ultramafic Belt is similar to Mediterranean ophiolites, alpine-type peridotite-gabbro complexes, and rocks beneath the oceans, as far as can be judged from samples available. These similarities are discussed below, and some hypotheses of origin of ophiolite and alpine-type complexes briefly reviewed.

Comparison with ocean-floor samples

Within the last few years dredge hauls from the ocean floors have recovered all the main rock types that make up the Papuan Ultramafic Belt. Thayer (1969a) discussed some of the samples and noted their similarity to rocks of the peridotite-gabbro complexes. Other discoveries which show similarities specifically to the Papuan rocks are:

1. Cross-fractures on the Mid-Atlantic Ridge between 2°N and 2°S show peridotite on lower slopes, basalt on upper slopes, and gabbro and greenschist between. A thickness of 3500 m of ultramafics is exposed in part of one of these, the Romanche Fracture (Bonatti, 1968).

2. Also in the Romanche Fracture, gabbro associated with peridotite has cumulus texture; ultramafic rocks from this and other North Atlantic localities look like 'mainly mobilized nearly crystallized cumulates' (Melson, 1969).

3. At 45°N in the Atlantic the Geological Survey of Canada has dredged dunite, harzburgite, gabbro, troctolitic gabbro, and amphibolic peridotite, many of which show crystal layering; also hornblende-rich quartz diorite and 'almost hornblende-free trondhjemite' (Aumento, 1969a, 1969b).

4. On the Mid-Indian ridge serpentinized peridotite is intruded by dolerite and gabbro; the main ultramafic rock type is serpentinized and fractured harzburgite (Udintsev, 1969).

5. A dredge haul from the East Pacific Rise contains both typical tholeiitic basalt and a 'differentiated andesitic variety' (Hart, 1969), which to judge from Hart's partial analysis might better be termed dacite.

Metamorphosed basalts of greenschist, amphibolite, and higher grades have been found on the floor of the North Atlantic and Indian Oceans (Melson & Van Andel, 1966; Miyashiro, 1969; Cann, 1969; Aumento, 1969b). These are thought to be up-faulted blocks of material which has been subjected to simple burial metamorphism, though some shearing and mylonitization is noted by Melson &

Van Andel. Metamorphosed basalts of greenschist, amphibolite, and pyroxene granulite facies are found along the basal Timeno fault system of the Papuan Ultramafic Belt, but these are products of dynamic rather than burial metamorphism.

Comparison with Mediterranean ophiolites

The ophiolite suite defined in Tuscany by Steinmann (1927) consists of basal serpentinite overlain by gabbro, which is in turn overlain by diabase-spilite with some pillow structures. The suite was extended and elaborated by Aubouin (1965), in a review of the ophiolites of Greece, Turkey, and Syria. Aubouin's tabulation of the ophiolite sequence is reproduced below; units are listed in descending order, and approximate typical thicknesses (A. Nicolas, pers. comm., 1968) are shown.

1. Fine-grained rocks:	Basalt, spilite, pillow lavas	300-400 m
2. Medium-grained rocks:	Dolerite	500-1000 m
3. Coarse-grained rocks:	Quartz diorite (occasionally)	
	Diorite	
	Gabbro	1000 m
	Pyroxeno-peridotite	
	Pyroxenite	2000-2500 m
	Peridotite	

The rock types and the sequence are similar to the Papuan Ultramafic Belt, but the typical total thickness, 4-5 km, is much less. Like the Papuan complex, the ophiolites show gabbro intruding peridotite (Steinmann, 1927; Vuagnat, 1963) and gabbro transitional upwards into dolerite and basalt. Cumulus textures may or may not be preserved in the ophiolites. One illustration in an account of the ophiolites of northwest Syria and adjacent Turkey is of an olivine cumulate with postcumulus clinopyroxene and altered plagioclase (Dubertret, 1955, pl. XI, fig. 4), although the cumulus texture is not discussed as such in the text. A similar cumulate is illustrated in an account of the Oman ophiolite, although it too is not described as such in the text (Reinhard, 1969, figs 4 and 5). Like the Papuan Ultramafic Belt, the cumulus layered gabbro of the Oman ophiolite was intruded by a later gabbro.

Ophiolites are not restricted to the Mediterranean (e.g. Gansser, 1959, 1964; Bezore, 1969), nor are they restricted in time. For instance, McCall & Doepel (1969) have described ophiolite sequences in Western Australia which are at least 2700 m.y. old. These are made up of ultramafic cumulates (harzburgite, bronzitite) overlain by gabbroic cumulates with some granophyre, which are in turn overlain by pillowed metabasalt. One complete sequence, 1050 m thick, is overlain by another, about 600 m thick, and the sequences are discontinuous laterally as though they might be former seafloor volcanoes.

Comparison with Troodos Complex, Cyprus

The Troodos Complex of Cyprus (Wilson & Ingham, 1959; Bear, 1963) consists of:

Pillow Lavas (Upper and Lower)		Basalt, relatively unaltered
Basalt Group		Pillow lavas of altered basalt with many dykes
Sheeted Intrusive Complex (Diabase)	about 2000 m	Basalt dyke swarm, north-south, steeply dipping; ratio of dykes to host is about 10:1
Granophyre		Diorite, quartz diorite, trondhjemite, granodiorite; small irregular bodies
Gabbro	about 500 m	Gabbro, olivine gabbro, norite
Ultramafics	thickness not known	Peridotite (mostly harzburgite), dunite, serpentinite

(Thicknesses are estimates based on Wilson & Ingham's sections.)

This sequence is similar to the ophiolite and Papuan Ultramafic Belt sequences except for one rock unit, the Sheeted Intrusive Complex or dyke swarm. Some of the Troodos ultramafic rocks are cumulates, though they apparently have not been recognized as such. Poikilitic harzburgite and wehrlite illustrated in two photomicrographs (Wilson & Ingham, pl. V, figs 2 and 3, p. 84) look like olivine cumulates with postcumulus enstatite and diopside respectively. The subhedral olivine grains and the poikilitic nature of the pyroxenes are diagnostic. In the harzburgite some of the olivine grains have partly reacted to form enstatite. Cumulus textures in rocks from the southeastern part of the complex are illustrated by Pantazis (1967), but again are not described as such in the accompanying text.

Gass (1968) has proposed that the Troodos Complex is oceanic mantle and crust and that the dyke swarm represents the feeder dykes of a fossil midocean ridge. Vuagnat & Çogolu (1967) have described a similar dyke swarm in the Kizil Dagh ophiolite complex, Hatay province, Turkey.

Comparison with alpine-type peridotite-gabbro complexes

Alpine-type peridotites are fault-bounded ultramafic bodies typically located in folded eugeosynclinal sediments in orogenic belts. Thayer (1967) emphasized that the alpine-type peridotites are commonly associated with Mg-rich gabbro, and these associations are better described as alpine-type peridotite-gabbro complexes. He cited the Canyon Mountain complex, Oregon, as a typical example, and discussed such other examples as the Bay of Islands complex, Newfoundland (Smith, 1958), and the Zambales complex, Luzon (Rossman et al., 1959).

Cumulus textures are preserved in some alpine-type complexes; for instance, Thayer (1969b) cited cumulus chromite, plagioclase, and olivine in a podiform chromite deposit in Camaguey district, Cuba (Flint et al., 1948). Also, some of the layers in gabbro at Canyon Mountain look like cumulus layers (Thayer, 1963, fig. 2) but any diagnostic microfabric has probably been obscured by recrystallization, for, from a brief field and petrographic study, I believe the Canyon Mountain complex to have been metamorphosed.

The plutonic part of the Papuan Ultramafic Belt is a typical alpine-type peridotite-gabbro complex in most respects, although the gabbros are less metamorphosed than, for example, the Canyon Mountain complex. Compositions of Papuan and Canyon Mountain (Thayer & Himmelberg, 1968) gabbros are very similar.

Alpine-type complexes and ophiolites

Thayer (1967) noted that the alpine-type peridotite-gabbro complexes are similar to the ultramafic and gabbroic parts of the Mediterranean ophiolite sequences, and suggested that all have a common origin. He proposed (Thayer, 1963, 1969b) that stratified peridotite-gabbro complexes form in the mantle by crystal-settling from fluid magma, and that these complexes are then emplaced in the crust as a stiff plastic crystal mush. However, such a process would not explain the association of basaltic lavas with the plutonic rocks in the ophiolite sequences. Thayer (1967, pp. 231-2; 1969b) argued that the basalts are not an integral part of the ophiolite sequence, but the bulk of field evidence and chemical data (e.g. Fig. 7) suggests that, in fact, they are.

If the alpine-type peridotite-gabbro complexes and the plutonic part of the ophiolite sequences are equivalent, as seems likely, it might be more logical to suggest that the alpine-type peridotite-gabbro complexes are ophiolite sequences in which the overlying volcanics have been either unrecognized, or removed by faulting or erosion.

Origin of ophiolites

Field observations have led various workers to propose that the volcanic, gabbroic, and ultramafic parts of any one ophiolite sequence are cogenetic, and this interpretation has been supported by chemical diagrams which show apparent fractionation trends with ultramafic rocks as one end member and basaltic volcanics as the other (Fig. 7, and Nicolas, 1966, pp. 127-47). These workers generally subscribe to what is termed the plutovolcanic hypothesis for the origin of ophiolites (Routhier, 1946; Dubertret, 1953; Brunn, 1956, 1960; Gansser, 1959; Maxwell & Azzarolli, 1962; Aubouin, 1965).

According to the plutovolcanic hypothesis, a massive lava flow up to 5 km thick is extruded upon the ocean floor. The outer skin is chilled to form pillow lavas, massive basalt and, inwards, dolerite. The inner part of the extrusion remains fluid and proceeds to differentiate by crystal settling to produce an ultramafic base and a gabbroic top capped by small irregular bodies of diorite. The end-product is a differentiated pluton several kilometres thick, encased in a shell of dolerite and basalt.

Critics of the plutovolcanic hypothesis (e.g. Vuagnat, 1963; Kaaden, 1964; Thayer, 1967) have pointed out five difficulties:

1. The typical ophiolite sequence is bottom-heavy, with too much ultramafic differentiate for the volume of the associated gabbro.

2. If the parent magma had been basaltic there should be a fraction enriched in Fe, Na, and K to provide a chemical counterweight for the Ca-Mg-rich ultramafic fraction.

3. Extrusion of a tongue of lava of such great volume at one time is unlikely.

4. Field evidence for the complete envelope of basalt and dolerite enclosing the plutonic rocks has not been established (e.g. Bortoletti et al., 1969).

5. The gabbro-peridotite contact instead of being transitional is commonly intrusive, gabbro into peridotite (Steinmann, 1927; Vuagnat & Cogolu, 1967; Maxwell, 1969).

These problems can be resolved if the plutovolcanic hypothesis is modified along the lines of the model suggested for evolution of the Papuan Ultramafic Belt:

1. The excess of ultramafic rocks includes a faulted segment of pre-existing mantle.

2. The parent magma is richer in Ca and Mg than are the basalts; the basalts represent residual liquids from crystal fractionation of the parent.

3. The complex developed in a tensional zone within the crust rather than as a thick flow on top of the crust, and was fed by successive pulses of parent magma.

4. Pillow lavas, massive basalt, and dolerite formed only at the top of the complex.

5. Gabbro intruded the pre-existing mantle, and later pulses of gabbro intruded the earlier-formed ultramafic cumulates.

Field observations in the Vourinos complex (Moores et al., 1966) support the suggestion that some or all ophiolites include a portion of ultramafics which has a separate origin from the rest of the complex, perhaps as a piece of pre-existing mantle. Whereas the gabbroic rocks of the Vourinos complex have simple layering 'probably of magmatic origin', the associated ultramafics have much more complex structure 'probably formed by nearly solid penetrative flows'.

Ophiolite chemical trends which are continuous from ultramafics through gabbro to basalt have been interpreted to show that the ultramafics are related to the other rocks (e.g. Nicolas, 1966, pp. 127-47; Bottcher, 1969). While this is probably true where the ultramafics are cumulates, it is probably not true for the great masses of apparently noncumulus dunite and harzburgite. The chemical diagrams may be misleading because dunite and harzburgite, by virtue of their high Mg and low alkali content, will plot as extreme differentiates on most graphical representations of chemical data. Neither Nicolas' review of Vourinos, Kizil Dagh, and Lanzo chemistry, nor Bottcher's plot of Troodos mineral compositions indicates any differentiation within the ultramafic rocks themselves. The lack of a differentiation trend within the ultramafic zone accords better with origin as a piece of mantle, than as a crystal cumulate.

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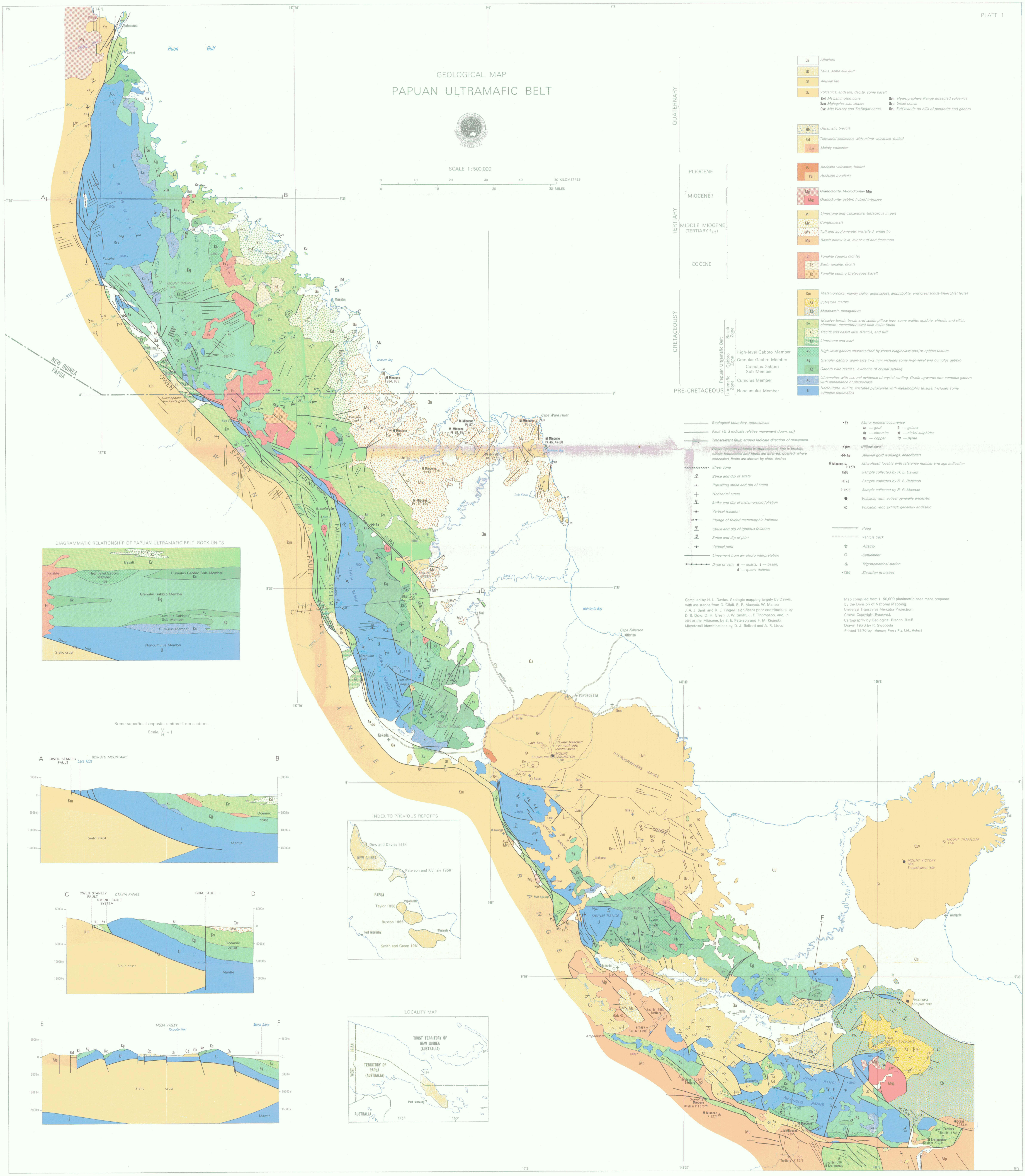
PLATE 1
GEOLOGICAL MAP
PAPUAN ULTRAMAFIC BELT
SCALE 1 : 500,000
0 10 20 30 40 50 KILOMETRES
0 10 20 30 MILES
QUATERNARY
Qa Alluvium
Qt Talus, some alluvium
Qf Alluvial fan
Qv Volcanics: andesite, dacite, some basalt
Qvl Mt Lamington cone
Qvm Malagalas ash, slopes
Qvv Mts Victory and Trafalgar cones
Qvh Hydrographers Range dissected volcanics
Qvc Small cones
Qvu Tuff mantle on hills of peridotite and gabbro
Qb Ultramafic breccia
Qd Terrestrial sediments with minor volcanics, folded
Qdv Mainly volcanics
TERTIARY
PLIOCENE
Pv Andesite volcanics, folded
Pa Andesite porphyry
MIOCENE ?
Mg Granodiorite. Microdiorite-Mg1
Mgg Granodiorite-gabbro hybrid intrusive
MIDDLE MIOCENE (TERTIARY f1–2)
Ml Limestone and calcarenite, tuffaceous in part
Mc Conglomerate
Mv Tuff and agglomerate, waterlaid, andesitic
Mp Basalt pillow lava, minor tuff and limestone
EOCENE
Et Tonalite (quartz diorite)
Ed Basic tonalite, diorite
Eb Tonalite cutting Cretaceous basalt
CRETACEOUS?
Km Metamorphics, mainly sialic; greenschist, amphibolite, and greenschist-blueschist facies
Ks Schistose marble
Kb Metabasalt, metagabbro
Papuan Ultramafic Belt
Basalt Zone
Kv Massive basalt; basalt and spilite pillow lava; some uralite, epidote, chlorite and silicic alteration; metamorphosed near major faults
Kd Dacite and basalt lava, breccia, and tuff
Kl Limestone and marl
Gabbro Zone
High-level Gabbro Member
Kh High-level gabbro characterized by zoned plagioclase and/or ophitic texture
Granular Gabbro Member
Kg Granular gabbro, grain-size 1–2 mm; includes some high-level and cumulus gabbro
Cumulus Gabbro Sub-Member
Kc Gabbro with textural evidence of crystal settling
Ultramafic Zone
Cumulus Member
Ku Ultramafics with textural evidence of crystal settling. Grade upwards into cumulus gabbro with appearance of plagioclase
PRE-CRETACEOUS
Noncumulus Member
U Harzburgite, dunite, enstatite pyroxenite with metamorphic texture. Includes some cumulus ultramafics
Geological boundary, approximate
Fault (D U indicate relative movement down, up)
Transcurrent fault, arrows indicate direction of movement
Where location of faults is approximate, line is broken; where boundaries and faults are inferred, queried; where concealed, faults are shown by short dashes
Shear zone
Strike and dip of strata
Prevailing strike and dip of strata
Horizontal strata
Strike and dip of metamorphic foliation
Vertical foliation
Plunge of folded metamorphic foliation
Strike and dip of igneous foliation
Strike and dip of joint
Vertical joint
Lineament from air-photo interpretation
Dyke or vein: q — quartz, b — basalt, d — quartz dolerite
• Py Minor mineral occurrence:
Au — gold
Cr — chromite
Cu — copper
G — galena
Ni — nickel sulphides
Py — pyrite
• pw Pillow lava
Au Alluvial gold workings, abandoned
M Miocene P 1274 Microfossil locality with reference number and age indication
1593 Sample collected by H. L. Davies
Pk 78 Sample collected by S. E. Paterson
P 1276 Sample collected by R. P. Macnab
Volcanic vent, active; generally andesitic
Volcanic vent, extinct; generally andesitic
Road
Vehicle track
Airstrip
Settlement
Trigonometrical station
• 1350 Elevation in metres
Compiled by H. L. Davies. Geologic mapping largely by Davies, with assistance from G. Cifali, R. P. Macnab, W. Manser, J. A. J. Smit and R. J. Tingey; significant prior contributions by D. B. Dow, D. H. Green, J. W. Smith, J. E. Thompson, and, in part of the Miocene, by S. E. Paterson and F. M. Kicinski. Microfossil identifications by D. J. Belford and A. R. Lloyd.
Map compiled from 1 : 50,000 planimetric base maps prepared by the Division of National Mapping.
Universal Transverse Mercator Projection.
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Cartography by Geological Branch BMR
Drawn 1970 by R. Swoboda
Printed 1970 by Mercury Press Pty. Ltd., Hobart
Huon Gulf
OWEN STANLEY RANGE
TIMENO FAULT SYSTEM
GIRA FAULT
NEW GUINEA
PAPUA
POPONDETTA
HYDROGRAPHERS RANGE
MOUNT VICTORY
MOUNT TRAFALGAR
DIAGRAMMATIC RELATIONSHIP OF PAPUAN ULTRAMAFIC BELT ROCK UNITS
Basalt Kv
Tonalite
High-level Gabbro Member Kh
Cumulus Gabbro Sub-Member Kc
Granular Gabbro Member Kg
Cumulus Gabbro Sub-Member Kc
Cumulus Member Ku
Noncumulus Member U
Sialic crust
Some superficial deposits omitted from sections
A B
OWEN STANLEY FAULT
BOWUTU MOUNTAINS
Oceanic crust
Sialic crust
Mantle
C D
OWEN STANLEY FAULT
TIMENO FAULT SYSTEM
GIRA FAULT
Oceanic crust
Sialic crust
Mantle
E F
MUSA VALLEY
Musa River
Sialic crust
Mantle
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Port Moresby
LOCALITY MAP
TRUST TERRITORY OF NEW GUINEA (AUSTRALIA)
TERRITORY OF PAPUA (AUSTRALIA)
WEST IRIAN
AUSTRALIA
Port Moresby