R i c h a r d   I r v i n g   W a l c o t t

S T R U C T U R E   A N D   P E T R O L O G Y

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R E D   H I L L   C O M P L E X,   N E L S O N

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CONTENTS

PART I. STRATIGRAPHY AND GENERAL GEOLOGY

PART II. PETROGRAPHY

SERPENTINITES, TECTONIC INCLUSIONS

PART III STRUCTURE

PART IV. PETROGENESIS.

ILLUSTRATIONS

ABSTRACT

The Red Hill Complex is an essentially concordant ultramafic body enclosed in Upper Paleozoic flysch facies sediments which include Pelorus Group (oldest), Lee River Group and Maitai Group. The Pelorus Group contains rare submarine lavas and is largely derived from spilitic volcanics. The Lee River Group consists of spilitic pillow lavas, volcanic breccias and spilitic basalts and dolerites. The Maitai Group consists of limestone, sandstone and argillite; an extensive conglomerate lens in the argillites is largely composed of andesitic pebbles.

The Red Hill Complex is a 12,000 ft. thick lens and is part of a sheet of peridotites which may extend 40 miles northward to Dun Mountain. The Complex is divided into a 3000 ft thick Basal Zone of massive harzburgite and a 9000 ft thick Upper Zone of layered harzburgite and dunite with minor variants, feldspathic-peridotite, eucrite, lherzolite, wehrlite and pyroxenite. The bulk composition of both zones is approximately the same but the Upper Zone contains about 0.2 per cent feldspar not present in the Basal Zone. There is no significant regional change in mineral chemistry throughout the Complex and the average composition is about; olivine Fo_{91}, 70 per cent; orthopyroxene En_{88}, 22 per cent; clinopyroxene, 5 per cent; feldspar An_{96}, less than 0.2 per cent; spinel 2 per cent.

Layering and foliation are common in the top of the Upper Zone. Layering is of at least two generations of which at least one is of metamorphic origin. Metamorphic layering was formed by metasomatic replacement probably along subhorizontal shear planes during intrusion of the ultramafic sheet. Pyroxene pegmatites formed after flow ceased. The diversity of rock types in the top of the Upper Zone is considered by the writer to have been caused by metamorphic differentiation of parent material the same composition as the Basal Zone.

The preferred orientation of olivine in lineated, foliated, laminated and layered rocks has the same pattern suggesting a close genetic relationship between those structures. Evidence strongly supports a tectonic origin for the preferred orientation.

Rocks in the Upper Zone are xenomorphic-granular in texture and those in the Basal Zone are typically protoclastic. Xenomorphic-granular textures are derived in part from protoclastic by post-deformational recrystallization.

The ultramafic rocks are cut by a number of dykes composed of hornblende-labradorite, hypersthene-augite-bytownite assemblages or minor variants of these. The dykes were intruded shortly after emplacement of the ultramafic rocks.

The Red Hill Complex is considered to have been emplaced as a sheet at shallow depths which intruded superficial deposits on the ocean floor and was later overlain by volcanics.

INTRODUCTION

GEOLOGICAL OUTLINE, FIELD WORK AND SCOPE OF THESIS

The Red Hill Complex is an Alpine-type peridotite-gabbro complex (Thayer, 1960) situated about 40 miles south of the city of Nelson in the South Island of New Zealand. It is part of a belt of ultramafic rocks, the Nelson Ultramafic Belt, that extends for more than 80 miles from D'Urville Island in the north to the Wairau Fault in the south (fig. 1). The ultramafic rocks occur as small sills and extensive concordant sheets in steeply dipping basic volcanic rocks which are of early Permian age (Waterhouse, 1964). Stratigraphically below and to the east of the volcanic rocks is a thick sequence of eugeosynclinal strata which grades eastward into schist. Overlying and to the west of the volcanic and ultramafic rocks are late Permian fossiliferous limestone, sandstone and argillite. To the west again are poorly exposed volcanics probably the same age as those to the east. The whole sequence from schist to the westerly band of volcanics is named here, the Nelson Upper Paleozoic Belt. A belt of similar ultramafic rocks occurring in the south of the South Island is here named the Otago Ultramafic Belt.

This thesis is concerned with the origin of ultramafic rocks and involves a structural and petrological study of the Red Hill Complex. Emphasis is on broad description and general structure and in this respect the study is complementary to detailed mineralogical and

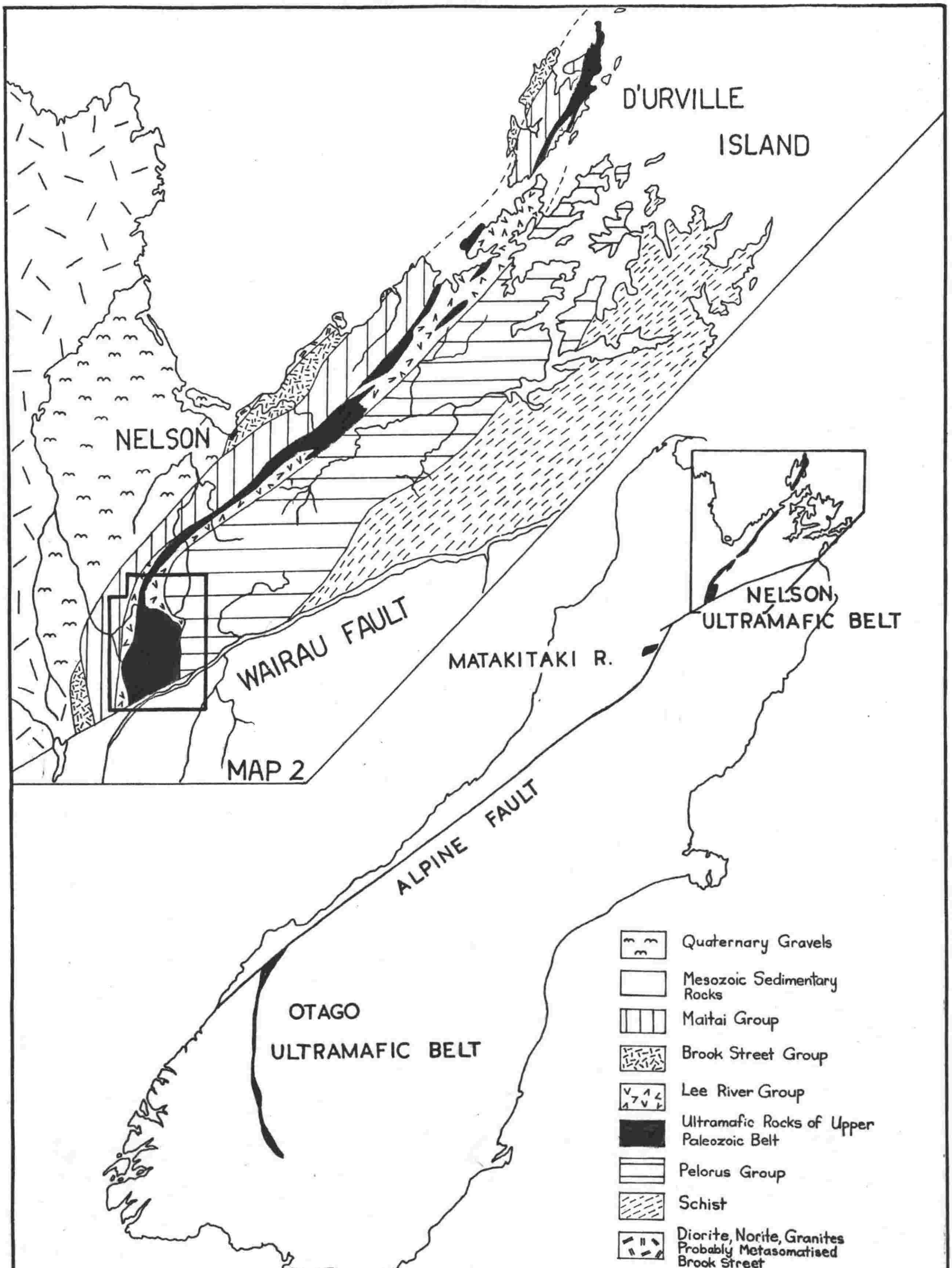

Fig. 1. Map 1. Distribution of New Zealand Permian Ultramafics.

Map 2. Nelson Ultramafic Belt. Inset shows mapped area.

petrological work done by Dr. G. A. Challis on part of the south-western margin of the Complex.

The thesis is presented in four parts. Part 1 covers stratigraphy and general geology of the area mapped. Part II gives the petrography of the mafic dykes, the peridotites and gabbro and the serpentinites and tectonic inclusions. Part III is concerned with structure and presents hypothesis of structural development of the Red Hill Complex. Part IV discusses petrogenesis and the mode and age of emplacement of the Red Hill Complex.

Field work commenced in November 1962 and in the following two years 165 days were spent in the field. Field sheets on a scale of 20 chains to the inch were compiled from aerial photographs. The final geological map (Map 1, in rear pocket) is presented in a scale of 1:50,000 and covers 160 square miles. Supplementary maps include a location map (Map III, rear pocket) showing the positions of petrological specimens and photographs referred to in text.

PREVIOUS WORK

The major contributions to the geology of the Nelson Upper Paleozoic Belt are few. Hochstetter (Fleming, 1959) first mapped and described the Ultramafic Belt and named the olivine rock, dunite, from Dun Mountain. McKay (1878) mapped the rocks south of Dun Mountain and drew attention to the stratigraphic restriction of the ultramafic rocks, pointing out that they are overlain throughout their length by a limestone formation. Dun Mountain was mapped on a scale of an inch to the mile by Bell, Clarke and Marshall (1911) who also described and named the hydrogrossular-diopside rock, rodingite. A small part of the Red Hill Complex was mapped by Henderson, Macpherson and Grange (1959) and the writer has drawn from their field sheets in determining the structure of the north-western part of the mapped area (in the valley of the right branch of the Wairoa River). No further regional maps of the Nelson Belt were published until recent years but several geologists, in particular H. W. Wellman, were active in determining the stratigraphy of the Upper Paleozoic rocks. Waterhouse (1964) in discussing the Permian Stratigraphy of New Zealand described many features of the Nelson Upper Paleozoic Belt and the writer has largely followed his terminology. Petrological work on the ultramafic rocks was given by Lauder (1965) who described the geology and petrology of Dun Mountain, and Challis (1965a, 1965b) who did detailed mineralogical and petrological work on rocks from part of the south western margin of the Red Hill Complex. A gravity survey covering in part the southern extremity of the Red Hill Complex was made by Malahoff (1962).

Part of the Otago Ultramafic Belt has been studied by Grindley (1956) who gave an account of earlier work in the area. Grindley's observations have been largely supported in the present study.

Recently, several geological maps on a scale of 4 miles to an inch of the Nelson and Otago Ultramafic Belts were published by the Department of Scientific and Industrial Research (Wood, 1962; Lensen, 1962; Beck, 1964).

ACKNOWLEDGMENTS

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The writer records his appreciation to the staff of the Geology Department, Victoria University of Wellington for valuable discussion and criticism. In particular he thanks Professors J. Bradley, R. H. Clark, and H. W. Wellmann, Dr. E. Ghent and Mr R. Lauder. Mr E. Hardy gave assistance in draughting the geological map and some diagrams. Dr. R. N. Brothers of Auckland University gave some detailed and stimulating criticism for which the writer records his appreciation. The following members of the Department of Scientific and Industrial Research gave the writer assistance over particular aspects of the work, Dr. B. Waterhouse, Dr. G. A. Challis of Geological Survey and Dr. T. Hatherton of Geophysics Division. The Director, Geological Survey kindly made available to the writer the field sheets of Henderson et al (1959).

The thesis would not have been completed without the considerable and able assistance of the writer's wife, Gennis R. Walcott.

STRUCTURE AND PETROLOGY

OF THE

RED HILL COMPLEX, NELSON

PART I

STRATIGRAPHY AND GENERAL GEOLOGY

STRATIGRAPHY

GENERAL GEOLOGY

The oldest sedimentary rocks of the Nelson Upper Paleozoic Belt are thick unfossiliferous strata and rare submarine lavas, which are named the Pelorus Group (Waterhouse, 1964) and are probably Carboniferous in age. They are structurally complex and little detailed mapping of these rocks has been previously attempted but in the Red Hill area they dip steeply and young westwards. Eastwards they grade into Marlborough Schist, probably derived from sediments of similar age and lithology.

Overlying the Pelorus Group are several thousand feet of strata comprising pillow lavas, thick sheets of spilitic basalt and albite dolerite and minor impure sandstone and argillites that are named the Lee River Group. The Lee River Group is in turn overlain by the Maitai Group which consists of a basal limestone followed by sandstone, thick laminated argillite and further sandstone and limestone.

The Maitai Group is folded into an asymmetric syncline with a steeply dipping eastern limb that is concordant with underlying Lee River and Pelorus Groups. The western limb is faulted against the Brook Street Volcanics, regarded by Waterhouse (1964) to be, in part, contemporaneous with the Lee River Group. Broadly, therefore, the structure of the Nelson Upper Paleozoic Belt is synclinal with the eastern limb consisting of Marlborough Schist, Pelorus, Lee River and Matai Groups and the western limb, of Brook Street Volcanics and Maitai

Group. The name Nelson Syncline has been given to this structure by Wellman (1956).

Peridotite, serpentinite and minor gabbroic rocks intrude the Lee River Group forming an almost continuous sheet from D'Urville Island to the Wairau Fault (fig. 1, map 2). The ultramafic rocks attain their greatest width of six miles at Red Hill, and thin to less than two hundred yards wide on D'Urville Island. Mostly, however, the sheet is about 1000 yards wide. Ultramafics also occur as small concordant lenses within the Lee River Group and one thin sill has been found within the Brook Street Volcanics on D'Urville Island (Waterhouse, 1964).

The geological map of the Red Hill Complex covers part of the Upper Paleozoic sequence on the eastern limb of the Nelson Syncline. Marlborough Schist lies about a mile to the east of the mapped area and the lowest rocks of the Upper Paleozoic sequence mapped are about two thirds from the top of the Pelorus Group. The axis of the Nelson Syncline lies west of the mapped area and the highest rocks in the sequence that have been mapped are the laminated argillites of the Maitai Group.

Attention is concentrated on the Upper Paleozoic rocks but the Mesozoic rocks south of the Wairau Fault and superficial Quaternary deposits are briefly mentioned.

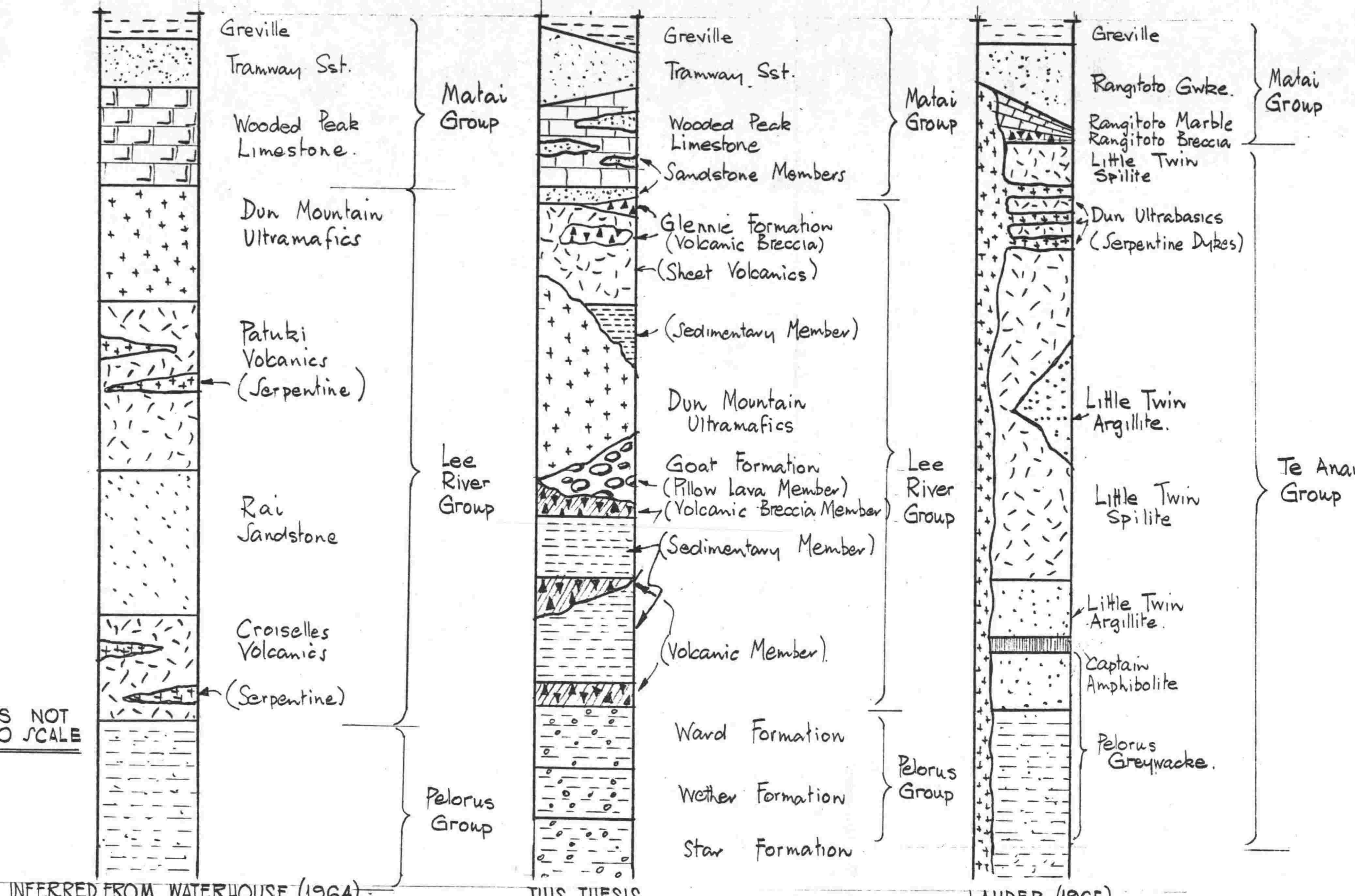
A summary of the stratigraphy of the Red Hill area is given in Table I. The age of the rocks is taken from Waterhouse, 1964. The nature of the contacts between the formations is given as 'conformable' where the contact is parallel to bedding planes and there is no evidence of an hiatus; 'disconformable' where the two formations have parallel bedding planes but the contact is marked by a possible erosion break; 'transgressive facies change' where the contact is transgressive to bedding planes, or 'fault' in which the only contact observed between the formations is occupied by a fault.

TABLE I. UPPER PALEOZOIC STRATIGRAPHY OF RED HILL AREA

AGE	GROUP	FORMATION	NATURE OF CONTACT	LITHOLOGY	THICKNESS
Tatarian		Waiua	Conformable	Red and green laminated argillite	
		Greville	Transgressive facies change	Grey and dark grey laminated argillite with sandstone and igneous conglomerate lenses	5000'
	Maitai	Tramway Sst.	Transgressive facies change	Sandstone, fossiliferous siltstone and minor argillite	500'-1300'
Kazanian		Wooded Peak Lst.	Disconformable	Limestone and minor sandstone	500'-1400'
Kungurian Artinskian	Lee River	Glennie	Fault	Sheet volcanics composed of spilitic basalt and dolerite, sandstone, siltstone and argillite	c. 5000'
Sakmarian		Goat	Fault	Pillow lavas, volcanic breccia, sandstone, siltstone and argillite	c. 6000'
Carboniferous		Ward	Conformable	Volcanic derived, impure, green grey sandstone, siltstone and argillite	c. 4000'
	Pelorus	Wether	Conformable	Volcanic derived, impure, red and grey argillite, flecked red and green sandstone, minor green sandstone. Rare lavas	200'-4800'
		Star		Volcanic derived, impure, grey, blue and green sandstone, grey argillite	c. 2000'

FIGURE. 1a.

IDEALISED COLUMNS ILLUSTRATING CORRELATIONS OF LITHOLOGICAL UNITS IN WATERHOUSE (1964), LAUDER (1965) AND THIS THESIS.



NOTE:
COLUMNS NOT DRAWN TO SCALE

Fig. 1a illustrates correlations with previously proposed subdivisions of the Nelson Upper Paleozoic Belt. The most complete discussion of the stratigraphy of the Upper Paleozoic rocks is given by Waterhouse (1964). His broad subdivisions are followed here; the only major differences are the formations recognised in the Lee River Group and discussed on page 31 - 33. Lauder (1965) in his study of the geology of Dun Mountain includes most of the volcanic rocks and closely associated clastic sedimentary rocks as a formation (the Little Twin) in the 'Te Anau Group', but includes a volcanic breccia which occurs as a narrow discontinuous band at the base of the Rangitoto Marble in the Maitai Group. Similar breccias are found in the Red Hill Area (p. 42) at the base of the equivalent Wooded Peak Limestone. Identical breccias have been found entirely enclosed within the Glennie Formation. For this reason the breccia is not regarded as a basal conglomerate of the Maitai Group, in the sense of Lauder (1965, p.6) but as an integral part of the Glennie Formation of the Lee River Group.

P E L O R U S   G R O U P

The Pelorus Group, defined by Waterhouse (1964) as the strata between Marlborough Schist and the Lee River Group consists of Upper Paleozoic volcanic wackes and rare submarine lavas. It is well exposed in the streams and bare ridges of the heavily bush covered and deeply dissected country north and east of Red Hill. To the east, rocks dip steeply and lie in a gently arcuate belt, named the Eastern Belt, but to the north rocks are folded into a broad, asymmetric anticline named the Ben Nevis Anticline, which plunges southwestwards at 30°.

The three formations recognised in the Pelorus Group are defined from the Eastern Belt. A number of sections through the Belt are given in figure 2. Rocks vary in grain size from argillites (silt or finer) to coarse sandstones (frequently with grit bands) and beds range in thickness from less than an inch thick to more than 50 feet thick. The beds are commonly graded. All formations show similar diversity of lithology, the main criterion for the recognition of formations being colour.

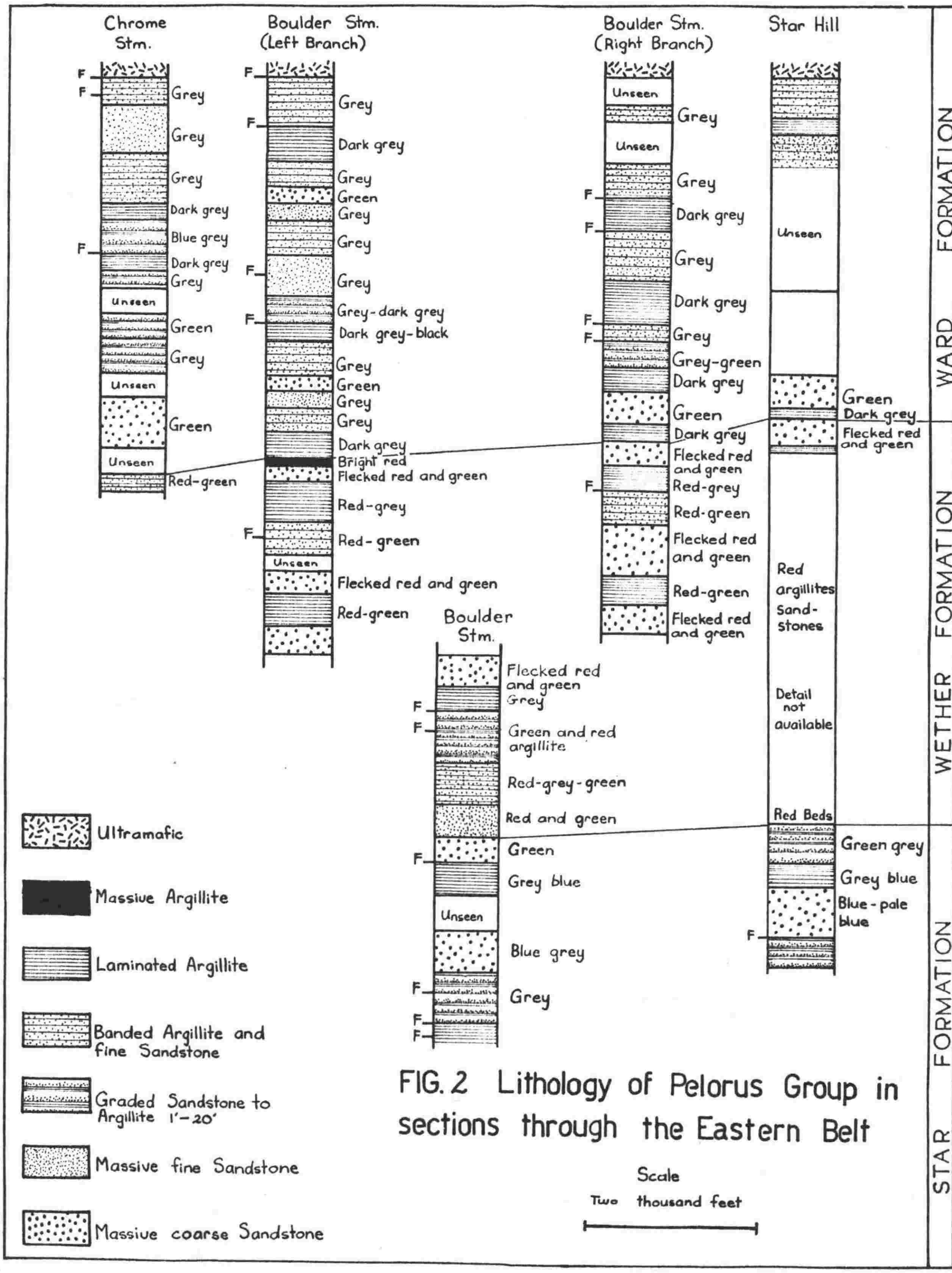


FIG. 2 Lithology of Pelorus Group in sections through the Eastern Belt

STAR FORMATION

The Star Formation is typically exposed on the eastern flanks of Star Hill (Map reference 488853)* and consists of graded beds, laminated argillites, massive green sandstones and a characteristic blue coloured, coarse-grained feldspathic sandstone. The top of the formation is marked by the first appearance of red rocks of the Wether Formation. The stratigraphically lowest rocks of the formation in the Eastern Belt are grey to green graded beds. The thickness of the formation is at least 2000' but complex structure makes it unlikely that this is a true stratigraphic thickness.

WETHER FORMATION

Wether Formation is named from the low hill at 469819 but the type section is in lower and right branch of Boulder Stream (fig. 2). The formation consists of red coloured argillites and sandstones interbedded with grey and green argillites and impure sandstones. Very rare submarine lavas are associated with red argillites. The top and base of the formation ~~is~~ are marked by ~~a~~ transitions into the adjacent formations which do not contain red rocks. The thickness of the formation in Boulder Stream is 4800 feet but faulting and isoclinal folding are common and it is unlikely that this is a true stratigraphic thickness.

Map references given below as six figure numbers without the words 'map reference' refer to the Geological Map of the Red Hill Complex.

WARD FORMATION

The formation is named from Ward Pass Stream but the type section is taken as the right branch of Boulder Stream (fig. 2). The formation consists of green and grey sandstone and laminated and banded argillites. The lower part of the formation is characteristically marked by a coarse green sandstone overlying dark grey laminated argillites which pass downward into the red rocks of the Wether Formation. The top of the formation is everywhere faulted. The stratigraphically highest rocks are banded grey argillites and fine-grained sandstones. The rocks generally become finer-grained towards the top of the formation. The thickness in Boulder Stream is 4500'. As indicated for the other formations this is unlikely to be a true stratigraphic thickness.

DISTRIBUTION OF THE FORMATIONS

Eastern Belt. A section of Pelorus Group rocks described by Waterhouse (1964) from the Lee River, 15 miles north of Chrome Stream is very similar to sections east of Red Hill. In the Lee River, red rocks outcropping over a distance of two and a half miles are flanked on the west by blue and green argillite and on the east by blue argillite and minor green sandstone. No sections through the whole width of the Eastern Belt were examined between Lee River and Star Hill but it is considered on the following evidence that the Eastern Belt continues northward at least as far as Lee River without major change in strike and only minor increase in width of formations.

(i) The strata exposed in tributary streams of the left branch of the Wairoa River strike uniformly at 020°.

(ii) Red pebbles in these streams indicate the presence of Wether Formation in their headwaters.

(iii) Red rocks, presumably of the Wether Formation, are exposed on the summit of Purple Top, a hill which lies on the trend of Wether Formation at approximately midway between Lee River and Star Hill.

It is therefore likely that the three subdivisions of the Pelorus Group given by Waterhouse (1964, fig. 3) are correlatives of the three formations defined above.

Ben Nevis Anticline. The rocks in the Ben Nevis Anticline are lithologically similar and have the same broad sequence as those in the Eastern Belt. Accordingly, red coloured rocks are correlated with the Wether Formation, and the rocks stratigraphically above and below are identified as Ward and Star Formations respectively.

The Wether Formation varies considerably in thickness; on Ben Nevis Ridge it is about 200 feet thick, in the Wairoa River section at least 1000 feet and probably near 2000 feet thick, in the type section 4000 feet thick, and in the Lee River it outcrops over a width of $2\frac{1}{2}$ miles (Waterhouse, 1964). Much of the thickening is probably facies controlled because there is a continuous section at least 400 feet thick in the Wairoa River; this is at least twice the formation thickness on Ben Nevis Ridge. It is also possible that much of the thickening is due to isoclinal folding and strike faulting (p.216).

The Wether Formation is the only formation of the Pelorus Group with a stratigraphically defined top and bottom and it is not possible to determine whether the Star or Ward Formations show similar changes in thickness, but it is unlikely that thickening applies only to the Wether Formation.

CONTENT AND PETROGRAPHY OF THE PELORUS GROUP

F i e l d   d e s c r i p t i o n. The Pelorus Group are highly indurated, dense rocks of varied colour, and grain size, cut by numerous thin veins of quartz, albite and epidote. The most distinctive and characteristic rocks are very poorly sorted green, grey or blue sandstones containing scattered angular fragments up to a centimeter across. McKay (1878) referred to these rocks as aphanitic breccias, a name which well indicates their poor sorting, general medium grain size and angularity of fragments. In sections a few hundred feet thick, the most characteristic feature is varied thickness of individual beds; massive sandstones are interbedded with finely laminated argillites, thick graded beds and banded sand- and siltstones. The predominant colours are grey and green, but the red rocks of the Wether Formation are most distinctive and blue sandstone and argillites are common.

Grading occurs on the scale of magnitude which varies from beds a fraction of an inch up to many tens of feet thick. Micro-cross-stratification is also commonly present but large scale cross bedding has not been observed.

Penecontemporaneous deformation expressed by intraformational folding, giving highly complex structures on the scale of about 10 to 100 feet is widespread. Sections of strata with parallel strike and dip over a distance of a hundred yards or so are not common, and for this reason measurement of strike and dip of individual bedding planes is of little general structural significance. The strikes and dips shown on the geological map indicate considerable thickness of strata of the orientation shown. Obliteration of bedding by penecontemporaneous deformation

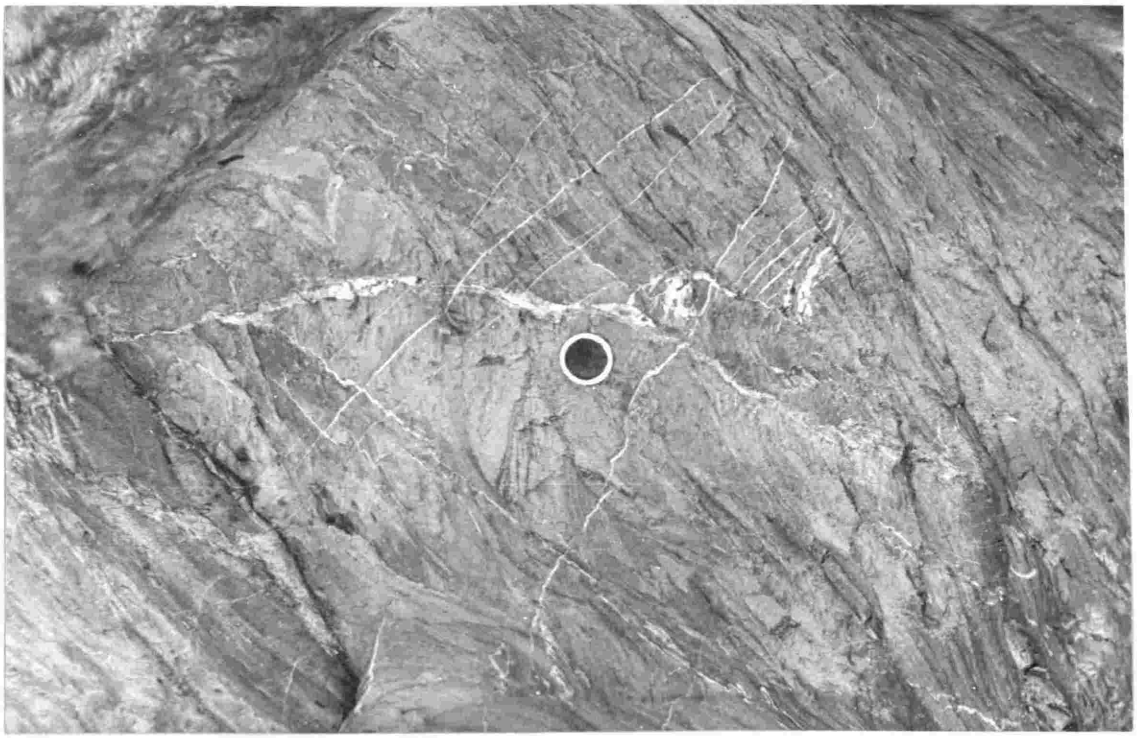


Fig. 3. Laminated argillites, bedding almost completely obliterated by penecontemporaneous deformation. Rock cut by quartz veins.

Fig. 4. Crumpled and sheared argillites of Pelorus Group.

is common in the laminated argillites. A typical example is shown in fig. 3. Only traces of the original bedding ~~is~~ are preserved.

Post-consolidation deformation is expressed by numerous small strike faults, isoclinal folding (amplitude of folds about 300 feet) and bedding plane slip. This last is the most conspicuous - the fine-grained beds in a sequence of thick graded beds or massive sandstones are usually crumpled, sheared and seamed with quartz veins (fig. 4).

General petrography The Pelorus Group are either lithic or feldspathic wackes. Quartz is rare, especially in the coarse-grained rocks which invariably contain less than 5 per cent quartz. Some finer-grained rocks contain as much as 20 per cent quartz. Twinned plagioclase is abundant and contains abundant inclusions probably of the calcsilicates, pumpellyite and epidote. The plagioclase is in all cases optically positive and has an extinction angle of about 18°. Because of highly refringent dust-like inclusions, measurement of refractive index was not always possible but in those in which it was, the refractive index was less than 1.545. The plagioclase is therefore albite, but the inclusions of calc-silicate suggest that the albite was formed by alteration of initially more calcic plagioclase. Lithic fragments are also abundant and are composed either of volcanic or sedimentary rocks. Volcanic fragments are generally spilitic with small acicular laths of plagioclase randomly orientated in an indeterminate matrix. More rarely, coarser-grained volcanic rocks, with a doleritic texture are present. The sedimentary rock fragments are composed of argillite, similar to the finer-grained rock of the Pelorus Group and are probably locally derived. Mono-mineralic fragments, present only in minor amounts, include iron-rich epidote, augite, and rarely strongly pleochroic hornblende. Fine-grained aggregates of chloritic composition are also present.

The matrix of the rocks is recrystallised but the degree of recrystallisation varies widely from rock to rock. In some rocks both matrix and lithic fragments are almost obliterated by recrystallisation. The recrystallisation is shown by small granular aggregates of green pumpellyite with its characteristic strong dispersion and highly birefringent epidote. These form clusters such as those shown in fig. 5.

D e t a i l e d   d e s c r i p t i o n

No. 10924* A coarse-grained grey sandstone from thick-graded beds at 482933. The rock is poorly sorted with subangular fragments up to 2 mm. in size. Recognisable fragments constitute 70 per cent of the rock and are made up of argillite fragments 15 per cent, spilitic fragments 15 per cent, plagioclase 60 per cent, quartz 5 per cent, with accessory chlorite aggregates and augite. The matrix is composed of small clusters of pumpellyite and epidote and unidentifiable, extremely fine-grained material. The rock is cut by veinlets of quartz containing a little pumpellyite.

No. 10925 A coarse sandstone from a massive bed of red and green flecked sandstone at 479920. The original clastic texture is considerably modified by extensive recrystallisation of the matrix and some of the lithic fragments. A third of the fragments are argillite, another third spilitic volcanics and the remainder composed of plagioclase, quartz (less than 5 per cent) and aggregates of chlorite.

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* The numbers refer to specimens in the petrological collection of the Geology Dept., Victoria University of Wellington.

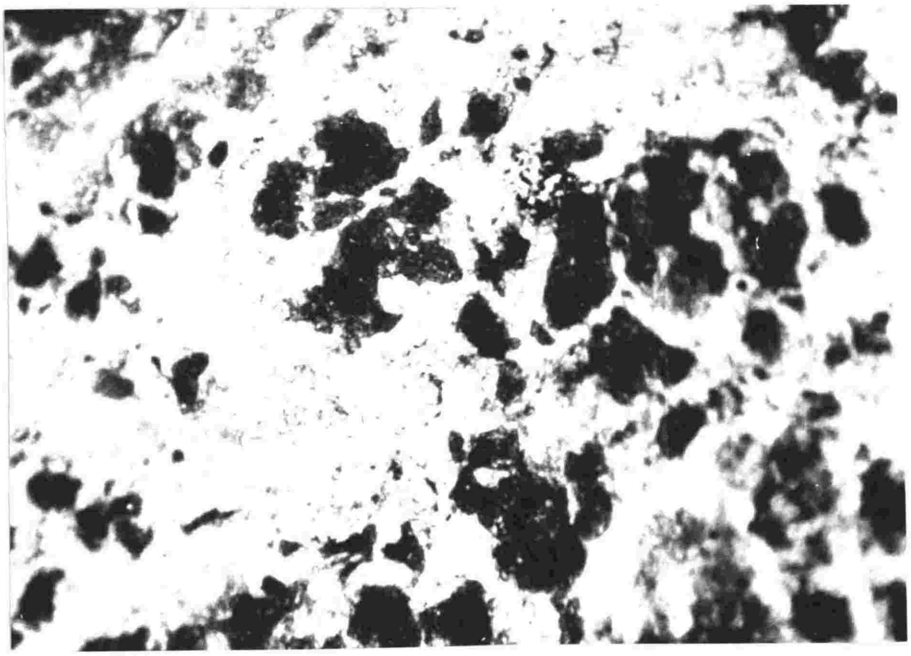

Fig. 5. Clusters of granular epidote and pumpellyite in recrystallised matrix of Pelorus Group impure sandstone spec. no. 10925 (X 150).

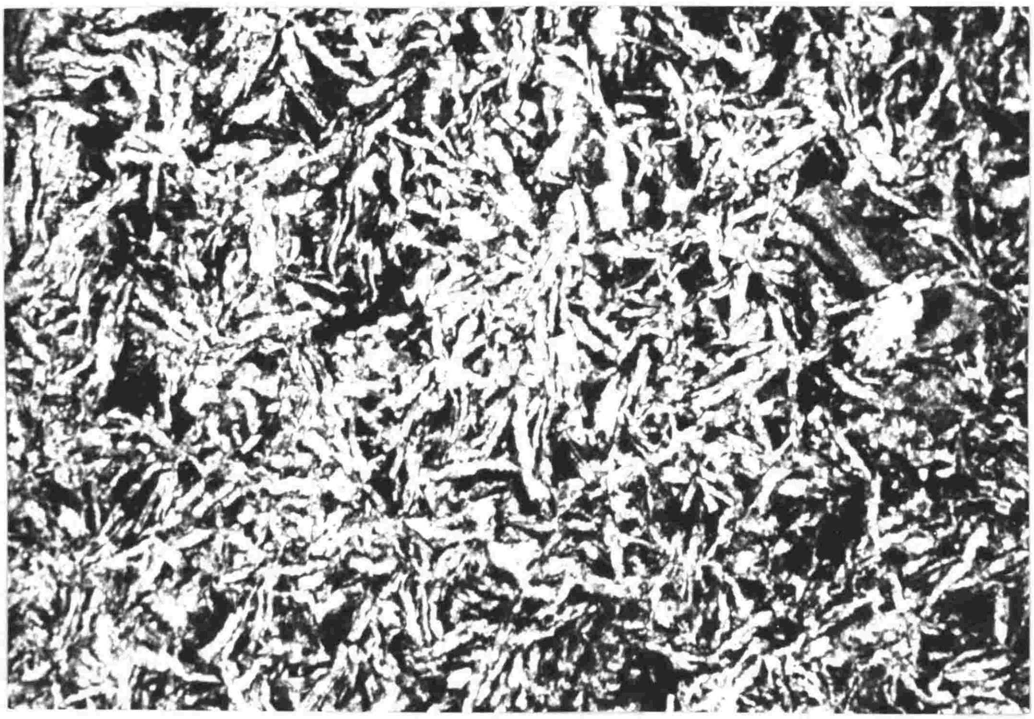

Fig. 6. Photomicrograph of submarine lava from the Pelorus Group. Original feldspar altered to a fine grained mixture of albite and sericite. The matrix is hematite stained and almost opaque. Spec.No.10929 (X 100)

The argillite fragments are packed with very small granules of hematite, and hematite stains much of the recrystallised matrix. Veins of quartz, with epidote, albite and carbonate cut the rock.

No. 10926 A massive coarse-grained green sandstone at 454894. In thin section the poorly sorted clastic texture is shown by aggregates of authigenic epidote and pumpellyite which outline the boundaries of very altered fragments. Few fragments are identifiable but most appear to be of sedimentary origin. It is probable, that volcanic and plagioclase fragments were initially present and are no longer identifiable. Quartz, epidote and albite are accessory, and the authigenic minerals include mica as well as epidote and pumpellyite. The mica occurs in altered argillite fragments and in veins of quartz.

No. 10927 A coarse-grained green sandstone from thick graded beds at 431916. The rock has a moderately well sorted texture of angular fragments up to 2mm. in size but mostly in the range of 0.75 to 1.5 mm. The matrix (grains less than 0.2mm. in size together with interstitial recrystallised material) makes up about 20 per cent of the rock - much less than is usual in the Pelorus Group rocks. Volcanic fragments constitute almost 40 per cent, argillite 25 per cent, and quartz, pyroxene and brown hornblende about 10 per cent of the rock.

Submarine lavas interbedded with laminated red and green argillites of the Wether Formation are found in Boulder Stream at 466830 (No. 10928) and in Chrome Stream at 480795 (No. 10929). In hand specimen the rocks are coloured purple and very fine-grained. They may be mistaken for massive highly indurated red argillite. The lavas occur as small pods not more than 10 feet across and have been only found in the Wether Formation. They are intimately mixed with sedimentary rock, but exposure is not good enough to determine their precise relationship. Probably they are pillow or brecciated pillow lavas. Microscopically, No. 10929 is a fine-grained altered spilitic basalt with randomly orientated elongate laths of altered plagioclase making up about 60 per cent of the rock. These are set in a matrix composed of finely granular hematite along with aggregates of epidote (fig. 6). The altered plagioclase laths are composed of plagioclase (presumably albite) and finely divided muscovite (determined by X-ray diffraction). No. 10928 is similar in composition to No. 10929 but is finer-grained. The feldspar laths are bent and only about 0.5mm. long. Rare 0.5mm. diameter amygdales are filled with iron-rich epidote.

PROVENANCE AND DEPOSITIONAL ENVIRONMENT

General. The Pelorus Group is typical of flysch facies in respect of sedimentary structures, absence of fossils and rarity of volcanic rocks. The large amount of volcanic sedimentary and feldspathic fragments indicate derivation from a rapidly eroding source and the presence of graded bedding suggests that redeposition was probably an important agent of sedimentation. In many respects, therefore, the association is typically eugeosynclinal with rapid sedimentation in a deep water trough far from a stable continental platform (Dunbar and Rodgers, 1957). Most of the fragments are of volcanic origin, and even the fragments of argillite have probably been derived by redeposition of volcanic mud. Thus it appears probable that the Pelorus Group formed in a linear trough closely related to rapidly eroding volcanic material - perhaps a line of volcanoes along the border of the trough.

Massive sandstone. The turbidite hypothesis (Kuenen and Migliorini, 1950) is widely held to explain the origin of graded beds, and Waterhouse (1964) has suggested an ancillary mechanism by which great thicknesses of laminated beds may be formed. The general hypothesis, however, only poorly explains the origin of massive coarse sandstones such as those of the Pelorus Group. Many are 400 feet thick and although some contain conglomerates and grit bands, they commonly show no trace of bedding. These massive sandstones have probably formed by continuous unbroken sedimentation. It is suggested that tectonic activity from time to time raised a part of the margin of the depositional trough within reach of wave erosion or perhaps even above sea level, and for a period erosion was rapid and continuous with the resulting detritus deposited directly over the floor of the basin.

If this is correct, then it may be expected that massive sandstones of the Pelorus Group would be of wide distribution and may be used as marker beds. This is, as yet, unproved.

Red colour of the Wether Formation is due to abundant, finely disseminated granules of hematite distributed throughout the matrix and fragments of the rocks.

It is possible that the origin of the hematite is related to the presence of submarine lavas. The only submarine lavas found in the Pelorus Group are restricted to the Wether Formation (red rocks) and Reed (1957) in describing the red argillites of the Mesozoic rocks of Wellington noted that "the evidence... indicates that submarine extrusion of lavas is closely connected with the formation of red and green argillites".

The relationship of red rocks and submarine lavas is however probably indirect. The very small amount of submarine lavas in the Wether Formation makes it highly unlikely that they could have caused the development of red argillites. Rather, it is considered that both red rocks and lavas are the result of a more fundamental cause. Such a cause would be vigorous vulcanism, which, acting over a long period formed a vast quantity of airborne volcanic ash. The hematite developed from this by oxidation and settled to form the red rocks. During the same period some magma rolled down the slope of the depositional trough to form the submarine lavas.

AGE OF THE PELORUS GROUP

No fossils are known from the Pelorus Group and the only evidence of the age of the rocks is indirect. In the Red Hill area the Pelorus Group rocks dip parallel to, and young towards, the Lee River Group. Although the contact between the two Groups is marked by a fault it is most probable that the Pelorus underlies and is hence older than the Lee River Group. The Lee River Group is on fossil evidence, assigned by Waterhouse (1964) to the early Permian and therefore the Pelorus Group rocks are probably Carboniferous.

L E E R I V E R G R O U P

The Lee River Group is defined by Waterhouse (1964) as the "sedimentary, volcanic and plutonic rocks exposed between the younger Maitai Group ... and the Pelorus Group." (p. 23) In the Red Hill area, apart from the rare submarine lavas in the Wether Formation, volcanic rocks are confined to the western flank of the Ultramafic Complex near Mt. Ellis and near the northeastern and southeastern corner of the Complex (see geological map). These together with closely associated sedimentary rocks are referred to as the Lee River Group.

Stratigraphic sections through the Lee River Group are given in fig. 7. Sections, numbers 1 to 4 are exposed on the south western contact of the Complex, and are shown on figure 20, page 82 ; sections 5 to 8 are exposed in Porters, Ellis and Diorite Streams and the Motueka River respectively. The most complete section, number 9, is exposed on the ridge that passes over Mts. Glennie and Ellis between 380940 and 430920.

Waterhouse, 1964 recognised three formations in the Lee River Group; the lowest (the Croiselles Volcanics) is separated from the highest (the Patuki Volcanics) by a 2000 feet thick sedimentary formation (the Rai Sandstone). The same overall sequence is shown in

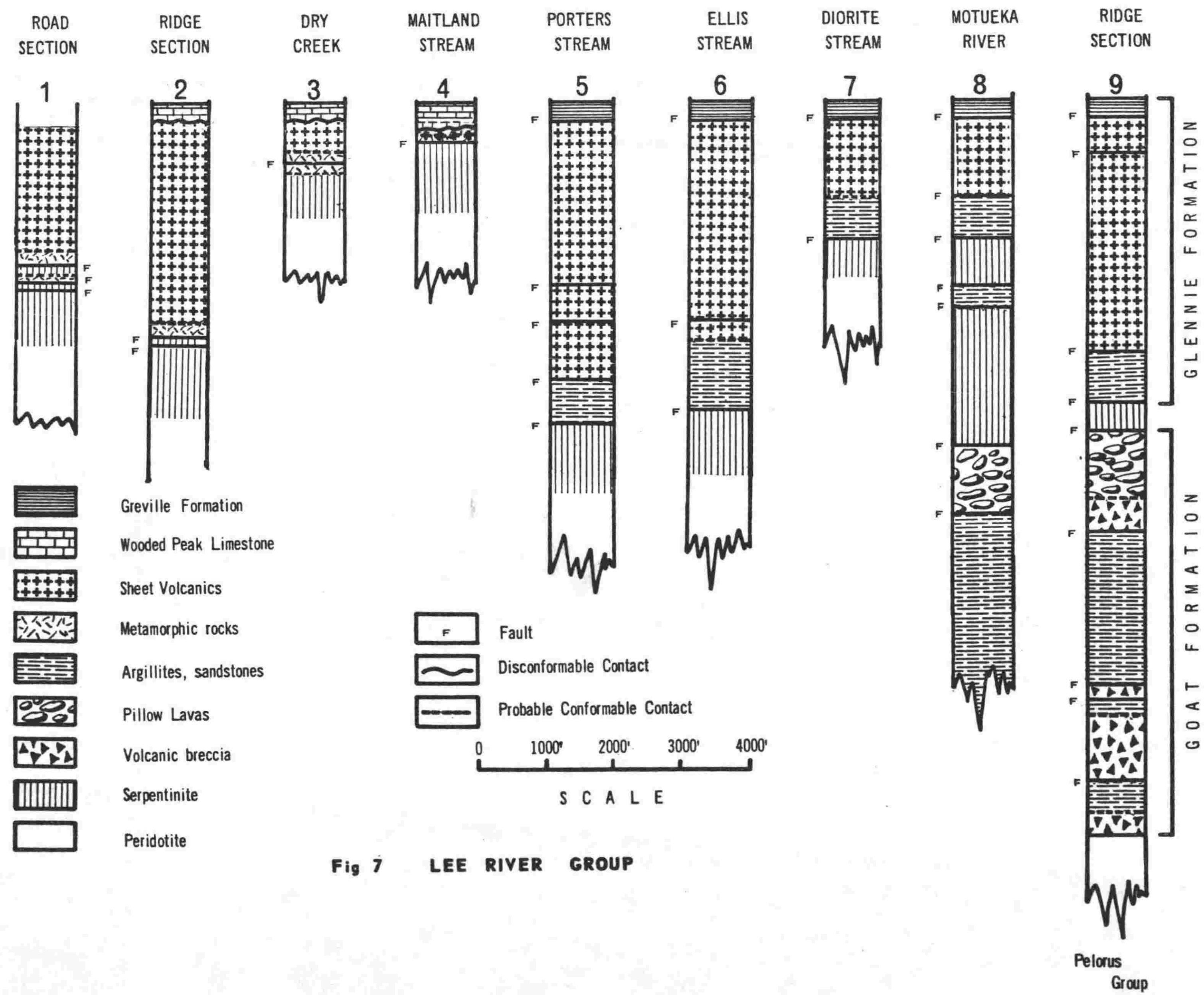

Fig 7 LEE RIVER GROUP

the Red Hill area (section No. 9) and probable correlations are indicated in fig. 1a. There is no lithological difference between the Croiselles and Patuki Volcanics, but in the Red Hill Area the highest volcanics are sheet volcanics, readily distinguished in the field from pillow lavas and volcanic breccias lower in the Group. Accordingly, two formations are recognised. Associated sedimentary rocks are included as members of one or other of those formations. The two-fold division reflects what are thought to be fundamental differences in the environment of deposition of the volcanics, not recognised in Waterhouse's three fold division.

Waterhouse also included concordant lenses of serpentinites as members of the Patuki and Croiselles Formations but there is no assurance, on present evidence, that the serpentinites are contemporaneous with adjacent strata and therefore it is preferred to separate all ultramafic rocks into one unit - the Dun Mountain Ultramafics described later.

GOAT FORMATION

The Goat Formation, defined as the sedimentary and volcanic rocks exposed on the ridge section (fig. 7, No. 9) between 400935 and 424920, is named from Goat Saddle at 404932. It characteristically contains pillow lavas and derived volcanic breccias and silt and sandstones. Sheet volcanics are very rare. The formation is probably equivalent to the whole of the Rai, and part of the Patuki and Croiselles Formations of Waterhouse (1964). The sedimentary and volcanic members of the formation are shown separately on the geological map.

The top of the formation is marked by a thick band of pillow lavas and volcanic breccias. These are in fault contact with dark grey argillites and green sandstones of the overlying Glennie Formation. At the base of the formation is a 100 feet thick band of volcanic breccia in fault contact with the Pelorus Group. The formation is about 6000 feet thick, but the thickness is not accurately known because of faulting and penecontemporaneous slumping.

Content

Volcanic member. - Three types of volcanic rocks can be distinguished in the field; sheet volcanics, pillow lavas and volcanic breccia.

Sheet volcanics is the name used here to describe tabular bodies of volcanic or hypabyssal rock. Only one sheet, about 100 feet thick has been identified as a tabular body in the Goat Formation

Fig. 8. Pillow lavas in the Gbat Formation.

(at 399913) and elsewhere rocks with a similar texture have indeterminate form. In contrast to the overlying Glennie Formation, sheet volcanics are very rare.

Well formed pillow lavas (fig. 8) are abundant in the stratigraphically higher parts of the Goat Formation and are well exposed at 398914. Most outcrops of pillow lavas show some evidence of penecontemporaneous movement, such as zones of brecciation and highly contorted interlayered sedimentary rocks. With increase in degree of deformation, the pillow lavas grade into volcanic breccias.

Volcanic breccia is the most common rock type. Outcrops, typically, show an ill defined but unmistakable bedding with individual beds up to 50 feet thick. These are composed of poorly sorted fragments of fine-grained volcanics and are generally red in colour. The fragments range in size from sand up to boulders several feet in diameter, and fragments of pillow lavas and boulders comprising a number of cemented pillows are common. Less typically, the volcanic breccia may be composed of fragments no larger than grit, but the similarity in composition is shown in thin section. With further decrease in particle size, the volcanic breccias grade into sandstones.

The general characteristics typical of volcanics of the Goat Formation are well summarised by Twenhofel (1932 p. 240):

> "All gradations exist between the somewhat broken and slightly disturbed fragments of lava and the farther-transported and better sorted lava derivatives which enter into a true sedimentary rock. The materials in

"all these cases are so similar and the rocks so intermingled structurally, that it is a matter of great difficulty at times to distinguish between the lavas and the volcanic breccia or tuff, especially since the lava not infrequently in its flow and pillow structures simulates the breccia structure in a most confusing fashion."

Sedimentary member. The sedimentary member is composed largely of massive or graded, coarse-grained sandstones, but massive black argillites, and laminated andbanded grey or grey-green rocks are common. The member is lithologically similar to the Pelorus Group.

Exposed in the head waters of the left branch of the Motueka River is a thick sequence of graded beds and massive sandstone in the middle of the formation. **This is complexly** intraformationally folded. The beds are isoclinally folded, as shown by frequent reversals in younging direction obtained from grading and micro-cross bedding; yet the folded rocks are confined to a belt which is only broadly flexured. The intraformational folding was presumably caused by slumping of unconsolidated sediments.

Petrography

The sheet volcanics (Nos. 10930 and 10931) are hypautomorphic granular albitised rocks with large (4mm. long) subhedral tabular laths of altered plagioclase (now albite with abundant inclusions) making up 60 per cent of the rock, subhedral titaniferous pleochroic augite 20 per cent, altered ilmenite (now leucoxene and hematite) about 5 to

10 per cent and an interstitial chloritic mesostasis making up only 5 to 10 per cent. The comparatively coarse-grain distinguished these rocks from the pillow lavas so that in handspecimen they are green coloured with a distinctly mottled appearance.

The pillow lavas and fragments from the volcanic breccias are all fine-grained spilites and generally with a variolitic (No.10932) or intersertal (No. 10933) texture. In No. 10932 crystallisation of original glass has given dendritic crystal growth which cuts through the earlier varioles of radiating albite aggregates. The matrix of the volcanics is usually packed with small granules of hematite. In Nos. 10933 and 10932 a small amount of weakly pleochroic (probably titaniferous) augite is present but in most, the ferromagnesians are altered to chlorite and/or actinolite. Altered titanomagnetite is a common accessory.

The sedimentary rocks are lithic or quartzo-feldspathic sandwackes. Lithic fragments are predominantly volcanic-derived and have a similar texture to the pillow lavas and fragments from the volcanic breccias. In No. 10934 there is a higher proportion of quartz, about 25 per cent, than is usually found in the sedimentary rocks of the Red Hill Area, but otherwise the sandstones are similar to those of the Pelorus Group.

<u>Distribution and thickness</u>

The Goat Formation has been examined closely only in the vicinity of Mt. Ellis but is presumed to continue northward in the headwaters of the Wairoa River. The thickness of the formation is given in section 9 of fig. 7 as about 6000 feet, but much of this may be due to post-depositional thickening, such as penecontemporaneous slumping.

Metamorphic rocks near the north-eastern contact are probably derived from basic volcanics (p.86) and together with associated sedimentary rocks, are placed in the Goat Formation. At 445780 volcanic rocks are exposed in the scree-covered slopes below the eastern contact of the Complex. The rocks are volcanic breccias similar to those of the Goat Formation with which they are correlated.

Depositional environment

The lithological similarity between the sedimentary rocks of the Goat Formation and Pelorus Groups suggests a comparable depositional environment which, because of absence of fossils, abundance of graded beds, and considerable thickness, is thought to be deep water and rapid accumulation.

The volcanics are thought to have derived, like the sediments, from a source marginal to the basin of deposition with vast submarine outpouring of basaltic magma flowing down the sides of the basin and forming pillow lavas. Then with redeposition perhaps in the manner of a submarine lahar, poorly bedded deposits of volcanic breccia were formed. Some magma probably intruded the volcanic pile to produce the rare sheet volcanics.

GLENNIE FORMATION

The Glennie Formation is defined as the sedimentary and volcanic rocks typically exposed in the section between 382940 and 400940 (fig. 7 section No. 9). The formation is about 3000 feet thick in the south and at least 4800 feet thick north of Porters Stream. It characteristically contains an abundance of sheet volcanics (spilitic basalts and albite dolerites). In contrast to the Goat Formation, this formation contains only rare pillow lavas if any, (none have been observed) and minor amounts of volcanic breccia. The formation is characterised by successive 10 - 100 feet thick sheets of massive albite dolerite and spilitic basalt separated by thin screens of coarse clastic sediments including brecciated fine-grained volcanics and fragments of angular quartz, sandstone and argillite. Its lower boundary is marked by dark grey argillites and green sandstone in fault contact with pillow lavas of the Goat Formation; its upper boundary by sheet volcanics in disconformable but regionally concordant contact with the Maitai Group.

Sedimentary member. The sedimentary member is composed of green sandstone and less common, black and grey argillites. There is a general increase in grain size from the base to the top of the member. Bedding is rarely preserved and the rocks are either strongly contorted or brecciated, with large angular fragments of coarse green sandstone, scattered through a finer sandstone matrix (fig. 9). These breccias probably formed by penecontemporaneous deformation.

Fig. 9. Penecontemporaneous breccia, sedimentary member of Glennie Formation. Angular fragments of green sandstone (light) surrounded by matrix of grey argillite (dark).

Volcanic member. The volcanic member comprises a succession of sheet-like bodies of spilitic basalt and dolerite. The sheet-like form of the volcanics is well shown on the east flank of Mt. Glennie, and in Ellis Stream and near the Wairau River at 344738. The sheets vary from 10 feet up to at least 100 feet thick, but most are about 25 feet thick. The centre of the sheets is medium-grained, with a doleritic texture (e.g. No. 10936) but about 4 feet from the contact the texture becomes intersertal. Within 2 to 3 feet of the contact, the sheets are brecciated and intimately mixed with sedimentary rock, which forms a thin unmetamorphosed screen between the sheets. Atomodesma prisms are common in the screens. The sheets are thought to be lava flows rather than dolerite sills because of the sedimentary screens, the wide zone of brecciation, and the wide chilled margin. It is estimated that sheet volcanics make up at least 90 per cent of the volcanic member, with volcanic breccias only locally abundant, and pillow lavas (if any) very rare.

Volcanic breccias occur as lensoid-shaped masses enclosed in sheet volcanics near the Wairau River and as a thin band underlying the Wooded Peak Limestone at 346770. They are more highly coloured than those of ~~Ellis~~ Goat Formation, with green volcanic fragments set in a red, hematite-rich matrix, and are composed of fragments of albite dolerite and spilitic basalt similar to the sheet volcanics. The breccias are probably channel fillings caused by erosion during the deposition of the formation.

No pillow lavas have been observed by the writer in the Glennie Formation.

Petrography

Sedimentary rocks. The sedimentary rocks are composed of poorly sorted angular fragments of plagioclase, quartz, sedimentary and volcanic rock fragments. Quartz is only abundant in the finer-grained rocks such as No. 10935 (a siltwacke) in which it makes up about 40 per cent of the rock and about 80 per cent of the identifiable fragments. The coarse-grained rocks are composed predominantly of lithic fragments but are so altered that relative proportions of fragments is impossible to estimate. The alteration products are authigenic micas (in the finer-grained rocks) and prehnite and pumpellyite. Accessory minerals are exogenic epidote, chlorite and pyroxene.

Volcanic rocks. At 344738, three steeply dipping, volcanic sheets are well exposed, They are each approximately 25 feet thick and have a brecciated zone about 3 feet thick separating them. The middle of the sheets (specimen No. 10936 has a medium-grained, sub-ophitic texture with augite grains (comprising 25 per cent of the rock) 2mm. in size, partly surrounding 1.5mm. tabular laths of albite (50 per cent) which are charged with minute granules of a highly refringent mineral, probably epidote. An interstitial chloritic matrix makes up about 20 per cent of the rock and the remainder is composed of altered skeletal ilmenite or titanomagnetite - now represented by leucoxene and iron ores - and accessory sphene, epidote and a few needles of apatite. About 2 feet from the contact, near the edge of the brecciated zone, the rock (No. 10937) has an intersertal texture (fig. 10) in which divergent laths of cloudy albite make up about 45 per cent of the rock

Fig. 10. Spilitic basalt, photomicrograph No.10937. Albite occurs as tabular laths and acicular grains; intergranular anhedral augite; mesostasis of brown chlorite devitrified glass.

with intergranular anhedral augite (15 per cent) granules of iron ore (5 per cent) and chloritic and iron-stained matrix (35 per cent). The volcanic fragments of the marginal breccia (No. 10938) are very fine-grained, with a variolitic texture of highly acicular feldspar crystals set in an indeterminate granular matrix. The clastic matrix of the breccia is composed of grains of augite, feldspar and rock fragments. Most of the other specimens collected throughout the formation are similar to those described above. In some the feldspar is clear, euhedral albite containing only a few discrete grains of epidote (No. 10939) and in others the feldspar is strongly zoned with inner zones altered to a mat of saussurite, and only the outer zones of oligoclase and albite persist clear and unaltered (No. 10940). In some rocks the augite is replaced by actinolitic amphibole (No. 10941) or uralitised to a fine grained mat of brown and green amphibole (No. 10939).

Apart from these minor differences the sheet volcanics show little variation in composition. Plagioclase is invariably very sodic and is usually albite, and because of the general abundance of granular inclusions of calc-silicates was probably formed by alteration of initially more calcic plagioclase. The relative proportions of different minerals is also remarkably constant, feldspar comprises about 55 to 65 per cent, ferromagnesians (pyroxene or alteration products) 20 to 30 per cent, ilmenite (or alteration products) between 4 and 10 per cent, and the chloritic matrix is variable but commonly about 20 to 30 per cent. Accessory minerals are epidote (probably deuteric origin) sphene and apatite. No rocks containing more than 50 per cent ferromagnesians have been found, but some rocks are composed dominantly of albite with only minor ferromagnesians (e.g. No. 10941).

The volcanic breccia is composed entirely of fragments with similar textures and compositions to the rocks described above.

D i s t r i b u t i o n   a n d   t h i c k n e s s

The sedimentary member of the formation, found only in and north of Porters Stream, occurs between the Red Hill ultramafics and the overlying sheet volcanics. South of Porters Stream, the sheet volcanics occur up to the contact with the Complex and in places are metamorphosed. In all stratigraphic sections examined the sedimentary rocks are faulted against the ultramafics, and therefore only a minimum thickness - 800 feet - can be obtained for the member.

North of Porters Stream, the volcanic member is faulted against the Greville Formation, and is at least 4000 feet thick. In the south the thickness between the top of the formation and the contact with the Complex varies from 3000 feet in the Wairau Valley to only 400 feet in Maitland Stream. Much of this apparent thinning is probably due to transgressive intrusion by the ultramafic rocks, but it is also possible that the formation does not maintain thickness laterally and may thin and thicken considerably.

The sedimentary member was probably deposited in deep water, because of the same reasons given for deep water deposition of the Goat Formation. Penecontemporaneous deformation was common in both formations, causing the development of the brecciated rocks and intra-formational slumping. In contrast, the volcanics were possibly deposited in relatively shallow water, because of the fossiliferous interbedded sedimentary rocks. If so they are the oldest rocks of the Upper Paleozoic sequence on the western limb of the Nelson Syncline to have been deposited in shallow water. In this respect they are more closely related to the overlying Wooded Peak Limestone and Tramway Sandstone, than to the rest of the Lee River Group.

M A I T A I   G R O U P

The Maitai Group is subdivided into five formations (Waterhouse, 1964) of which only the four lowest are present in the Red Hill area:

Stephens Formation (absent)

Waiua Formation

Greville Formation

Tramway Sandstone

Wooded Peak Limestone

Rocks of the Maitai Group form a gently arcuate belt of steeply dipping rocks on the west of the mapped area. The Wooded Peak Limestone and Tramway Sandstone are exposed only in the south of the belt; north of the right branch of the Motueka River, the Greville Formation is faulted against the Glennie Formation.

WOODED PEAK LIMESTONE

Waterhouse gives this name to the basal limestone of the Maitai Group. The name was initially applied only to the lowest member of the limestone (Waterhouse, 1959) but, "it can be used in the broader sense as well." (Waterhouse, 1964 p. 28). Here, it is used in the broader sense as the calcareous and sedimentary rocks which overlie the Glennie Formation of the Lee River Group and underlie the Tramway Sandstone.

In the Red Hill area the Wooded Peak Limestone is covered with beech forest and is not well exposed except near the contact with the volcanics.

Content

In the south of the area the limestone (which is 1400 feet thick there) rests directly upon the Glennie Formation but at 340798 a blue-green sandstone appears separating the two and progressively thickening toward the north. The sandstone resembles the blue-green sandstone member of the Tramway Formation but is included as a member of the Wooded Peak Limestone. A thin section of the sandstone (No. 10942) shows that it is poorly sorted feldspathic-sand-wacke with a predominence of feldspar fragments and abundant fragments of volcanic rocks, clinopyroxene and chlorite. Sub-rounded quartz grains make up less than 5 per cent of the section.

The limestone in contact with the volcanics is a dark to light grey banded limestone, and intense folding of the bands is believed to be caused by penecontemporaneous slumping of poorly consolidated calcareous debris (fig. 11). Higher in the formation unfolded shale laminae and silt bands are abundant. Within the limestone are rare thick lenses of grit and sandstone. These have been mapped as a separate member of the formation along with the blue-grey sandstone described above. The grit beds, some of which are 20 feet thick, are composed of about 50 per cent fragments of calcite and the rest of feldspathic, volcanic and chloritic fragments. Near the top of the formation a sandy limestone which is widespread, passes into the non-calcareous rocks of the Tramway Sandstone.

Distribution and thickness

On the watershed between the Wairau and Motueka River systems the Wooded Peak Limestone is 1400 feet thick and limestone is by far the dominant member. The formation thins northwards to less than 500 feet thick at 337799 and the lower sandstone member is also about 500 feet thick. Limestone is absent in the Motueka Gorge at 335808 but large limestone blocks in the stream bed at 337812 probably indicate a small outcrop of limestone nearby. This has not been found in situ.

Relationship to Glennie Formation and mode of deposition

The contact between limestone and volcanics is in part well exposed and follows a broadly sinuous curve (fig. 20, p. 82) Penecontemporaneous slump folds in the limestone are common at the base of the formation. The limestone was clearly deposited on an uneven and undulating surface on the volcanics in the first instance as slumps of calcareous debris into basins in the sea floor and later as a thick accumulation of limestone. Waterhouse (1964 p. 29) suggested that the limestone "represents a vast calcareous bank derived from the steady accumulation of broken shells of Atomodesma" and that "...Deposition was probably at shallow to moderate depths...". The thinning of the limestone to the north suggests that the bank may have been discontinuous as a series of reefs rather than a continuous bank. The attitude of the volcanics near the contact is parallel to that of the limestone. There is no evidence however for a major erosion break between the two groups, but it is probable that the contact is disconformable.

Fig. 11. Shaly lamination in Wooded Peak Limestone. Slump folding of limestone is shown by tracing such lamination over distances of several yards.

Fig. 12. Autoclastic breccia in Tramway Sandstone. Fragments are of black argillite.

TRAMWAY SANDSTONE

The name Tramway Sandstone is given to that formation of sandstone with interbedded banded siltstones that overlies the Wooded Peak Limestone and underlies the laminated argillites of the Greville Formation (Waterhouse, 1959).

Content

On the ridge between Maitland and Beebys Streams the Tramway Sandstone comprises:-

		South	North
Greville Formation	Laminated argillites.............		
Tramway Sandstone	Coarse grey-green sandstone......	50	100 ft
"	Fine steel-blue sandstone........	600	200 ft
"	Fine grey-dark grey fossiliferous siltstone	200	- ft
"	Black sulphurous argillite.......	50	- ft
"	Coarse green sandstone...........	100	200 ft
	Thickness	1000	500 ft

The formation thins northward by lateral gradation into laminated argillites of the Greville Formation. The topmost sandstone is lithologically indistinguishable from the sandstone member of the Greville Formation and the lowest sandstone is lithologically very similar to the lower sandstone member of the Wooded Peak formation but both are mapped as Tramway by definition.

The section cut by the Motueka River at 335808 comprises:-

Greville Formation	Laminated argillites	
Tramway Formation	Graded beds 5-10 ft thick	100 ft
"	Massive green sandstone	300 ft
"	Blue-green sandstone	200 ft
"	Banded silt and sandstone	200 ft
"	Coarse green sandstone	300 ft
"	Banded grey argillite	200 ft
Fault		
Peridotite		
	Thickness	1300 ft

In the absence of limestone, the whole of the section is mapped as Tramway Sandstone. The lateral facies change between massive green sandstone, graded beds and laminated argillites of the Greville Formation is particularly well shown in this section. The graded beds contain intraformational breccia of argillite which form elongate flattened fragments up to 8 inches long (fig. 12). These tend to lie parallel with long axes pitching south west and flattened in the planes of bedding. The southwest lineation probably indicates direction of transport. The Tramway Sandstone is commonly fossiliferous, containing abundant fossils of the Permian pelecypod Atomodesma. The finer grained members commonly have a well developed slaty cleavage (fig. 13).

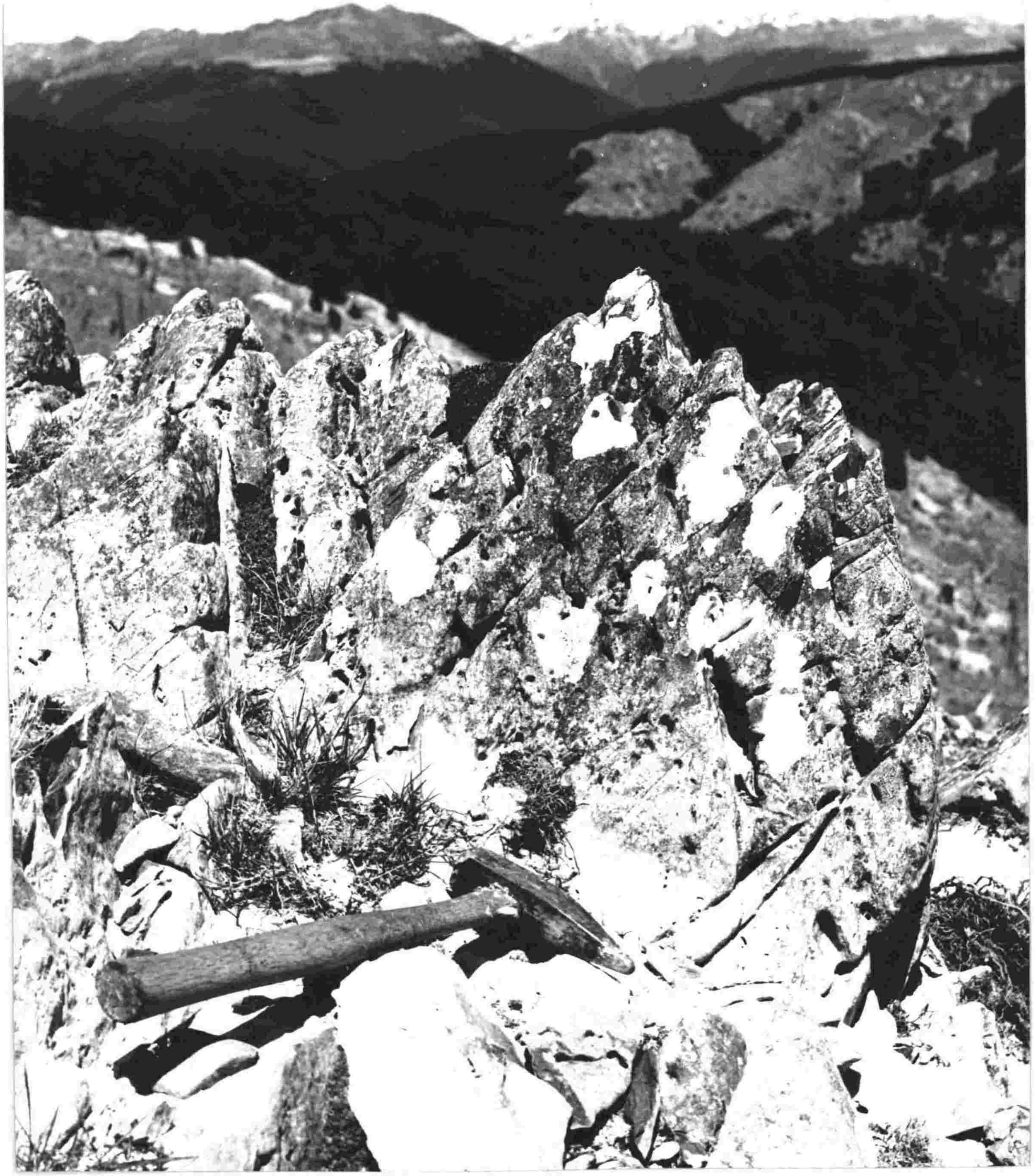

Fig. 13. Fossiliferous Tramway Sandstone.
Slaty cleavage dips to the left, bedding marked by aligned cavities formed by weathering of Atomodesma shells, dips steeply to the right of the photograph.

Petrography

A thin section (No. 10943) cut from the sandstone fraction of the graded beds in the Motueka River section shows that the rocks are moderately well sorted (grain size 1 mm. ranging from 0.2mm. to 2mm.) and composed largely of subangular fragments of spilitic and doleritic volcanic rocks. Altered feldspar is abundant and clinopyroxene, calcite and chlorite fragments are common while sub-rounded quartz grains make up less than 2 per cent of the rocks. The rock is best described as a lithic arenite.

The steel-blue sandstone is a typical member of the Tramway Sandstone Formation and is described by Waterhouse (1964) elsewhere in the Nelson area. It is poorly sorted but of similar composition to the graded beds.

Thickness

The Tramway Sandstone in the Red Hill area varies from 500 to 1300 feet thick, as described above.

Relationship to Wooded Peak Limestone

The Wooded Peak Limestone grades laterally into the Tramway Sandstone. This is most clearly shown on the north end of the ridge between Maitland and Beebys Streams. There it is possible to walk along the strike of bedding planes from limestone to Tramway Sandstone. At right angles to the strike however, the change appears to be abrupt, sandy limestone passing into non-calcareous sandstone within a distance of a few feet.

Mode of deposition

Waterhouse (1964) suggested that the Tramway Sandstone was deposited at moderate depth, probably slightly greater than that of the limestone. This view is supported by the evidence of facies change between limestone and sandstone and between sandstone and laminated argillites described below.

GREVILLE FORMATION

Content

In the Red Hill area the Greville Formation contains three facies; dominant and characteristic laminated argillites, a coarse-grained green sandstone, and a volcanic conglomerate. Waterhouse (1964, p. 34) described similar facies elsewhere in the Nelson Upper Paleozoic Belt.

The coarse green sandstone facies forms large lenses which grade laterally into laminated argillites, and has been mapped as a separate member. Watters (in Waterhouse, 1964) has given the petrography of several thin sections of this facies from the Greville Formation north of the Red Hill area. The rocks are feldspar-quartz-, volcanic-, and altered feldspathic- arenites. Similar rocks are present in the Red Hill area. One rock (No. 10944) from 325850 is unusual in containing a high proportion of calcite as clastic grains and veins. It is a calcareous arenite and is a moderately well sorted, coarse sandstone, with calcite 40 per cent, lithic fragments 40 per cent (volcanic and sedimentary rocks in equal abundance) and feldspar 10 per cent. Fragments of quartz, chlorite and epidote are accessory.

The conglomerate member was studied in detail because a petrographic description of these rocks has not been previously published. The conglomerate is largely massive but well defined bedding and grading is widespread. It varies from a coarse cobble to a pebble conglomerate with a sandy matrix (fig. 14). It is exposed in Harvey Stream and has been traced southward for 1 mile to the Motueka River where it thins out. Large boulders of conglomerate in the lower part of the Motueka River suggest that another lens may be present further south but that has not been found in place. By far the most common rock type in the cobbles and pebbles of the conglomerate is a green, porphyritic volcanic rock, but also present are fine-grained marble (only one small pebble found) and rare sedimentary rocks. Several thin sections of the porphyritic volcanic rocks were studied. All rocks are considerably altered, the feldspar to albite, pumpellyite and/or calcite; the matrix to pumpellyite and chlorite; and the ferromagnesians now represented only by pseudomorphs of secondary minerals. However the original texture is well preserved and identification of the volcanic rocks must depend almost wholly on texture and relative proportions of felsic to femic phenocrysts as the original composition of the minerals is lost by metamorphism.

No. 10945 A porphyritic pyroxene andesite. Euhedral, altered plagioclase occurs in two generations as 4mm. tabular crystals and as small elongate laths, 5mm. in size. There is a weakly defined parallelism in the crystals. Chlorite and calcite pseudomorphs

Fig. 14. Igneous conglomerate from the Greville Formation.

(probably of pyroxene) have short, stumpy prismatic and equidimensional octagonal cross sections. With this interpretation the proportion of pyroxene to feldspar is about 1:3 suggestive of an andesite rather than basalt.

No. 10946 An altered hornblende andesite. It has a porphyritic texture with euhedral 3mm. plagioclase crystals making up 40 per cent of the rock and euhedral hornblende 15 per cent. The hornblende is now represented by pseudomorphous finely divided hematite and chlorite but the characteristic lozenge and prismatic cross section are well preserved (fig. 15). The groundmass is of finely granular chlorite and calcite, pumpellyite is absent.

No. 10947 A fine-grained volcanic rock with a felsitic texture of randomly oriented small acicular laths of feldspar in a matrix (30 per cent) of finely divided sphene, pumpellyite and chlorite.

No. 10948 A medium-grained igneous rock with subhedral diopsidic augite 35 per cent, subhedral tabular feldspar 50 per cent and a chloritic matrix 15 per cent. Sphene surrounds relict opaque grains and occurs as fine-grained aggregates in the matrix.

The sedimentary rocks of the conglomerate pebbles are predominantly green sandstone, and red and green laminated argillites. These rocks are lithologically very similar to rocks of the Pelorous Group.

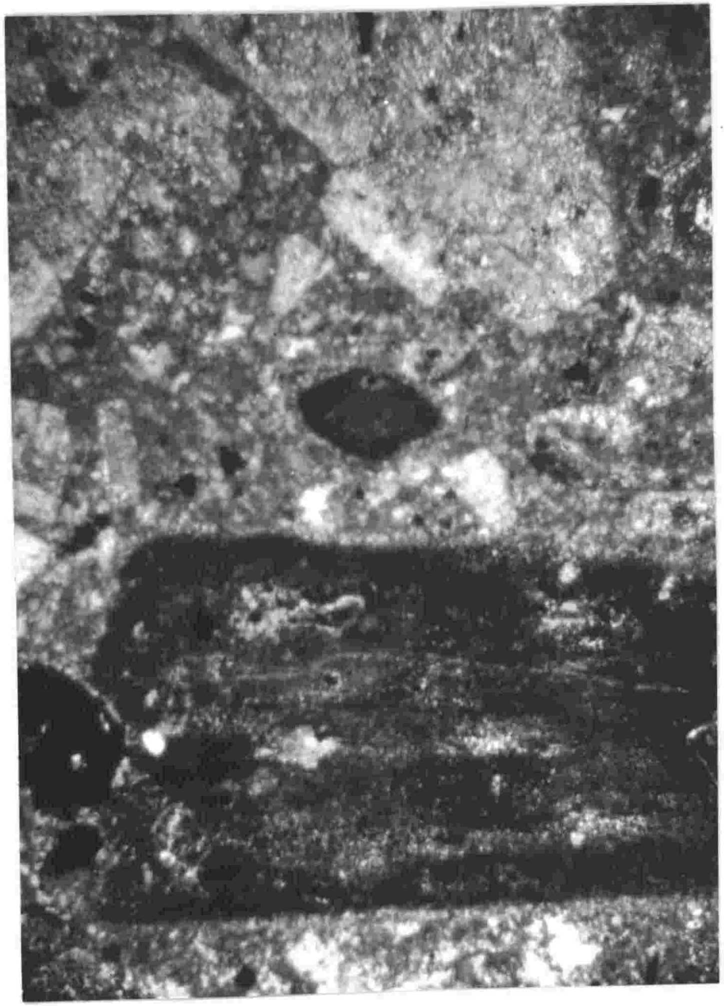

Fig. 15. Hornblende andesite. Euhedral hornblende (dark) altered to finely disseminated hematite granules.

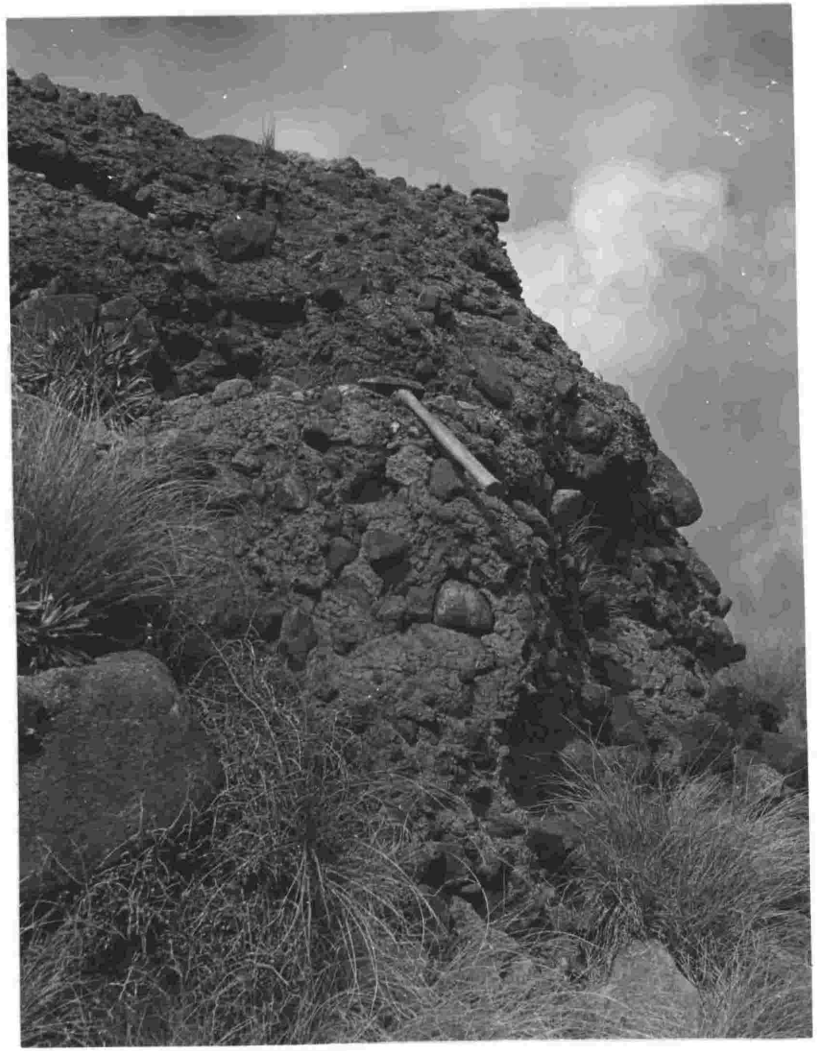

Fig. 16. Peridotite moraine exposed on edge of the Plateau. above Motueka River. Although probably late Pleistocene in age the moraine is very strongly cemented.

Distribution and thickness

The Greville Formation forms a gently arcuate belt on the west side of the mapped area (see geological map). The rocks dip uniformly steeply west, but small asymmetric zig-zag folds with an amplitude of about 100 feet are common in the upper part of the formation. Two such folds are well exposed in the Motueka River at 325850. Their axial planes, dip at about 60^{o} to the east, parallel to the axial plane of the Nelson syncline (p. 11).

Because of the unknown extent of folding the thickness of the Greville Formation cannot be measured accurately. The thickness between formation boundaries in the upper part of Beeby Stream is about 6300 feet and the stratigraphic thickness is there estimated at between 4000 and 6000 feet, probably of the order of 5000 feet.

Relationship to the Tramway Sandstone

The laminated argillites in the lower part of the formation grade laterally into Tramway Sandstone, but most of the Greville Formation is younger than the Tramway Sandstone.

WAIUA FORMATION

The Waiua Formation was not studied in detail. The boundary between it and the Greville Formation was taken as the western limit of the mapped area. As elsewhere (Waterhouse, 1964) it consists predominantly of laminated red and grey argillites but a coarser facies of a speckled red and green sandstone is exposed in the Motueka River at 320855.

AGE OF THE MAITAI GROUP

The age of the Maitai Group is discussed by Waterhouse (1964) who gives a Kazanian age for the Wooded Peak, Tramway and Greville Formations.

TORLESSE GROUP

The name Torlesse Group was formally proposed by Suggate, (1961) for poorly fossiliferous, non-schistose greywackes and argillites that lie to the east of the schist belt in the South Island of New Zealand. These rocks have been earlier described as undifferentiated greywackes (Grindley, Harrington and Wood, 1961) or Alpine Facies (Wellman, 1956).

In the Red Hill area the Torlesse Group occur to the south as a monotonous succession of beds of redeposited facies. Generally, the rocks are grey to greenish-grey greywacke sandstone and dark grey argillites. Grading and micro-cross-stratification are widespread.

Microscopically the rocks are poorly sorted and composed of angular fragments of quartz, microcline and plagioclase. A typical example, No. 10949 is a quartz-feldspathic wacke sandstone from 463759. The matrix (fragments less than 0.1mm. in size) constitute 40 per cent of the rock. The large fragments range up to 1.5mm. but most are in the range 0.5 to 0.75 mm. Of these, quartz predominates (40 per cent of the rock), microcline (10 per cent), plagioclase (5 per cent) and accessory sphene, biotite and muscovite, Incipient recrystallisation of the matrix is shown by the weak development of pumpellyite and veinlets of prehnite. Quartz veins are rare in the Torlesse Group.

A few miles east of the mapped area, a lens of fossiliferous limestone occurs in the Torlesse Group. (Lensen, pers. comm.) The fossils have not been definitely identified but indicate a probable Triassic age.

Although of similar depositional environment, the Torlesse Group differs from the Pelorus Group in age and provenance. Pumpellyite and epidote are the usual metamorphic minerals of the Pelorus Group whereas pumpellyite and prehnite are the characteristic minerals of the Torlesse Group in the Red Hill area. Also, in general, the rocks of the Pelorus Group are more altered than the Torlesse Group. Whereas the Pelorus Group is derived from basic volcanics, the provenance of the Torlesse Group, as indicated by microcline, sphene, biotite and an abundance of quartz, is probably granitic. In this respect the Torlesse Group of the Red Hill area is similar to the Mesozoic greywackes and argillites of Wellington (Reed, 1957).

QUATERNARY DEPOSITS

The geological map of the Red Hill area shows two types of Quaternary deposits; river gravels and moraine.

River gravels cover the Wairau valley bottom and extend several hundreds of feet up the valley walls. Aggradation terraces of several ages are preserved, the highest of which is about four hundred feet above the river. Some river gravels contain interbedded consolidated siltstones which dip at about 20°. These are probably mid to early Quaternary in age. Others are superficial recent deposits.

Small areas of moraine cover parts of the peridotites of the Red Hill Ultramafic Complex and are evidently the remains of the more extensive sheet that developed during Quaternary glaciations. Most of the moraine is composed entirely of rounded and subangular boulders and pebbles of unserpentinised peridotites set in a very poorly sorted, strongly cemented matrix. (Fig.16). The cement is thought to be serpentine developed through weathering of the matrix. This is because x-ray diffraction patterns of the matrix have revealed only serpentine, but it is possible that a non-crystalline amorphous material may also be present.

The patch of moraine at 356740 is composed of angular and rounded fragments of limestone, fossiliferous sandstone and argillite of the Maitai Group. It occurs 1200 feet above the Wairau River and was presumably transported at least a mile and a half from the nearest outcrop of the Maitai Group to the east.

D U N   M O U N T A I N   U L T R A M A F I C S

Dun Mountain Ultramafics is a name used, though not formally defined, by Waterhouse (1964) for the major sheets of ultramafic rocks in the Nelson Upper Paleozoic Belt. The smaller concordant sills and bodies of ultramafic rocks, distinguished by him from the Dun Mountain Ultramafics, are included as members of the one or other formations of the Lee River Group. In this thesis the name Dun Mountain Ultramafics is applied to all ultramafic rocks of the Nelson Ultramafic Belt, regardless of their form or size. In this sense, 'Dun Mountain Ultramafics' is a term recognising regional association and distinctive lithology of mappable igneous rock masses. The term has the same significance as 'Formation' and 'Group' in stratigraphy.

The Dun Mountain Ultramafics in the Red Hill area comprise rocks of the Red Hill Complex and numerous small bodies that are distributed widely throughout the Lee River Group and to a lesser extent, the Maitai Group and referred to as the Minor Intrusions.

The Minor Intrusions are composed of serpentinites or serpentinised peridotite.* The rock types of the Ultramafic Complex

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* The distinction made here between serpentinite and serpentinised peridotite is based on petrographic evidence. Serpentinised peridotite is an ultramafic rock that can be shown to be derived from an initial peridotite. It possesses mesh texture after olivine, relict grains of pyroxene or olivine, bastite pseudomorphs etc. The parent material of serpentinite cannot be shown on petrographic evidence to have ever been peridotite. In texture, it commonly consists of bladed antigorite crystals, or interlocking fibres of serpentinite. Commonly, however, it will have been derived from peridotite and the texture has been modified by deformation.

are much more varied. Serpentinites occur at the margins of the Complex and partly serpentinised periodite is common but most of the rocks are fresh and unserpentinised. Harzburgite, containing about 20 per cent orthopyroxene and subordinate diopside and spinel is the predominant rock type and dunite is the next most common. Other rock types are pyroxenite, wehrlite, lherzolite and feldspathic peridotite (feldspar constitutes about 2 to 5 per cent of the rock). Highly feldspathic rocks, eucrites and anorthosite veins also occur but their extreme rarity makes it possible to refer to the rocks of the whole Complex as ultramafics. When referring specifically to feldspar-rich rocks the term 'gabbroic' is used.

MINOR INTRUSIONS

Ultramafic rocks in orogenic belts characteristically occur as small concordant lenticular intrusions. Benson (1926, p.69) wrote, "It is very clear from many instances we have described how correctly are the basic and ultrabasic rocks described as concordant sills in dislocated mountains..." More recent work shows that it is also common for the ultramafic rocks to occur as highly sheared, slickensided bodies crosscutting the enclosing strata. Taliafferro (1943) observed many such bodies in the Franciscan-Knowville rocks of California and concluded that some serpentinites had "moved, generally upward, long after consolidation as a cold, probably rather plastic mass into the crest of folds or along faults." (op. cit., p. 102). Emplacement in this manner was called by him cold intrusion and was distinguished from 'ordinary intrusion' which involved "Emplacement of the igneous rocks as a magma rising from the depths in a molten or plastic state..." (ibid.). Cold intrusion is a widely accepted mechanism of emplacement and, as Hess (1955) pointed out, complicates dating the initial emplacement of peridotites. Long after initial intrusion, serpentinites and any involved ultramafics may move upward into younger overlying strata, and in the absence of any criteria to distinguish 'cold' from 'ordinary' intrusions the ultramafic rocks, as a whole, may appear younger than they are.

Marginal faulting and internal shearing are insufficient by themselves to distinguish ultramafic cold intrusions because it is to be expected that 'ordinary' intrusions may be affected by post-emplacement deformation giving similar features. High temperature contact metamorphism, if present, would indicate an 'ordinary' as opposed to a 'cold' intrusion but such contacts are rare or unknown for smaller Alpine-type ultramafic bodies. Chesterman (1960) has described a mineral assemblage of the glaucophane-schist facies in greywackes adjacent to small serpentinite sills. The development of the assemblage is attributed to thermal metamorphism, but Coleman (1961) ascribed similar mineral assemblages to low temperature - high pressure metasomatism, and the intrusion of the sills as a hot magma cannot be regarded as proved.

Conversly, the absence of a high temperature contact cannot be regarded as proof of 'cold' intrusion. Coleman (in press) described in detail low temperature contacts of the ultramafic bodies in the New Zealand Upper Paleozoic and deduced from this, together with the

Fig. 18. Peridotite breccia. Rounded to subangular fragments of unsheared serpentinised peridotite in a matrix of sheared serpentinite. This occurs as a tabular body, 10 feet thick and several hundred yards long and is probably a ' cold intrusion' of peridotite along a fault.

extensive marginal faulting of the bodies, that the ultramafic rocks were emplaced as 'cold' intrusions. Subsequently, Challis (1965b) recognised high temperature metamorphism on part of the contact of the Red Hill Complex. A similar contact is described later in this thesis (p. 81). This indicates that the Complex was emplaced as a very hot body.

The width of the contact aureole of the Red Hill Ultramafic Complex is only about 600 feet (p. 82) and probably the contact around smaller bodies was proportionately less. Thermal metamorphism adjacent to thick dolerite sills is commonly limited to within a few inches of the contact and if this relation is true also of ultramafic intrusions, it is possible that any high temperature effect was so small as to be readily obliterated by faulting. Therefore, although some concordant ultramafic bodies may have been emplaced into their present position by 'cold' intrusion, evidence of shearing or the absence of a high temperature contact is not sufficient to recognise them as such.

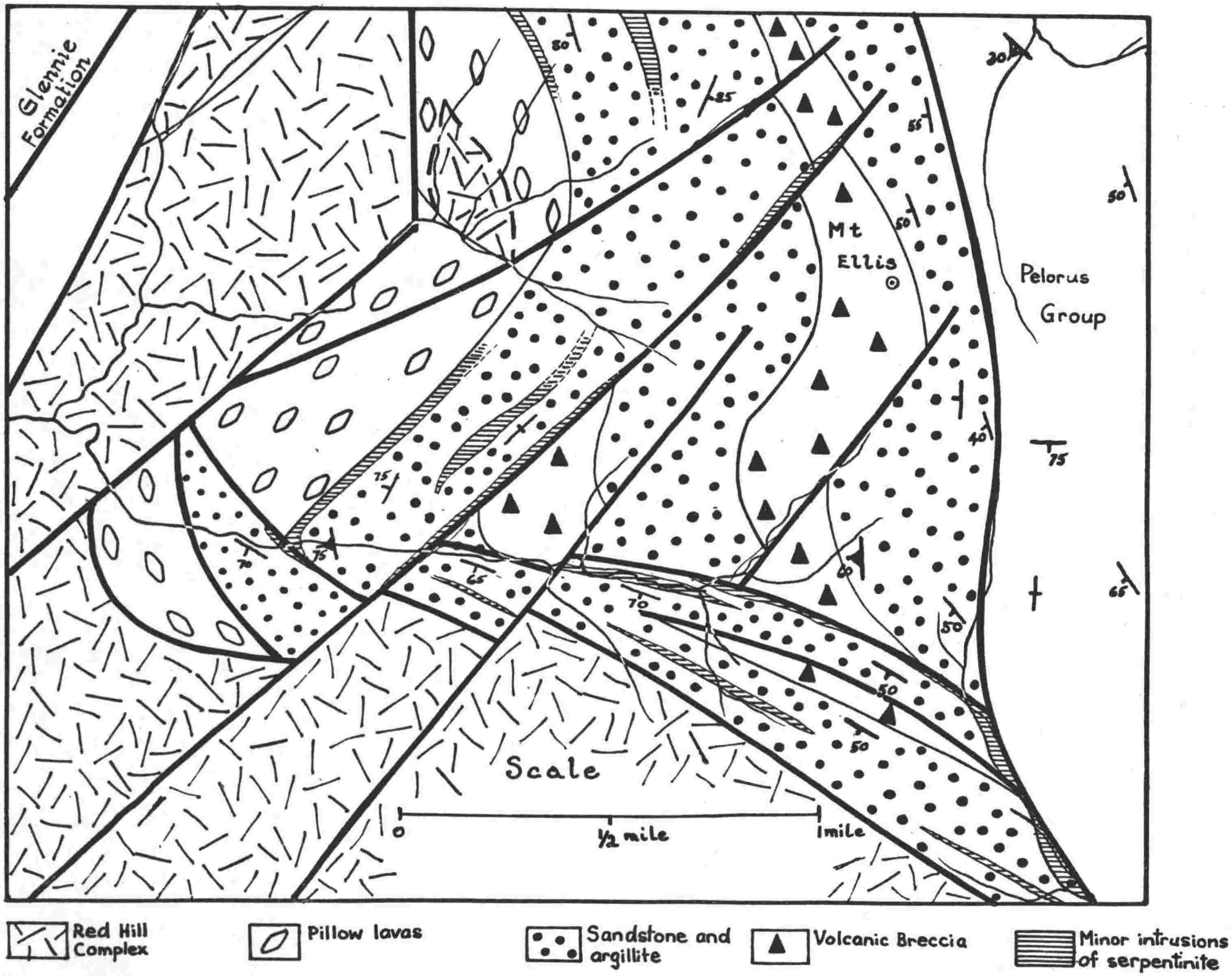
The only ultramafic bodies which may be reliably interpreted as 'cold' intrusions, are the sheared and slickensided bodies occupying fault zones discordant to the surrounding strata. 'Cold' intrusions of this type are common in the highly deformed rocks of the Goat Formation north of the Red Hill Complex and in the brecciated fault zones which cut the Maitai Group. They are shown on the Geological Map by the Letter 'S'. Those within the Maitai Group are commonly small, isolated bodies only a few feet in length surrounded by fault pug but some occur as large lenticular bodies like that in the fault exposed in

Porters Stream. That body is about 8 feet wide and at least 30 yards long.

Serpentinite in the Maitai Group invariably occurs in fault zones. 'Cold' intrusion appears to be the only reasonable hypothesis to account for their presence. The serpentinites are probably derived from peridotite enclosed within the Lee River Group underlying the Maitai Group. The date of emplacement of the 'cold' intrusions is not considered relevant in determining the age of the ultramafic rocks as a whole.

A large scale map of a small area near Mt. Ellis (fig. 17) shows both concordant and transgressive ultramafic Minor Intrusions in the Goat Formation. It has been necessary to exaggerate their width in order to show them on the map (fig. 17) but their general lenticular form is still evident. The transgressive intrusions are both more numerous and individually larger (20 feet thick) than those in the Maitai Group. Some have been traced for more than half a mile and usually thin and thicken throughout their length. Unlike intrusions of the Maitai Group they are composed of serpentinised peridotite rather than serpentinite. Commonly the intrusions are highly deformed with serpentinised peridotite preserved only as blocks in a matrix of sheared serpentinite (fig. 18). Wilkinson (1953) described similar serpentinite breccias from Queensland which are interpreted by him as fault fissure intrusions of peridotite which have been later disrupted by movement on the fault. The faults along which many of the serpentinite breccias are found in the Goat Formation (such as that of figure 18) are thought to be much younger (Upper Mesozoic) than the

Fig. 17. MAP OF THE AREA NEAR MT. ELLIS



initial period of emplacement of the ultramafic rocks (Permian). The breccias are therefore thought to be 'cold' intrusions of serpentinite which carried large blocks of peridotite into the fault.

Several concordant, lenticular bodies of serpentinised peridotite occur in the mapped area of figure 17. They are commonly sheared but this is consistent with the general deformation that caused the development of slaty cleavage in argillites in surrounding strata. They are composed of serpentinised periodite which occurs in large massive blocks separated by only thin shear zones of serpentinite. Contacts are invariably sheared. No evidence of high temperature contact metamorphism has been found but marginal discolouration of adjacent sediments and volcanics interpreted as low temperature metasomatism similar to that at the contacts of the Red Hill Complex is usual.

Transgressive intrusions have not been described elsewhere in the Upper Paleozoic. Waterhouse (1964) describes many instances of concordant bodies and it is possible that the transgressive intrusions of the Red Hill area are common only because of the strong deformation and folded structure of the rocks in that area.

R E D   H I L L   U L T R A M A F I C   C O M P L E X

The name Red Hill Ultramafic Complex is given to the body of ultramafic and genetically related rocks that occurs between Mt.Glennie and the Wairau Fault.

The Complex is considered to be a thick and tectonically widened part of an extensive single sheet of ultramafic rocks (p. 93) that extends northwards at least as far as Dun Mountain (New Zealand Geological Survey, 1:250,000 map sheets, Numbers 14,1964; 16,1962).

In the Red Hill area the sheet outcrops north of Mt. Glennie as a belt of rock about 2000 feet wide in the headwaters of the right branch of the Wairoa River (Geological Map). Due to poor exposure and difficulty of access, this part of the ultramafic sheet was not examined closely but is clearly shown in aerial photographs. It is, however, well exposed in a section cut by the left branch of the Wairoa River, a mile north of the mapped area. There, the ultramafic rocks, about 3000 feet thick, are composed of partly serpentinised harzburgite which is, in places, weakly foliated caused by parallel planar aggregates of orthopyroxene grains. The margins of the sheet are sheared and dip parallel to adjacent bedding in the Lee River Group. Internal faulting is widespread, but large blocks of unsheared peridotite, several chains wide are common. The sheet is also exposed in the Lee River about six miles to the north of the Wairoa River and has been mapped by Waterhouse (1964).

The Red Hill Ultramafic Complex covers 48.8 square miles and according to gravity data (Malahoff, 1963; and later discussion p. 90) it is at least a mile and probably almost two miles thick. The Complex is divided into two lithological units named the Basal and Upper Zones. Only the Basal Zone is shown on the Geological Map. It consists of massive non-layered harzburgite which has a strongly developed protoclastic texture. The Upper Zone, which covers the remainder of the Complex, is composed of layered and foliated rocks of varied composition. Harzburgite, dunite, and eucrite are present, together with veins of pyroxenite, anorthosite and pods of chromite. The terms 'Basal' and 'Upper' are derived from the relative positions of the two zones. Basal Zone rocks occur near the northern and eastern contacts which from structural considerations (p. 88) are considered to be at the base of the Complex. Upper Zone rocks cover most of the area of the Complex. The Complex is cut by numerous dykes called the mafic dykes composed of intermediate or basic plagioclase, hornblende, ortho-, and clinopyroxene. The dykes are dense black rocks, which cut across the layering of the peridotites and are closely parallel over the whole of the Complex.

The petrology and detailed description of the peridotites, serpentinites and gabbroic rocks of the Complex, and the intrusive mafic dykes, is given later in the thesis. Here a general description of the relationship of the Complex to surrounding strata is given.

The Marginal Fault

The contact of the Complex with the enclosing rocks is nearly everywhere marked by crushed and sheared serpentinite in a zone here called the marginal fault. The only exception occurs near Maitland Stream (at 348760) where massive serpentinised peridotites are in direct contact with metamorphosed volcanics (p. 81). At that point a 10 foot wide sill-like body of sheared serpentinite, about 20 yards from the contact, is thought to mark the continuation of the marginal fault.

Crushed serpentinite is easily eroded, but sedimentary rocks near the contact are usually strongly indurated and very resistant to erosion. Consequently at the contact of the Complex (against sedimentary rock) a broad asymmetric trench surmounted on one side by knolls of sandstone or argillite (fig. 19) has developed. Volcanic rocks on the other hand, are comparatively easily eroded and the marginal trench is broad and the contact defined only by vegetational contrast.

Relationship to Maitai and Pelorus Groups

Maitai and Pelorus Groups are in direct contact with the Complex in a few places but nowhere do ultramafic rocks cut into, or develop high temperature contacts with these rocks. The absence of Lee River Group rocks at these places is ascribed to post-emplacement faulting.

Fig. 19. Western contact of the Red Hill Complex.
Left branch of the Motueka River on right of photograph.

Maitai rocks are in contact with the Complex for a short distance north of the right branch of the Motueka River (at 338810) and there it is evident that the Glennie Formation has been faulted out by the conjunction of the marginal fault with a splinter fault. Movement on the splinter fault shown by the displacement of the Tramway/Greville boundary, indicates an apparent downthrow displacement to the west.

The Ellis Fault between Pelorus and Lee River Groups cuts into the marginal fault at 428896, and accounts for the disappearance of the Goat Formation to the east. The Wards Pass Fault probably follows the eastern contact for some distance from the north-eastern corner of the Complex and the only rocks of the Goat Formation that have been preserved occur in a small area to the south.

Thus it is likely that the Complex, before faulting,was entirely surrounded by Lee River Group rocks and was generally concordant with respect to the Maitai and Pelorus Groups.

At Dun Mountain Lauder (1965) believes that the ultramafic rocks intrude and are younger than the basal Maitai rocks. In particular he shows in his idealised column the Dun Ultrabasics as cutting across the Rangitoto Marble and intruding the Rangitoto Greywacke (fig. 1a, p.14 this thesis or fig. 3., p. 8 Lauder, 1965). This relationship is not evident on the geoloical map (Lauder, 1965. fig. 4).

The nature of the base of the Maitai Rocks in the Dun Mountain area has been discussed by Waterhouse (1959) who suggests that map data are consistent with either faulting, non-deposition or assimilation by the ultramafics. Non-deposition is eminently reasonable since on

Lauder's Map the Rangitoto Marble thins rapidly northward of Wooded Peak and may lens out. Lauder also gives a description of several altered rocks near to the contact of the Maitai Group with the serpentinites. One of these deserves special mention. It is described as a prehnite-cordierite-chlorite-sphene rock and is considered to have probably formed from greywacke. The presence of cordierite suggests that the rocks may have been thermally metamorphosed at high temperatures but it is by no means clear whether the parent rock is of Maitai age. Lauder states with reference to these rocks (p. 27) "Most of these rocks are similar to those produced by metasomatism and metamorphism of Little Twin Spilite and it is difficult to say whether they have been derived from marble, greywacke or volcanic rocks particularly as relations are complicated near the contact and as no unmetamorphosed rocks are present near by". If the parent rock were of Maitai age and the cordierite unequivocally definitive of high temperature metamorphism the age of intrusion would be established as post-Maitai. But there is only one example of cordierite rock and the age of that rock is uncertain. The other contact rocks mentioned by Lauder exhibit relatively low temperature assemblages (Coleman, in press).

RELATIONSHIP TO THE LEE RIVER GROUP

Apart from local transgressive contacts which can, in most cases be ascribed to displacement on major splinter faults that pass into the marginal fault, the Complex is essentially concordant with respect to the Lee River Group. But on the western margin of the Complex a transgressive intrusive contact is well displayed. The sedimentary member of the Glennie Formation is only found in, and north of Porters Stream; to the south, volcanic or metamorphosed volcanic rocks are in direct contact with the Complex.

The restricted distribution of the sedimentary member cannot be explained by faulting. The volcanic rocks in the vicinity of the Maitland Stream are in direct contact with the Complex and because of the thermal metamorphism of volcanic rocks the sedimentary rocks must have been absent at the time of intrusion.

Two possible explanations may be put forward to explain the absence of the sedimentary member in the south of the area.

(a) The western contact of the Complex is a transgressive, intrusive contact. The ultramafic rocks intruded at least part of the Glennie Formation cutting across the sedimentary member. This hypothesis carries the implication that high temperature metamorphism presumably developed over the whole margin of the Complex, has been faulted out in the north of the western contact.

(b) It is possible that the uppermost volcanics of the Glennie Formation are younger than the ultramafic rocks. If so, then the sedimentary rocks may also be younger, and were perhaps deposited in restricted basins separated by elevated blocks of metamorphosed volcanics. Later, the metamorphic and sedimentary rocks were overlaid by further volcanics of the Glennie Formation.

Because the sedimentary rocks of the Glennie and Goat Formations are lithologically similar (in the case of (b) it would be expected that the two would be considerably different) the first explanation is preferred.

Concordance of the Red Hill Complex

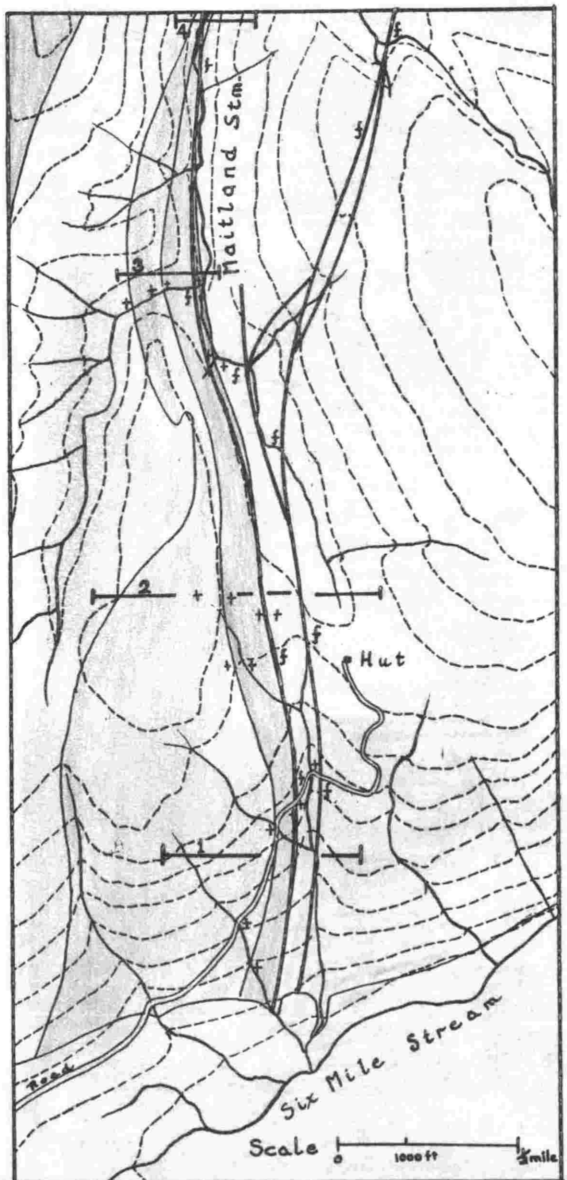
The thickness of strata cut during intrusion need not be greater than 800 feet to account for the present distribution of the sedimentary member of the Glennie Formation. This is small in comparison with the scale of intrusion and total thickness of the Lee River Group. Thus despite the minor transgressive western contact, the Complex is essentially concordant with respect to the Lee River Group; the Goat Formation strikes parallel to the contact where it is adjacent to the Complex and the sheet volcanics of the Glennie Formation, found only to the west of the Complex, are also regionally parallel to the contact.

The argument for concordance is strengthened when the whole of the Nelson Ultramafic Belt is considered. From D'Urville Island to the Wairau Fault, a distance of more than 80 miles, ultramafic rocks occur enclosed within the Lee River Group (Fig. 1). As in the Red Hill area, Maitai and Pelorus Groups are in contact with the ultramafic rocks in various places but nowhere with a possible exception do the ultramafic rocks cut either of the two Groups. The regional concordance is of importance in deducing age of emplacement of the ultramafic rocks as well as supporting the contention of essential concordance of the Red Hill Complex. The recognition of concordance is of considerable structural significance and is required for construction of a geological cross-section (p. 88).

Contact Metamorphism

High temperature. Dr. C. A. Challis has described high temperature contact metamorphism of volcanic rocks on the south-western contact of the Complex (Challis, 1965b). The metamorphic sequence passes from unaltered spilitic basalts to amphibolite and pyroxene hornfels facies close to the contact with ultramafic rocks. Challis considers that the grade of contact metamorhpism indicates a temperature of emplacement of the ultramafic rocks - an initial magma temperature - of about 1200°C. (Challis, 1965b).

The nature of the contact adjacent to the metamorphosed volcanic rocks is of considerable interest. The serpentinite zone surrounding the Red Hill Complex (p. 195) is here much narrower than elsewhere, approximately 100 to 50 feet wide. In fine detail the contact between serpentinised peridotite and metamorphic rocks is highly irregular and blocks of metamorphic rock are entirely enclosed within the ultramafics and vice versa. The contact may be regarded as a zone, about 100 feet wide, of an intimate mixture of ultramafic and metamorphic rocks. Within this zone some of the boundaries between serpentinised peridotite and metamorphic



fault f crush zone 1 Sections given in Fig. 7.

Form lines at approximately 100' intervals. + Identified specimens.

Ywa Wooded Peak Limestone Ygl Glennie Formation

Contact metamorphics h-f Quaternary

dm Red Hill Complex

Fig. 20 SOUTH-WESTERN CONTACT

rocks are marked by a thin crush zone but in many cases massive serpentinised peridotite is in direct contact with the metamorphics. This is very clearly demonstrated by Specimen No. 11015 which comes from the contact between a large block of amphibolite surrounded on three sides by massive, non-sheared serpentinised peridotite. In thin section brown hornblende of the amphibolite is separated from mesh textured serpentinised peridotite by a 5mm wide zone of chlorite, antigorite, vesuvianite and magnetite. It is highly unlikely that the two rock types in this case can have been faulted together and therefore this particular contact and the general contact zone in the area is very probably a true intrusive contact.

A recently made road-cut through the contact near the Wairau River provides a good section of metamorphic and volcanic rocks which was briefly examined by the writer. Metasomatism and later regional metamorphism has altered earlier plagioclase to albite and calc-silicates (hydrogrossular, prehnite, pumpellyite or epidote). Of the ferromagnesian minerals only hornblende and pyroxene persist, but these, with texture, suffice to distinguish volcanic rocks from the coarse-grained, hornblende-bearing hornfels that occur near the contact. The width and extent of the metamorphic aureole at the south-western contact is shown in figure 20. The positions of specimens identified as volcanic (containing volcanic texture) or metamorphic are given. Slightly crushed rocks observed between volcanic and metamorphic outcrops in the road section, and crush zones elsewhere may indicate that some metamorphic rocks have been faulted out. It is likely however that the present width of about 600 feet approximates closely to the initial width of the aureole. This is considerably narrower than the 6000 foot contact aureole of the Alpine-type peridotite body described by MacKenzie (1960) and considering the size of the Red Hill Complex, is very small.

Metamorphic rocks are also exposed at the north-eastern corner of the Complex where the following sequence is exposed:-

	Thickness
Massive peridotite	
Serpentinised peridotite	400 feet
Crushed serpentinite	70 feet
Coarse-grained amphibolite........ grading into Fine-grained hornfels.............	300 feet
Crushed serpentinite sill.........	10 feet
Concordant contact	
Unmetamorphosed sedimentary rocks.	

The 10 foot serpentinite 'sill' is probably a 'cold' intrusion of serpentinite in a fault separating the metamorphic from non-metamorphic rocks.

The fine-grained hornfels (No. 10950) has a fine-grained (0.05 to 0.1mm) granoblastic texture, and is composed of pleochroic, strongly birefringent epidote and elongate prismatic crystals of actinolite with a small amount of quartz, sphene and albite. The rock is cut by a quartz, albite and stilpnomalane bearing vein. Aggregates of a weakly pleochroic, pale yellow-green chlorite are found in small amounts throughout the slide. Both texture and association with higher grade amphibolite metamorphics suggest that the rock is a hornfels. Its mineral assemblage is that of the albite-epidote-hornfels facies of contact metamorphism (Fyfe, Turner and Verhoogen, 1958).

The fine-grained rocks grade into the coarser-grained rocks. A thin section of the coarse-grained amphibolite (No. 10951) 20 feet away from the ultramafic rocks, shows a granoblastic texture of stout prisms (4mm.) of brown-green amphibole ($N_y = 1.675 \pm .003$, 2V negative and very large, $Z \wedge c = 27^{\circ}$) comprising 75 per cent of the section, moderately birefringent epidote (2V negative and large) about 15 per cent, and the remainder non-magnetic opaque grains, sphene, actinolite (rimming the brown-green amphibole) and prehnite (as sieved xenoblastic crystals). Plagioclase is absent but actinolite and prehnite indicate retrogressive metamorphic effects and original plagioclase may have given up lime in the formation of prehnite. The large extinction angle, brown colour, and large 2V indicate that the amphibole is a hornblende rather than actinolite.

According to Fyfe, Turner and Verhoogen (1958) the assemblage hornblende-epidote is restricted to regional metamorphic facies. In contact metamorphic rocks the equivalent assemblages are hornblende-plagioclase (hornblende hornfels facies) or actinolite-epidote (albite-epidote hornfels facies). However it is considered that the field relations are unambiguous and indicate a contact metamorphic origin of these rocks. It is probable that the hornblende-epidote-plagioclase assemblage indicates higher load pressure than usual in contact metamorphosed rocks. The coexistence of intermediate plagioclase and epidote in regional metamorphic facies is attributed to Fyfe, Turner and Verhoogen (1958, p.229) to "high load pressures; perhaps augmented by non-hydrostatic stress".

The parent rock of the amphibolite and hornfels is probably basic igneous rock and is inferred to be volcanic rocks of the Goat Formation.

Low Temperature contacts. High temperature contact metamorphism is present only over a small part of the total circumference of the Complex. Elsewhere, volcanic and sedimentary rocks immediately adjacent to the contact show only incipient, low grade metamorphism. For instance, specimen No. 10952, a typical sand-wacke of the Pelorus Group collected within 4 inches of the contact at 435890 shows unaltered clastic texture although the matrix is extensively recrystallised with clusters of epidote and very coarse-grained, interlocking fibres of tremolite. In hand specimen, the rock is bleached, very brittle and hard, but there is no evidence of high temperature metamorphism. Similarly volcanic rocks near the eastern contact are relatively unaltered compared with the rocks of the south-western contact. The section exposed in the stream bed at 440771 is:-

	Thickness
Serpentinised peridotite	
Crushed serpentinite	50 feet
Black, indurated argillite......	25 feet
Indurated, green volcanics.....	50 feet
Crushed serpentinite lens	0-60 feet
Quartzose-feldspathic sandwacke..	

The black argillite (No. 10953) has traces of the original, very fine-grained, clastic texture preserved and contains scattered grains of angular quartz (with secondary growth of quartz around their edges) and epidote. Extensive recrystallisation is shown by abundant, very fine-grained fibrous tremolite. The green volcanics (No. 10954) have about a third of the rock composed of anhedral augite, 1mm. across, some showing shapes inherited from original ophitic texture. These are set in a matrix of very fine-grained chlorite and calc-silicates (x-ray diffraction patterns indicate the presence of chlorite and prehnite not distinguished microscopically). Original feldspar is presumably completely saussuritised. Accessory minerals are sphene and iron oxides.

High temperature contacts were probably developed over the whole of the intrusive contact but in most places these have since been faulted out.

The low temperature contacts, characterised by alteration limited to within a few feet of the ultramafic rocks, involve partial development of a low grade metamorphic mineral assemblages containing such minerals as prehnite, epidote and tremolite. Such minerals are typical of low temperature metasomatism related to serpentinites (Coleman, in press), and have probably developed during regional metamorphism.

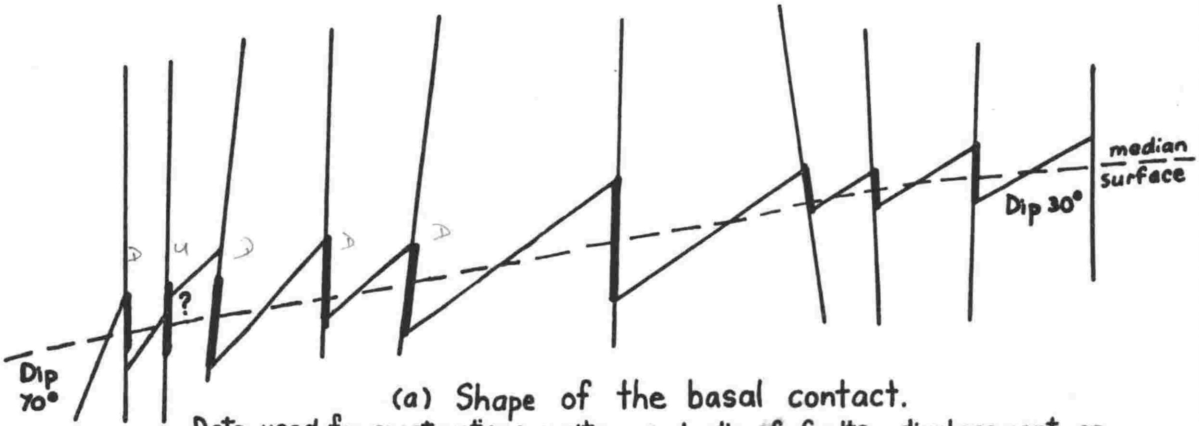
THE GEOLOGICAL CROSS-SECTION

A geological cross-section of the Red Hill Complex accompanies the geological map. The line of section is shown on the map as A - A'. Its construction was based on the following deductions and data.

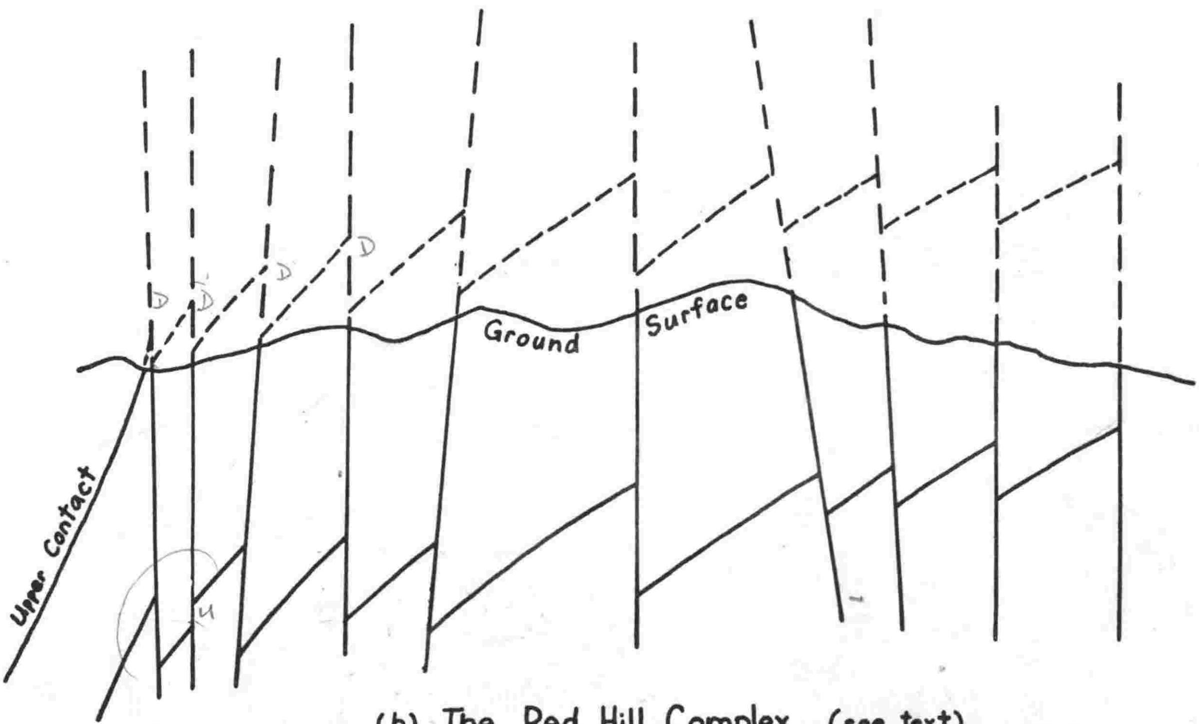
The Shape of the Basal Contact

(i) Dip of the basal contact of the Complex. The Complex is generally concordant with respect to the Lee River Group and the basal contact of the Complex may be expected to be parallel to the strata of the adjacent Goat Formation. On the map these contacts are shown as faulted. They are in fact zones of crushed serpentinite and are regarded as planes of movement between formations of widely different strengths. On the southern end of the eastern contact the basal contact of the Complex dips 50° N.W. and at the eastern and western parts of the northern contact it dips at 60° S.W. and 70° S.W. respectively (Map 1). Although the base of the Complex is not exposed on the eastern end of the line of section from consideration of those parts exposed to the north and south it probably too dips west at about 50°. This gives an apparent dip of about 30° in the plane of the cross-section. At the western end of the line of section the basal contact is presumed to dip more steeply at about 70° N.W., sub-parallel to the top of the Complex. The dip of the basal contact is assumed to change gradually between these values.

The great width of outcrop of the Complex may be due to folding or to repetition by faulting.



(a) Shape of the basal contact.
Data used for construction: position and dip of faults, displacement on faults (heavy lines), dip of basal contacts. (see text)



(b) The Red Hill Complex (see text)

Scale
0 5000' 10,000'

Fig. 21

CROSS-SECTION OF THE RED HILL COMPLEX.

Faulting. A number of north-easterly striking faults cut almost at right angles across the line of section. Only vertical movement is assumed to have occurred (p. 215) the amount of which may be estimated from the apparent horizontal displacement of the northern contact of the Complex. Considering faulting alone and assuming that the dips of the basal contact regularly increase in dip from east to west a profile of the basal contact may be deduced as in figure 21a. Faulting it is apparent, will cause considerable tectonic relief of the base of the Complex and for simplicity a hypothetical surface which passes through the centre of relief is defined. This is the median surface and is shown in figure 21a.

Folding. Possible folding of the basal contact of the Complex is suggested by the southward plunging Ben Nevis Anticline and the folded layering within the Complex. The period of such folding (if any) will be much greater than the tectonic relief caused by faulting and will hence be represented by folding of the'median surface'.

Gravity data is of special value in determining the form and the approximate depth of the median surface, as the depth to the base of the Complex may be estimated from isostatic anomalies due to the denser ultramafic rocks. The depth values so obtained will be related to the median surface rather than the true base of the Complex because gravity data tends to smooth out irregularities in relief. Long period folding will therefore be expected to be apparent from a gravity survey.

Gravity Data. A gravity survey covering, in part, the southern extremity of the Red Hill Complex was carried out by Malahoff (1962). The maximum gravity anomally due to ultramafic rocks is 40 mgals. and changes little over the examined area of the Complex but decreases sharply towards its edge. Malahoff interprets the data as indicating an almost horizontal sheet of ultramafic rocks about 7000 feet thick.

The estimation of thickness is based on density contrast of 0.6gms/ cm^3. obtained by assigning a value of 3.3 gms/cm^3. for the ultra-mafic and 2.7 gms/cm^3. for the enclosing rocks. This value of density contrast is probably excessive. Two values for wet densities of Pelorus Group (which surround the Complex on the east and north) are 2.78 and 2.80 gms/cm^3.; the value of 2.7 gms/cm^3. applies only to Greville, Waiua and Torlesse Group rocks. The density contrast should, on this account, be reduced to 0.5 gms/cm^3. Also, the density of ultramafic rocks varies from 2.34 and 2.68 for serpentinites to 3.34 gms/cm^3. for unaltered dunite and therefore a value of 3.3 gms/cm^3. is an absolute upper limit for the average density of the Complex.

It is believed a value of 3.2 gms/cm^3. would be more realistic. The gravity contrast is therefore further reduced to 0.4 gms/cm^3. Consequently the depth to the basal contact of the Complex should be increased to 11,000 feet.

Malahoff did not carry the gravity survey further north than the Plateau, (Map 1) but recently 6 widely separated gravity values to the north have been obtained. The data has been kindly made available by

Dr. T. Hatherton of Geophysics Division, D.S.I.R. who writes, "Isostatic anomaly values range from about -10 to 15 over the Maitai (Strata) on the west side through a maximum value of +22 mgals. in the centre to about -20 mgals. on the east side giving an effect probably due to the Red Hill Complex of about 40 mgals." (pers. comm.). Thus on present evidence Malahoff's interpretation of a flat lying sheet may be extrapolated with reasonably certainty over the whole of the Complex.

Long period folding such as may be expected from extending the Ben Nevis Anticline beneath the Complex is not apparent from the gravity Survey. Rather the form of the base of the Complex approximates to that illustrated in fig. 21a where only faulting is presumed to have occurred. The base of the Complex in the line of section is therefore inferred to be of that form and the median surface is considered to be approximately horizontal. The median surface occurs at a maximum depth of about 11,000 feet beneath the ground surface, but is probably shallower than this beneath the line of section because the base of the Complex outcrops only a short distance to the north. It is assumed in construction of the geological cross-section to occur at a depth of 7,000 feet.

SHAPE OF THE RED HILL COMPLEX

The top of the Complex is exposed along the western contact and in figure 21b has been extrapolated to the east on the assumption that it is sub-parallel to the basal contact of the Complex.

The Red Hill Complex (fig. 21b) is therefore inferred to be a tilted, and faulted sheet-like body of variable thickness, on the west about 3,000 feet, and in the middle of the line of section about 8,000 feet thick. Further to the south where the depth of the median surface is greater, the thickness of the sheet is also likely to be greater - probably about 12,000 feet.

Before faulting the sheet must have been at least 12 miles wide (east-west direction) and 10 miles long. Because its form is similar to that of the sheet that can be traced north for 40 miles to Dun Mountain, the Red Hill Complex is probably the southward extension of that sheet. The break in the sheet north of Mt. Glennie (geological map) is explained as due to Upper Mesozoic and Tertiary faulting.

The original sheet of ultramafics may have extended further north than Dun Mountain. Therefore, although the known limits of the sheet are 40 miles long, 12 miles broad and 3,000 to 12,000 feet thick the original areal extent is likely to have been several times greater. Detailed mapping along the Upper Paleozoic Belts may show how much the short breaks now evident in the geological maps (Sheets 14,16 and 22 of the New Zealand Geological Survey, 1: 250,000 series Beck, 1964; Lensen, 1962; Wood, 1962) can be attributed to post-emplacement faulting.

STRUCTURE AND PETROLOGY OF

THE RED HILL COMPLEX,

NELSON.

PART II.

PETROGRAPHY.

M A F I C   D Y K E S

INTRODUCTION

Cutting the peridotites and gabbros of the Red Hill Complex are a large number of calc-alkali basic dykes which are petrographically of three kinds. The most abundant are medium-grained, dark green, usually mottled rocks composed of brown-green hornblende, labradorite and subordinate quartz, named the hornblende microgabbros. Derived from these are banded leucocratic-melanocratic rocks probably formed by segregation of the felsic and femic components of the hornblende microgabbros. Leucocratic veins composed of andesine and quartz cut the peridotite near some banded dykes and are believed to have formed from filter pressed felsic material derived from the dykes. The third kind, the pyroxene microgabbros, are dark, almost black, fine grained rocks composed of bytownite, augite, hypersthene and commonly some red-brown hornblende.

Dykes are found in all parts of the Complex but are most numerous and of greatest individual thickness in a wedge shaped zone extending diagonally north-west to south-east across the Complex (Map 1). Within the zone the dykes, uniformly composed of hornblende microgabbro or banded rocks, attain a thickness of 60 feet but most are in the range 5 to 25 feet. At least 60 per cent of the total number of dykes occur within the zone which constitutes only about 30 per cent of the area of the Complex. Outside the zone, dykes are generally more basic, composed of pyroxene microgabbro, and are only about 6 inches (fig. 22) to 4 feet thick.

The strike of the dykes is generally uniform over the whole of the Complex and all dip steeply, usually to the north-east.

Fig. 22 Pyroxene microgabbro dyke (covered with lichens) cutting massive harzburgite.

HORNBLENDE MICROGABBROS

M i n e r a l o g y

The hornblende microgabbros are composed of hornblende, plagioclase, quartz, opaque minerals and usually a trace of biotite and apatite.

Hornblende is moderately birefringent and strongly pleochroic from dark green (Z) through brownish-green (Y) to pale yellow (X). The optic axial angle, estimated from curvature of the isogyre, is about 80° (negative). $Z \wedge c$ is about 23°. Some hornblende is rimmed with pale green amphibole, probably actinolite.

Plagioclase is strongly twinned on Carlsbad and albite laws. Combined Carlsbad-albite twins have been used to estimate composition. The large grains are zoned with inner zones of bytownite and outer zones about An_{50}. Low grade metamorphism has affected some rocks and plagioclase is replaced with albite, prehnite and hydrogrossular. Commonly only the calcic inner zones have been altered (fig. 23).

Opaque minerals are black under reflected light and strongly magnetic. Since the opaque mineral in some metamorphosed dykes has altered to sphene and iron oxides, it is probably titaniferous magnetite.

P e t r o g r a p h y

Compositions of the hornblende microgabbros varies within narrow limits. Hornblende and plagioclase together make up 85-90 per cent of the rock; plagioclase between 40 and 60 per cent and hornblende between 35 and 50 per cent. Quartz is a common subordinate constituent. Modal analyses of five specimens from widely separated dykes is given in Table VI.

TABLE VI

MODAL ANALYSES OF HORNBLENDE MICROGABBROS AND COMPOSITION OF FELDSPAR

No.	Position	Plagioclase %	Plagioclase Comp	Hornbl.	Mag.	Quartz	Biotite	Apatite
11006	392770	47	An_{80-48}	41	4	8	tr	tr
11011	368760	51	An_{78-47}	37	3	9	x	tr
11012	361739	45	An_{78-55}	45	5	5	tr	tr
11013	371832	57	An_{80-48}	35	3	5	tr	tr
11014	378856	43	An_{55} groundmass	51	5	tr	tr	x

Texture is usually porphyritic or xenomorphic-granular. The porphyritic rocks contain 2-4mm. phenocrysts of strongly zoned plagioclase constituting not more than 20 per cent of the rock with a groundmass of hornblende and weakly zoned plagioclase grains 0.5 - 0.75mm. in size. The xenomorphic-granular rocks are coarser grained with grain size of plagioclase and hornblende about 3-4mm. Quartz occurs as small interstitial grains and biotite is usually found as very small grains near titaniferous magnetite. A typical example of xenomorphic-granular texture is shown in figure 23. Less commonly, the hornblende microgabbros are hypautomorphic-granular and fine to medium grained. The plagioclase occurs as small (0.75mm.) zoned tabular laths with subhedral crystal form, the hornblende is generally anhedral and has a smaller grain-size (0.5mm.).

The hornblende microgabbros are usually massive, unfoliated rocks but a planar fabric produced by parallel tabular plagioclase laths, and planar segregation of hornblende and quartz and plagioclase is present in some rocks. By increase in the amount of segregation, the dykes grade into the banded rocks discussed below.

Order of crystallisation

Few of the hornblende microgabbro dykes have a "chilled contact" although the marginal rocks of many dykes are strongly altered, presumably by metasomatism related to adjacent serpentinites. However, one dyke near the base of the Complex (at 427764) has a well preserved fine-grained contact and shows several interesting petrographic features. The dyke is one of a series of aligned lensoid masses 70 feet long and

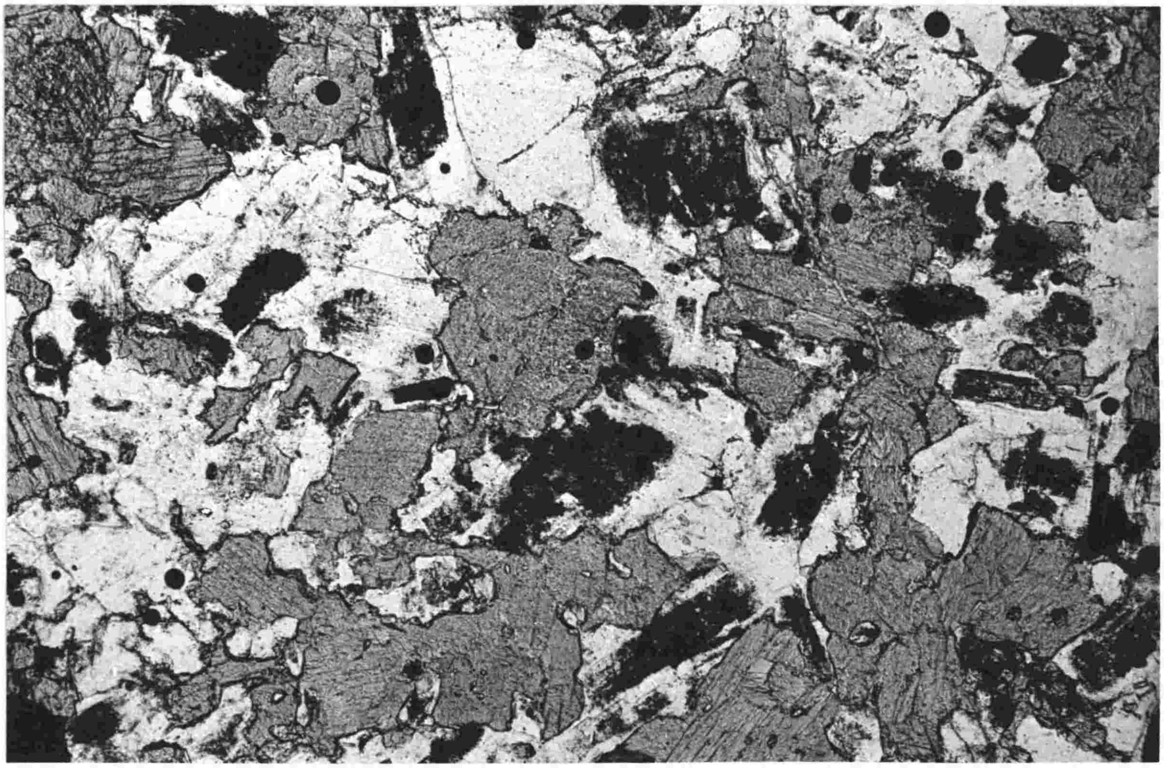
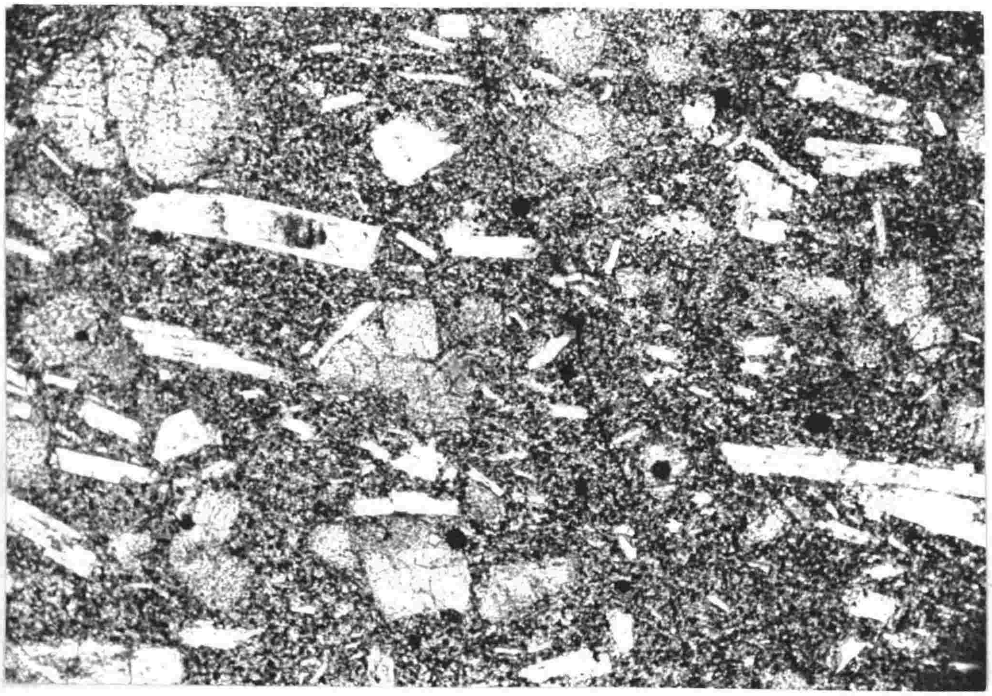


Fig. 23. Xenomorphic-granular texture in hornblende microgabbro. Hornblende (grey), plagioclase (light). Inner zones of plagioclase altered to hydrogrossular (black, turbid material). (20).



Fig, 24. Porphyritic contact rock of hornblende microgabbro. Spec. no. 11003. Plagioclase laths are parallel to dyke walls. Subhedral phenocrysts of pyroxene show as flecked grey grains.

25 feet broad and dips steeply north-east. The surrounding peridotite is partly serpentinised within 20 feet of the intrusion and within a few inches is strongly sheared. At the contact (No. 11003) the intrusive rock is very fine grained, dark coloured and dense. The texture is porphyritic with 2mm. long, euhedral laths of zoned plagioclase (An_{80-58}) and euhedral equidimensional crystals of augite (1mm. in size) equally making up about a quarter of the rock. The groundmass of finely granular pyroxene, magnetite and untwinned plagioclase has a grain size of about 0.01mm. at the contact, increasing to 0.1mm. at 4 cm from the contact. Augite phenocrysts are mantled with hornblende which increases in abundance towards the centre eventually replacing most of the pyroxene. The texture of the contact rock is shown in figure 24 and the pronounced parallelism of plagioclase phenocrysts shown in the photograph is thought to have developed by magmatic flow. Specimens (No. 11004 and 11005) from 2 feet and 12 feet from the contact respectively, contain brown-green hornblende as the only ferromagnesian of the groundmass and only in the phenocrysts does pyroxene exist as small relict grains, surrounded by wide reaction rims of hornblende.

In the centre of the intrusion (No. 11005) the rock is composed of 55 per cent plagioclase, 40 per cent hornblende and accessory quartz, opaque minerals, epidote biotite and apatite. (The presence of epidote is unusual, and found only in this intrusion.) The plagioclase is zoned from An_{65} to An_{50}. The texture is porphyritic with plagioclase phenocrysts constituting 25 per cent of the rock.

While the rocks at the centre of the intrusion are petrographically similar to other hornblende microgabbros, the contact rock has the mineralogy of a dolerite. Evidently during cooling, pyroxene and calcic plagioclase crystallised first and at a lower temperature the pyroxene altered to hornblende. A few other hornblende microgabbros have relict pyroxene but most do not, and hornblende may have crystallised directly from the magma, just as the hornblende in the groundmass of the rocks described above.

BANDED DYKES AND LEUCOCRATIC VEINS

The banded dykes have alternating leucocratic and melanocratic layers or lenticles about 1 to 3 cm. thick parallel to the dyke walls. Veinlets of leucocratic material cut across melanocratic layers, and leucocratic veins up to 10 cm. thick occupy the margin of the dykes and in some places cut into the surrounding (serpentinised) peridotite. Banded dykes have been traced along their strike into massive, unfoliated hornblende microgabbros and some dykes have banded rocks near the margin and hornblende microgabbro in the middle.

Specimen No. 11006 is a typical example of a banded rock near the margin of a 10 feet thick hornblende microgabbro dyke. The leucocratic layers are 2 cm. thick and are composed of 15 per cent fine-grained quartz, 55 per cent fine-grained unzoned plagioclase, 20 per cent weakly zoned plagioclase phenocrysts about 3mm. in diameter and 5 per cent hornblende. The melanocratic layers consist of about 90 per cent hornblende and 10 per cent plagioclase.

The leucocratic veins (e.g. No, 11007) are quartz-rich, composed of about 50-60 per cent fine-grained quartz, 40 per cent phenocrysts of unzoned plagioclase and the remainder hornblende. The plagioclase is An_{55} with little variation in composition in different leucocratic veins throughout the Complex. The texture is protoclastic (fig. 25); the plagioclase phenocrysts are rounded with ragged edges, twin lamellae are commonly bent and the grains show undulatory extinction. The quartz grains have strongly sutured margins and also have undulatory extinction. Because of their close relationship the banded rocks and leucocratic veins were very probably derived from the hornblende microgabbros. The felsic components must have been very mobile as many leucocratic veins are found at least 100 feet from the nearest dyke.

The banded and leucocratic rocks are considered to have developed through deformation accompanying crystallisation of the hornblende microgabbros. The mobile felsic components, the residual fluids of the crystallising dykes, were filter-pressed out into segregation laminae and layers within the dyke, or if very mobile (probably depending on the amount of quartz) as veins along joints or fractures into the surrounding peridotites.

Fig. 25. Leucocratic vein showing plagioclase porphyroclasts in a matrix of quartz. Note directional fabric and undulatory extinction of quartz.

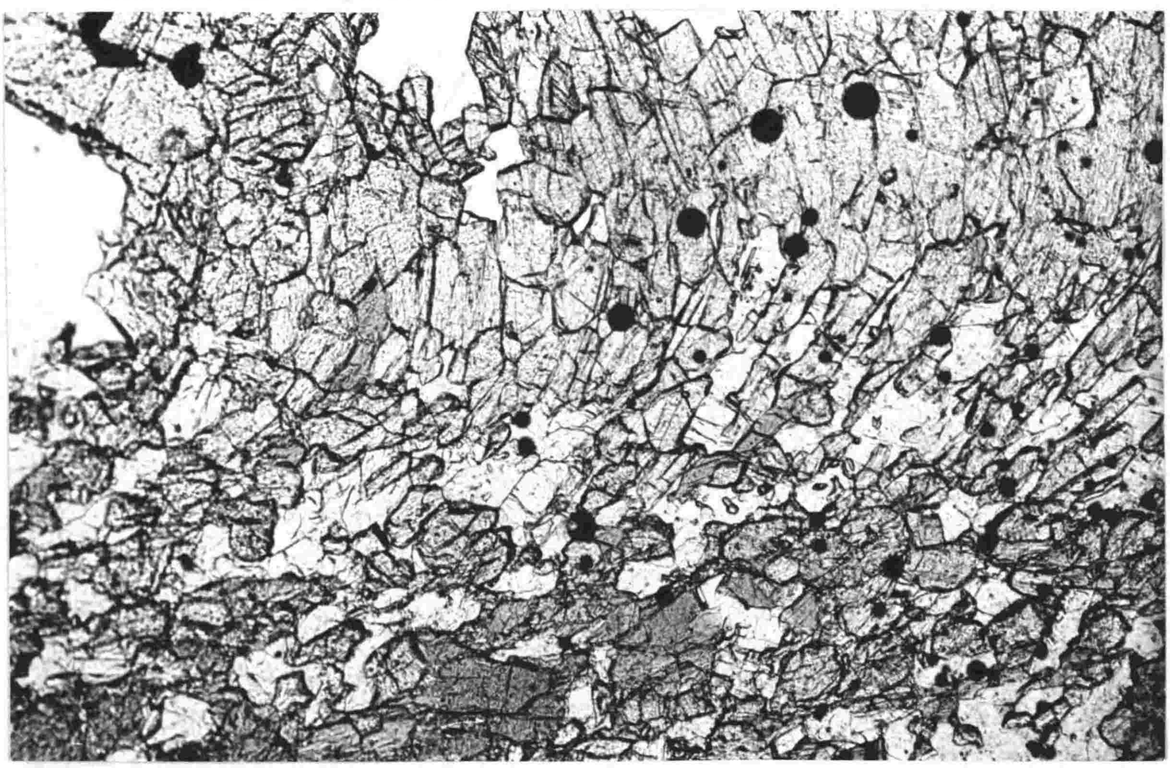


Fig. 26. Pallisade structure at the contact between peridotite (above) and pyroxene microgabbro (below). In this example the pallisade has been bent through about 30°.

PYROXENE MICROGABBROS

M i n e r a l o g y

The pyroxene microgabbros are composed of orthopyroxene, clinopyroxene, amphibole, plagioclase and opaque minerals. The compositions of minerals from three specimens are given in Table VII. The compositions were estimated by optical methods.

Orthopyroxene composition was estimated by measurement of Z refractive index of cleavage fragments by the oil immersion method and, in Specimens Nos. 11010 and 11008, by universal stage (orthoscopic) measurement of 2V. There is good agreement between the values obtained from refractive index and optic angle measurements.

The compositions vary between En_{60} and En_{65} and the mineral is therefore hypersthene. In thin section, the hypersthene is weakly pleochroic, from very pale reddish-brown (X) to very pale green (Z).

Clinopyroxene composition was estimated in Specimen No. 11010. The Y refractive index was measured to a probably accuracy of $\pm$ 002, and 2V to an accuracy of about 1° and is likely to be too low a value rather than too high (Munro, 1963). Using the determination diagram in Deer and Others, 1963b, p. 132, the composition is estimated at $Ca_{39}\ Mg_{36}\ Fe_{25}$. The mineral in thin section is very pale green and non-pleochroic.

Amphibole is strongly pleochroic from straw to rich red-brown. The birefringence is fairly high for a hornblende (estimated at about .025), the Z refractive index moderate (1.672) and the mineral is

TABLE VII

MINERAL COMPOSITION OF THE MICROGABBROS

No.	Position	Mineral	Measured properties	Composition
11008	420800	Feldspar	Carlsbad-albite	An_{75} - An_{68}
		Orthopyroxene	Z = 1.710 2V = 60°	En_{65}
		Amphibole	Z' = 1.672	
11009	430788	Feldspar	Carlsbad-albite	An_{76}
		Orthopyroxene	Z = 1.714	En_{60}
		Amphibole	Z' = 1.672 2V = 82° (-)	
11010	407890	Feldspar	Y = 1.575 Carlsbad-albite	An_{84}- An_{78}
		Orthopyroxene	Z = 1.710 2V = 58°	En_{64}
		Clinopyroxene	Y = 1.700 2V = 47°	$Mg_{39}Ca_{36}Fe_{25}$
		Amphibole	Z' = 1.670	

optically negative with a large 2V (82°). Optical data alone are insufficient to estimate the composition of the amphibole but are consistent with common hornblende in which the 100 $Mg/Mg+Fe^{2}+Fe^{3}+Mn$ ratio is about 60 (Deer, Howie and Zussman, 1963b, p. 296).

Plagioclase composition was estimated from Carlsbad-albite twins. The plagioclase is zoned and in the larger phenocrysts may range in composition from An_{84} to An_{75}. The zoning is not as marked as in the plagioclase of the hornblende microgabbros. The Y refractive indices of several grains from Specimen No. 11010 were measured and the greatest value gave an estimated composition of An_{84}. This is similar to the composition estimated from Carlsbad-albite twins (An_{82}).

Opaque mineral is, like that of the hornblende microgabbro, strongly magnetic, and on low grade metamorphism alters to give sphene. It is therefore likely to be a titaniferous magnetite.

Petrography

TABLE VIII

Modal analyses of microgabbros

No.	Position	Opx.	Opx.	Amphi.	Fels.	Opaque
11008	420800	8	25	19	42	6
11009	430788	20	24	2	45	9
11010	407890	19	30	6	39	5

Composition. Modal analyses of three specimens are given in Table VIII. The rock composition averages at about 40 per cent bytownite, 28 per cent augite, 25 per cent hypersthene plus hornblende and 7 per cent titaniferous magnetite. Hornblende occurs as large poikilitic plates preferentially replacing hypersthene. Consequently the modal proportions of hornblende and hypersthene vary considerably in different thin sections from one specimen.

Texture

The pyroxene microgabbros are porphyritic (Nos. 11008 and 11009) and glomeroporphyritic (No. 11010) with 2-6mm. phenocrysts of zoned plagioclase and 1 to 2mm. phenocrysts of hypersthene and augite altogether totalling not more than 30 per cent of the rock. The groundmass is xenomorphic-granular and consists of fine-grained (0.1mm.) plagioclase, ortho- and clino-pyroxenes, hornblende and titaniferous magnetite.

The hypersthene phenocrysts are usually rimmed by optically continuous augite and augite commonly has a reaction rim of hornblende. Magnetite occurs as inter-granular small grains and as orientated rods within groundmass pyroxene crystals.

Deformation

Like the hornblende microgabbros the pyroxene microgabbros are commonly deformed. Some have narrow shear zones near the margin of the dykes. In these zones the rock has a clastic texture and much of the pyroxene is altered to brown hornblende. The hornblende indicates that deformation occurred when the dyke was still very hot.

Other dykes are strongly sheared and the minerals granulated. The secondary minerals developed are brown-green, and green hornblende, actinolite, clinozoisite, hydrogrossular, prehnite, albite and quartz. The variety of secondary minerals ranging from low temperature facies to high temperature brown hornblende indicates that deformation and recrystallisation occurred locally throughout the period of cooling.

Contacts

The pyroxene microgabbros usually cut unserpentinised peridotites. Immediately adjacent to the olivine and orthopyroxene of the peridotite there is a reaction zone of small (0.05mm.) prismatic crystals or orthopyroxene orientated with their long axes approximately perpendicular to the contact. These form a pallisade-like structure. That illustrated in figure 26 has been bent through about 30° probably by flow of magma during development of the reaction zone. As the orthopyroxene crystals are considerably more magnesian than the hypersthene in the centre of the dyke (2V of the contact crystals is large and positive in sign) they probably developed by reaction between the olivine of the peridotite and high silica content of the dyke.

CHEMICAL ANALYSES

Two new chemical analyses of the mafic dykes are given in Table IX. Analysis I is of a hornblende microgabbro from the Basal Zone of the Complex. The petrography of the specimen is given on page 101. The analysed specimen was taken 2 feet from the contact of a dyke and has suffered comparatively little deuteric alteration apart from the development of hornblende from original augite. Analysis 4 is of a pyroxene microgabbro which is described on pages 106-9.

The chemical constitution of the hornblende microgabbro shows many similarities to that of the average tholeiite from Hawaii (Analysis 3), differing significantly only in higher soda and lower magnesia content. The wide difference in the normative components, such as absence of quartz, appearance of 13 per cent olivine and the much higher albite content of the hornblende microgabbro may be directly attributed to its higher soda content. These differences in the norm and the chemical analyses are minor however, and the rock is essentially tholeiitic in character. In several respects the analysis of the hornblende microgabbro is also very similar to that of a partly metamorphosed volcanic rock (Analysis 2) from the Glennie Formation and quoted by Challis (1965b). Challis's rock however is described by her as an amphibolite (contains volcanic texture with original augite partly replaced by actinolitic-hornblende) and is possibly metasomatised by spilitization and thermal metamorphism. Too much reliance cannot therefore be placed in the similarity and indeed other analyses given by Challis show fewer common features. The volcanic rocks are however tholeiitic in character and are broadly the same magma type as the hornblende microgabbros.

The pyroxene microgabbro also shows some obvious similarities with the tholeiite but contains considerably more lime and less soda. Quartz does not appear in the norm, presumably because of slightly lower silica and the higher lime content. Although no precisely similar rock has been described by McDonald and Katsura (1964) from Hawaii the low alumina and alkali content of the pyroxene microgabbro shows that it too is broadly tholeiitic.

METAMORPHISM

The assemblages hornblende-labradorite-quartz and bytownite-hypersthene-augite-(hornblende) although common in plutonic and metamorphic rocks are most unusual in dykes. If the dykes were intruded into cold peridotite the replacement of augite by hornblende would be expected to be of very limited extent. Yet in many dykes almost complete or total replacement has occurred. This may be explained as due to a long period of deuteric alteration such as may be expected if the dykes were intruded into hot peridotite or, possibly, to a later period of metamorphism.

This point is returned to in discussion of the mode of intrusion of the dykes (p.   ); it is sufficient here to indicate that metamorphism post-dating dyke intrusion has apparently occurred.

P E R I D O T I T E S A N D G A B B R O S

Internally the Red Hill Complex consists of two stratiform zones which may be distinguished in terms of composition, texture and structure of the rocks. In its lower part, the Complex consists of massive (non-layered or-foliated) harzburgite with a strongly developed protoclastic texture whereas in its upper part the rocks are compositionally more varied ranging from dunite to eucrite (gabbro), texturally, xenomorphic-granular and structurally, layered or foliated. Cutting the peridotites are veins and pegmatites, generally of pyroxenite but some of anorthosite. The two zones have been called the Basal and Upper Zones respectively, but only the Basal Zone has been shown on the geological map (Map 1). The boundary between the two zones has been arbitrarily defined at the first appearance of mesoscopic structures such as foliation or layering. This definition is arbitrary to the extent that while layered rocks above the boundary may contain protoclastic texture this texture is characteristic of the Basal Zone.

The strike and dip of layering and foliation, where present, is shown on the geological map (Map 1). In most cases the symbol indicates the orientation of the only planar structure present in the rocks but in some outcrops two cross-cutting planar structures are present. In these few cases the strike and dip of the last formed structure is that shown. Reference may be made to Map II for the orientation of both the early and late structures which occur in the vicinity of Porters Knob.

In the following section the petrography of the rocks of the Red Hill Complex is given. This includes a description not only of the composition and texture of the rocks but also their structure.

COMPOSITION OF THE ROCKS

MINERALOGY

Introduction

The fresh, unaltered rocks of the Red Hill Complex are composed of one or more of the five isomorphous mineral groups, olivines, orthopyroxenes, clinopyroxenes, feldspars and spinels. The primary purpose of the mineralogical study was to investigate the variation in chemical composition of the minerals (with respect to major components) over the whole of the Complex.

Determinative methods. The composition of the silicate minerals was estimated by measurement of refractive indices by the oil immersion method (using sodium vapour light) and in some cases by direct, orthoscopic measurement of 2V on the universal stage. The composition of the spinel was not determined.

Oils spaced at 0.005 were used for approximate determination and for more accurate work oils were mixed to give intermediate values. The refractive indices of the standard oils was determined before and after use on an Abbe Refractometer and the refractive index of mixed oils was measured on a liquid-filled prism refractometer calibrated against one of the standard oils. The measurement of refractive index by this method is considered to be accurate to $\pm$ .002.

Olivine does not have a good cleavage and there is no definite assurance that measurements of the maximum refractive index will give the true value for Z. However, careful measurement of the maximum (Z') and minimum (X') refractive indices of powdered samples gave in most cases a value for Z'-X' close to the true birefringence of olivine and therefore the measured values must lie close to the true values for Z and X. The composition was estimated from the curves given in Deer, Howie and Zussman (1963a, p.22) and the mean value of the two values so obtained is that quoted (Table I[1]). A variation of $\pm$ .002 in refractive index gives a variation in composition of $\pm$ 0.8 mol. per cent.

Orthopyroxene has a perfect {210} cleavage and therefore in crushed samples grains are common with Z parallel to microscope stage and Z can be measured directly. Variation of $\pm$ .002 in refractive index of orthopyroxene gives a variation in composition of $\pm$ 3 mol. per cent. The determinative curves used are those of Deer and others (1963b, p.28).

Clinopyroxene fragments commonly have some grains giving optic axis sections and therefore Y can be measured directly. No attempt has been made to estimate composition of individual clinopyroxene specimens as optic properties may vary considerably due to minor substitutions of Cr_2O_3, Na_2O and Fe_2O_3 for the major components of CaO, MgO and FeO. In general the optic properties are similar to a chrome-diopside analysed by Hess (1949), (Table II).

Feldspar has two very good cleavages and cleavage fragments are common. Bytownite and anorthite have the Y direction almost in (001) and it can be measured directly on (001) fragments. In practice it was found that measurement of maximum and minimum refractive indices gave values fairly close to Z and X. The determinative curves used for estimation of composition are those given by Smith (1957). A variation

TABLE II

MINERAL COMPOSITION IN THE RED HILL COMPLEX

No.	Rock type	Locality	Mineral	Measured properties				Composition
				X	Y	Z	2V	
10955	harzburgite	422762	olivine	1.652	x	1.689	x	Fo_{91}
10956	feldspar vein	417808	feldspar (chemical analyses, Table II)	1.574	1.583	1.589	77.0	$An_{97.2}$
10957	harzburgite	387738	olivine		x	1.688	x	Fo_{91}
10958	harzburgite	"	"		x	1.688	x	Fo_{91}
10959	"	"	"		x	1.686	x	Fo_{92}
10960	"	"	"		x	1.689	x	Fo_{91}
10961	"	"	"	1.648	x	1.686	x	Fo_{92}
10962	felspathic-harzburgite	414886	olivine	1.651	x	1.689	88 (+)	Fo_{91}
			Opx *	x	x	1.693	87 (+)	En_{88}
			feldspar	1.575	1.582	1.587	x	An_{98}
			Cpx	x	1.685	x	58	
10963	eucrite	348828	olivine	1.657	x	1.695	x	Fo_{88}
			Opx	x	x	1.686	x	En_{82}
			feldspar	1.574	x	1.586	x	An_{95}
			Cpx	x	1.680	x	x	

10964	eucrite	380854	olivine	1.657	x	1.695	x	Fo_{88}
			Opx			1.		
			feldspar	1.572	1.580	1.586	x	An_{92}
			Cpx		1.680		57	
10965	harzburgite	440862	olivine	1.653	x	1.691	x	Fo_{90}
			Opx	x	x	1.676	x	En_{90}
10966	harzburgite	373882	olivine	1.653	x	1.689	x	Fo_{90}
			Opx	x	x	1.678	x	En_{88}
10967	harzburgite	436851	olivine	1.653	x	1.691	x	Fo_{90}
			Opx	x	x	1.676	x	En_{90}
10968	olivine-eucrite	385815	olivine	1.655	x	1.692	x	Fo_{89}
			Opx	x	x	1.679	x	En_{87}
			feldspar	1.574	1.580	1.586	x	An_{94}
			Cpx	x	1.679	x	x	
Hess (1949)	bronzite	Bushveld	Cpx (Cr_2O_3 1.31%)	1.674	1.680	1.702	$55\frac{1}{2}$	Mg_{50} Fe_5 Ca_{45}

(Cpx = clinopyroxene, Opx = orthopyroxene, x = no measurement) of $\pm$ .002 gives a variation in composition of $\pm$ 3.6 mol. per cent. Using the mean of the three independent measurements of X, Y and Z the probable error of composition is about $\pm$ 2 mol. per cent. One feldspar, estimated by the above method to be composed of 98 per cent anorthite, was chemically analysed (after optical measurements were made) and gave a composition of 97.2 per cent anorthite.

Accuracy of optical methods in determining composition.

Composition determination by optical methods is open to errors other than those of measurement, e.g. considerable changes in refractive index of olivine may be caused by small amounts of Mn, Ti, or Fe^{3+}, replacing Mg or Fe^{2+}. Sometimes this may cause as much as .005 change from the 'normal value' giving an error of 4 mol. per cent. Despite potential errors of this sort, Deer and others (op. cit.) consider that more precise determination can be obtained from refractive index than measurement of 2V. Wyllie (1958) stated that a single conoscopic measurement of 2V on the universal stage may be in error by as much as 3°, corresponding to 6 mol. per cent. Munro (1963) showed that orthoscopic measurement of 2V using equipment similar to that available to the writer, usually gave results at least 2° smaller than the true value, for large angles at 2V and for conoscopic measurement, between 0.5 and 1.0° smaller. The values of 2V quoted in Table II are therefore likely to be about 2° lower than true value for larger angles, and about 1° lower for intermediate angles.

Variation in composition

The estimated composition of minerals from a number of different localities and rock types is given in Table II. Specimens Nos. 10957-61 inclusive were collected at 50 foot intervals at right angles to a steeply-dipping layering but others are widely separated throughout the Upper and Basal Zones of the Complex. Apart from the feldspar of specimen No. 10956, all determinations were made of the 'country rock' as opposed to veins. The following general conclusions seem justified.

Mg/Fe ratio of olivine and orthopyroxene. (a) There is no significant variation in the Mg/Fe ratio in the minerals from rocks of the same type from different parts of the Complex. Hence there is no evidence of 'cryptic layering'. (b) The limits of variation in Mg/Fe ratio are small; olivine ranges from Fo_{93} to Fo_{88} and orthopyroxene from En_{90} to En_{82}. The average Mg/Fe ratio of the femic minerals should however be weighted in favour of the more magnesian members as the iron rich minerals occur in the eucrites which are of small amount in comparison to the other rocks of the Complex. The average values are therefore estimated as Fo_{91} for olivine and En_{88} for orthopyroxene.

Clinopyroxene. The range in measured optic properties of clinopyroxene is small suggesting only a small variation in chemical composition. Also all specimens from 'country rock' are green in handspecimen and are usually tinted green in thin section. Thus the clinopyroxene is likely to be similar to the chrome-diopside described by Hess (1949), (Table I[1]).

Feldspar. Calcium-rich plagioclases occur in feldspathic-peridotites low in the Upper Zone and the more sodic (although still anorthite in composition) in the eucrites, high in the Upper Zone. This distribution may be interpreted as indicating 'cryptic layering' but as there is no parallel variation in the femic minerals of the adjacent rocks this interpretation is not considered to be correct. The average composition of feldspar in the Complex should be weighted in favour of the more calcic members as these are more abundant. Accordingly it is estimated that the average composition is An_{96}.

ROCK TYPES

Introduction

Nomenclature. The nomenclature used for rocks composed of olivine, orthopyroxene and/or clinopyroxene has been adapted from the classification in Williams, Turner and Gilbert (1954) and is summarised in the triangular diagram of figure 28. Subordinate amounts (less than 10 per cent) of a mineral are indicated by adjectives, e.g. diopsidic-harzburgite. Spinel however, is a constant accessory in the peridotites and is not normally referred to.

Rocks containing subordinate feldspar are distinguished as 'feldspathic'. Rocks containing essential feldspar (greater than 10 per cent) and clinopyroxene are termed eucrites (the feldspar is invariably very calcic). The eucrites generally contain some olivine and orthopyroxene and those containing a higher proportion of olivine than feldspar are termed olivine-eucrites.

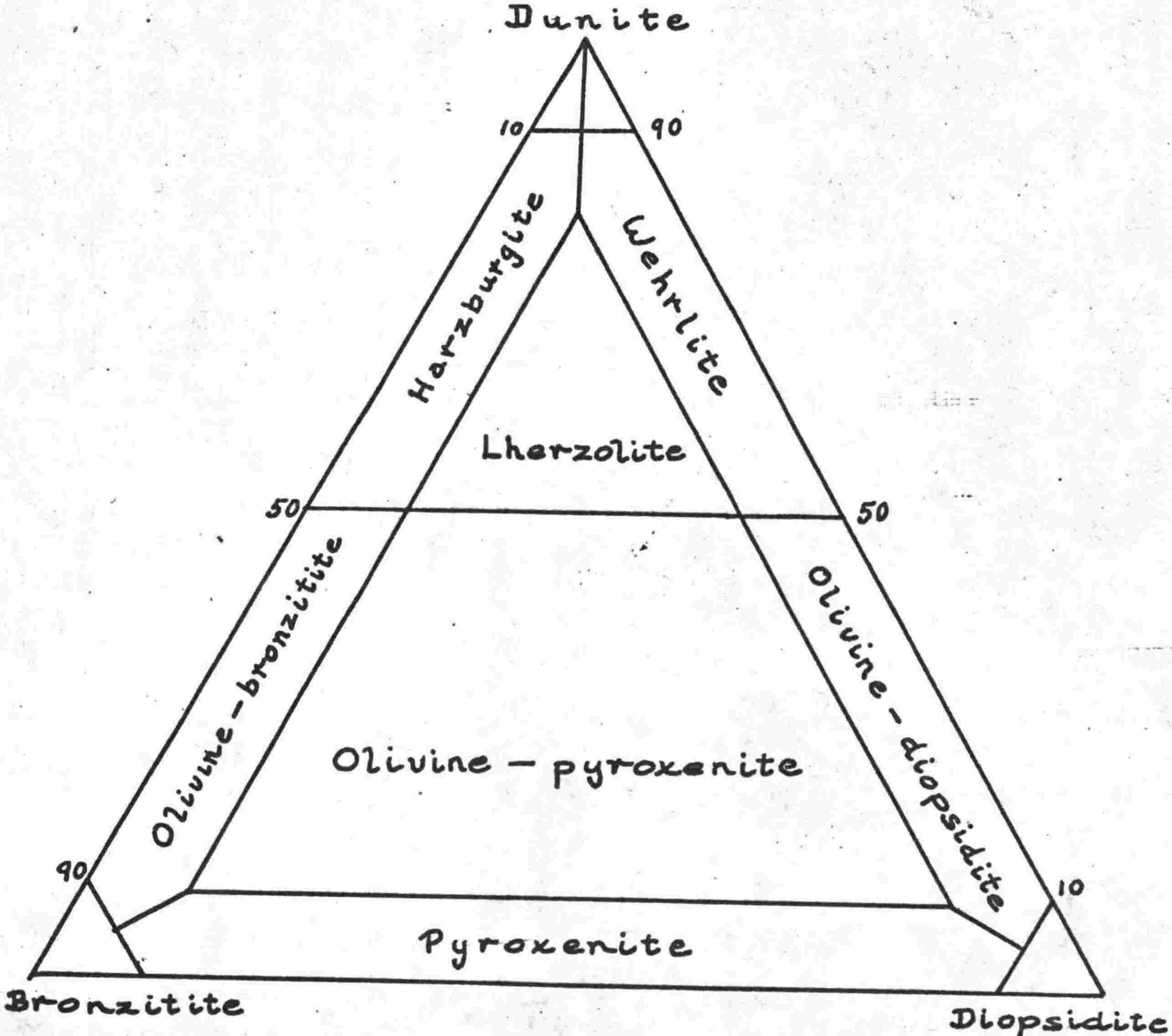


Fig. 28. Classification of pyroxene bearing peridotites.

TABLE III. MODAL ANALYSES

No.	Rock type	Map ref.	Zone	Olivine	Opx	Cpx	Fels	Spinel
10955	harzburgite	422762	B	72	26	tr	x	2
10969	chromite-dunite	350755	U	92	x	x	x	8
10970	harzburgite	387738	I	86	12	tr	x	2
10957	harzburgite	"	I	74	23	2	x	1
10971	diopsidic-harzburgite	412885	I	77	13	8	x	2
10972	diopsidic-harzburgite	"	I	59	34	5	x	2
10973	diopsidic-harzburgite	368752	U	71	23	4	x	2
10974	feldspathic-diopsidic-harzburgite	415882	I	71	20	4	3	2
10975	feldspathic-diopsidic-harzburgite	388845	U	69	13	7	9	2
10976	harzburgite	414897	B	69	24	3	x	4
10977	harzburgite	355735	U	69	25	4	x	2
10978	enstatite-dunite	382857	U	91	7	x	x	2
10964	eucrite	380854	U	8	11	38	53	x
10979	olivine-eucrite	382857	U	44	6	15	34	1
10980	eucrite	346826	U	14	7	21	58	x
10965	harzburgite	440862	B	75	18	5	x	2
10981	olivine-diopsidite	372881	U	29	6	64	x	1
10967	harzburgite	436851	B-I	71	27	tr	x	2
10968	olivine-eucrite	385815	U	41	tr	37	20	2

B = Basal Zone I = Lower part of Upper Zone U = Upper part of Upper Zone tr = trace

Rock identification. The major rock types can be easily distinguished in the field. Olivine weathers preferentially from peridotite and leaves the associated minerals in relief. Careful examination of the weathered surface therefore enables identification and estimation of the relative proportion of the rock forming minerals.

Modal analyses of a number of rock types is given in Table III, and are believed to be representative of the Complex. The average grain size is about 2mm. and thin sections covered at least 8, and usually, 10 square centimetres. All modal analyses were made with between 1500 and 2000 points.

Basal Zone

The Basal Zone is composed entirely of harzburgite (usually containing subordinate diopside) and rare pyroxenite veins. Examination of the field shows that there is little change in composition throughout the Zone. The average composition of the Zone is therefore given reasonably accurately by the average of the modal analyses of Basal Zone rocks (Table III): 72 per cent olivine; 23 per cent orthopyroxene; 3 per cent clinopyroxene and 2 per cent spinel.

Upper Zone

The lower part of the Upper Zone is also composed largely of harzburgite but feldspathic peridotite and dunite are common. The feldspathic peridotite invariably contains a conspicuous foliation (p.150) caused by parallel orientation of feldspar aggregates. Even when the rock consists of less than 5 per cent feldspar, the foliation is well developed and easily observed. The feldspathic-peridotite occurs only in pockets a few feet in diameter within non-feldspathic harzburgite.

The relative proportion of feldspathic to non-feldspathic harzburgite in the lower part of the Upper Zone is estimated at about one to twenty.

Large irregular-shaped masses of coarse-grained dunite and rare, widely separated veins or layers of medium to coarse-grained dunite occur in the lower part of the Upper Zone. The large masses are roughly lenticular bodies measuring about 300 feet long by 100 feet wide. The contacts of the dunite with enclosing harzburgite is sharp on the scale of the map but diffuse over a distance of 5 feet or so. Knife-edge contacts between harzburgite and dunite have not been observed.

Towards the upper part of the Upper Zone dunite and pyroxenite veins increase in abundance and rock types become more varied. Wehrlite, lherzolite, eucrite and pyroxenite have all been observed but the most common rock types are still harzburgite and dunite. Feldspathic-peridotite is not as common as in the lower part of the Upper Zone.

A typical section through the upper part of the Upper Zone is given below. The section is exposed on a ridge between 364855 and 380850. Layering, lamination and the attitude of the rock units are all parallel and the only cross-cutting structures are coarse-grained pyroxenite pegmatites.

Contact serpentinites.

100 ft.	Dunite
15 ft.	Olivine-diopsidite (80-90 per cent pyroxene)
60 ft	Harzburgite (25 per cent pyroxene) and dunite layered each 2 inches thick
100 ft.	Dunite (cut by 3 inch-thick veins of pyroxenite)

15 ft.	Olivine-pyroxenite (70 per cent pyroxene) laminated.
30 ft.	Harzburgite (20 per cent pyroxene)
2 ft.	Eucrite
60 ft.	Dunite (a few veins of pyroxenite)
150 ft.	Harzburgite (20-40 per cent pyroxene) and dunite layered 3 inches and 2 inches thick respectively
80 ft.	Dunite
4 ft.	Eucrite
10 ft.	Harzburgite (30 per cent pyroxene)
2 ft.	Eucrite
400 ft.	Harzburgite-dunite, layered 6 inches to 2 inches. Cut by pyroxenite veins
20 ft	Lherzolite (40 per cent pyroxene)
50 ft.	Dunite (pyroxenite veins up to 3 inches thick)
15 ft.	Wehrlite (30 per cent pyroxene) laminated
100 ft.	Harzburgite (20 per cent pyroxene)
60 ft.	Dunite.

(Note the abrupt changes in rock types, e.g. from harzburgite to eucrite to dunite. Thayer (1960) considers abrupt variation in rock types to be a typical feature of Alpine-type peridotites.)

It is very difficult to estimate the average composition of the Upper Zone because of the small and large scale layering. However, some measure of the relative proportions of the minerals can be made using the composition of the Basal Zones as a basis for comparison. The large proportion of dunite, suggesting comparative olivine-enrichment, is offset by the higher proportion of orthopyroxene in the pyroxene-rich rock units and the large number of pyroxenite veins. Overall, the relative proportion of olivine to orthopyroxene is likely to be about the same as in the Basal Zone. Clinopyroxene also appears to be more common but this may be misleading, as clinopyroxene is widely distributed throughout the Basal Zone and in the described section it is concentrated in layers or veins and is therefore more conspicuous. The average content of clinopyroxene is probably between 3 and 7 per cent. Feldspar occurs almost wholly within the eucrites and constitutes about 40 per cent of the rock (Table III.) As there is a total of 8 feet of eucrite in the 1600 foot thick section, the proportion of feldspar is of the order of 0.2 per cent.

The section is regarded as typical of the upper part of the Upper Zone, the average composition of which is therefore estimated as 70 per cent olivine, 22 per cent orthopyroxene, 5 per cent clinopyroxene, 2 per cent spinal and 0.2 per cent feldspar.

TABLE IV

CHEMICAL ANALYSES

	1	2	3
Spec. No.	10967	10974	10979
SiO_2	44.1	43.1	43.2
TiO_2	-	-	
Al_2O_3	0.7	1.5	12.2
Fe_2O_3	0.4	1.4	0.8
FeO	7.5	7.4	5.1
MnO	0.13	0.11	0.1
MgO	44.4	42.8	25.6
CaO	0.7	1.8	9.4
Na_2O	0.1	0.1	1.1
K_2O	0.1	0.1	0.1
H_2O^-	0.05	0.15	0.1
H_2O^+	0.1	0.8	1.75
P_2O_5	0.03	0.05	0.01
Cr	4000ppm	3000ppm	3500ppm
Ni	4500 "	4000 "	4000 "
Co	100 "	150 "	100 "
V	50 "	70 "	70 "
Ti	30 "	100 "	350 "
Mo	2 "	2 "	2 "
Cu	20 "	50 "	150 "
Total	99.47	100.16	100.38

Analysis 1. Protoclastic harzburgite representative of Basal Zone. Map reference 436851

Analysis 2. Feldspathic harzburgite from near the base of Upper Zone. Map reference 410886

Analysis 3. Olivine-eucrite from near top of Upper Zone. Map reference 374857

Modal analyses of specimen given in Table III.

Chemical analyses by J.A. Ritchie

Spectrographic analyses by W.C. Tennant.

TABLE V.

CHEMICAL ANALYSES OF ANORTHITE

Vein cutting harzburgite near Mt. Chrome. Map Ref. 417808.

Specimen No. 10956

	Weight per cent	Molecular Proportion	Number of ions ((32(O)).	
SiO_2	43.2	0.7192	Si	7.996
TiO_2	----			
Al_2O_3	36.3	0.3551	Al	7.939
Fe_2O_3	0.3	0.0018	Fe^{3+}	0.026
FeO	0.15	0.0021	Fe^{2+}	0.024
MgO	less than 0.1			
CaO	20.1	0.3584	Ca	3.996
Na_2O	0.35	0.0056	Na	0.106
K_2O	----			
P_2O_5	0.00			
H_2O+	0.5			
H_2O-	0.1			
Total	101.00			

Cr	10ppm	Mn	70ppm
Ni	5ppm	Ti	30ppm
Co	10ppm	Mo	2ppm
V	5ppm	Cu	10ppm

Molecular percentage Anorthite : Albite 97.2 : 2.8

Optical Properties.

N_x = 1.574

N_y = 1.583

N_z = 1.589

2V = $77.0 \pm 1^\circ$

CHEMICAL ANALYSES

R o c k a n a l y s e s

Three new chemical analyses of rocks from the Red Hill Complex are given in Table IV. The first is of a lineated protoclastic harzburgite which although by definition is included in the Upper Zone of the Complex, occurs near the boundary with the Basal Zone, is petrographically similar to Basal Zone rocks (c.f. Tables II and III), and is therefore regarded as representative of the Basal Zone. The second, a feldspathic-peridotite, is very similar in chemical composition, but contains slightly more aluminium and ferric iron. The third is an olivine eucrite from the upper part of the Upper Zone and although not an abundant rock type, it is considered to be petrologically significant.

Comparison of the trace elements Cr. Ni, Co. V, Tc, Mo and Cu of the three specimens clearly indicate the close genetic relationship of the eucrite with the more abundant harzburgite and feldspathic peridotites.

M i n e r a l a n a l y s e s

A new analysis of feldspar, from a vein cutting harzburgite near Chrome, is given in Table V. The computed composition, $An_{97.2}$, indicates it to be a basic anorthite similar to a feldspar recorded by Tsuboi (1935) from an olivine-eucrite nodule enclosed in tuff. The determination of the feldspar composition from optical properties (given in Tables II and V) and the determinative curves of Smith (1957) gives results in very close agreement with the chemical analysis.

Challis (1965a) showed that the clinopyroxene of the ultramafic rocks from the Red Hill Complex have lower Al_2O_3, Na_2O and TiO_2 content than clinopyroxenes of peridotite nodules in volcanic rocks. In this respect the clinopyroxenes of the Red Hill Complex are similar to those of Dun Mountain (refer to analyses given by Ross, Foster and Myers, 1954).

T E X T U R E

DESCRIPTION

Rock textures of the Red Hill Complex can be divided into two groups, clastic and xenomorphic-granular. Clastic textures may be described, in a broad sense, as in Williams, Turner and Gilbert (1954, p. 25) as rock textures in which the constituent minerals appear to be fractured. Xenomorphic-granular textures refer to non-clastic textures in which almost all of the constituent minerals are anhedral.

Clastic textures

Protoclastic. Rocks of the Basal Zone and some in the lower part of the Upper Zone have a strongly developed clastic texture. Because this texture is inferred to have formed at high temperatures (during emplacement of the Complex) it is referred to as protoclastic. A typical example of the texture (specimen No. 10976) is illustrated in figure 29. Fragments range in size from 6mm. down below resolution of the microscope. The matrix (grains less than 2 mm. across), composed of the same minerals as the porphyroclasts (larger grains), forms a dense mosaic of grains commonly showing little evidence of deformation. The porphyroclasts on the other hand, are considerably deformed and occur as rounded grains with highly sutured margins (fig. 30). Several of the porphyroclasts are split into one or two slightly separated parts. Along the fractures is a fine-grained mosaic of the matrix. Bent lamellae of orthopyroxene (fig. 31), very strongly

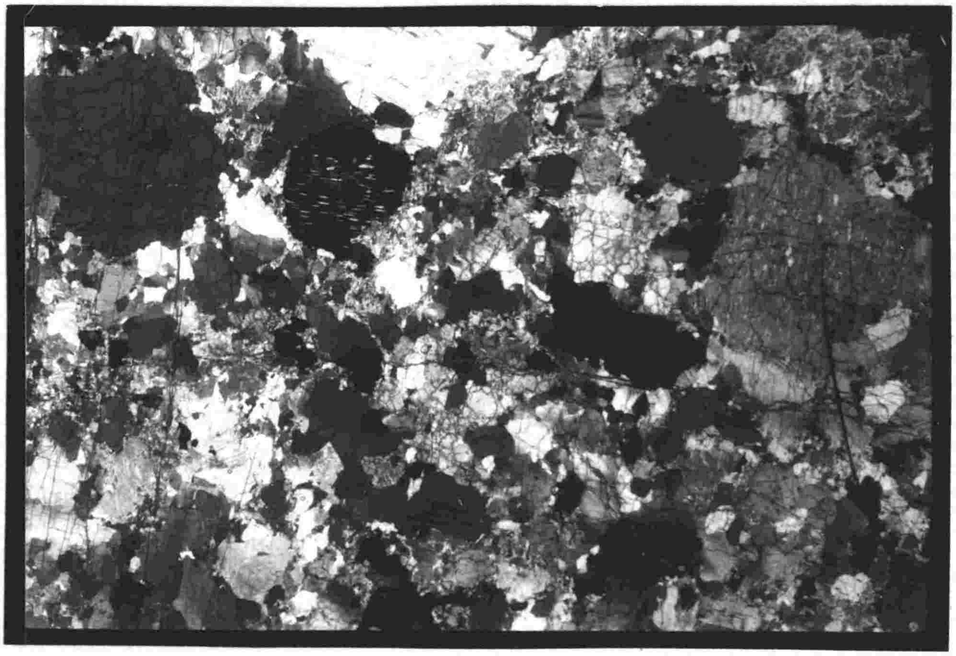


Fig. 29. Protoclastic harzburgite. Rounded porphyroclasts of olivine and orthopyroxene in a granulated matrix of the same composition. Spec. No. 10976 (X 8).

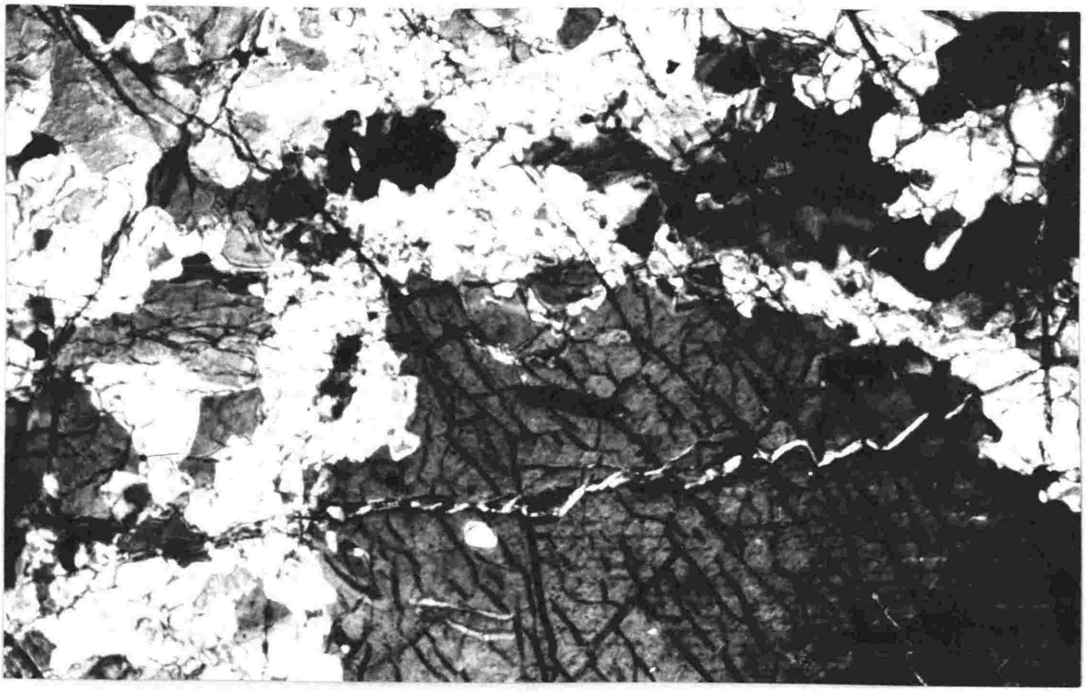


Fig. 30. Margin of orthopyroxene porphyroclast in protoclastic harzburgite. (X 50).

Fig. 31. Bent lamellae of orthopyroxene grains in protoclastic harzburgite. (× 70).

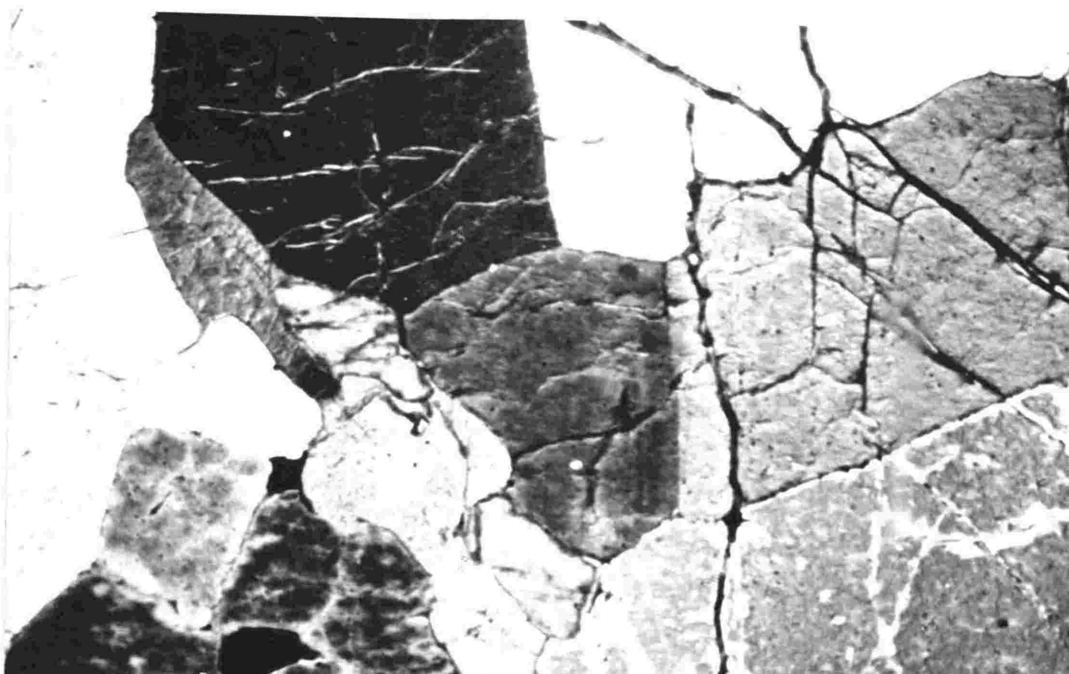


Fig. 32. Xenomorphic-granular texture in dunite. (×20)

developed translation bands with ruptured boundaries, and the fractured appearance of the larger grains indicate that the texture is deformational in origin.

Original protoclastic texture in some rocks has been considerably modified by post-deformational recrystallisation (see below) and in some cases has been completely obliterated by development of replacement veins of dunite. For instance, at 412885, veins of dunite cut at varying angles across country rock which has a well-developed protoclastic texture. The dunite (No. 10982) has a xenomorphic-equigranular texture (fig. 32) while the harzburgite within 4 inches of the vein (No. 10983) is distinctly protoclastic. Protoclastic textures also occur in other specimens of the country rock and are of regional extent. The dunite veins, presumably formed during cooling of the ultramafic rocks, are clearly younger than the protoclastic texture which therefore, must have developed when the rocks were still very hot. The most likely period was during or before emplacement of the Complex, when the rocks near the base of the ultramafic sheet must have been considerably deformed.

Cataclastic texture. Ragan (1963) who described very clear examples of protoclastic texture in the Twin Sisters Dunite referred to that texture as 'cataclastic'. It is preferred here to distinguish deformational clastic textures formed at high temperatures (protoclastic) from those formed at low temperatures. The latter are referred to as cataclastic, the same meaning as given in Williams, Turner and Gilbert (1954).

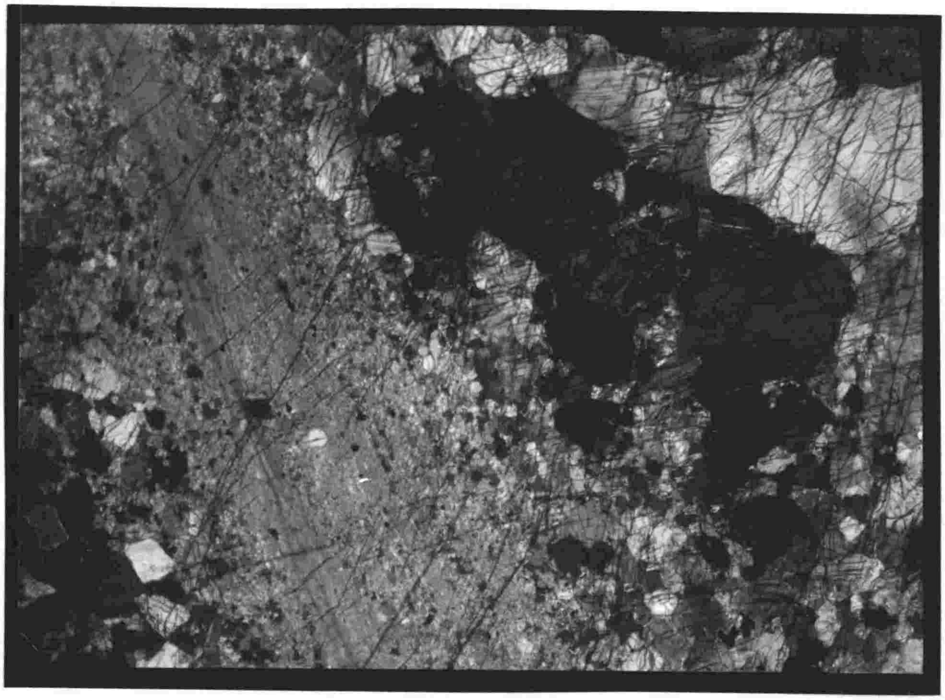


Fig. 33. Cataclastic texture developed in a shear zone traversing peridotite. Specimen No. 10985
(X 5)

Cataclastic textures are found locally developed in rocks of both the Upper and Basal Zones. They have formed along shear fractures in the ultramafic rocks and are seldom more than a few inches wide. Although wider zones of cataclasis may have developed, these probably permitted sufficient water to enter and completely serpentinise the peridotite and thus to obliterate the cataclastic texture.

A good example of cataclastic texture is shown by specimen No. 10985 an olivine pyroxenite from 396874. A shear zone about 1cm. wide composed of very finely-crushed material cuts across the rock (fig. 33). The shear zone grades laterally through increasingly large breccia fragments into the normal xenomorphic-granular texture of the country rock. In hand specimen the shear zone is observed only as a linear, weathered indentation in the pyroxenite.

The low temperature origin of cataclastic texture is suggested by its only local development and the absence of post deformational recrystallization.

Xenomorphic - granular textures

Most rocks of the Upper Zone, all veins, and layered rocks have a xenomorphic-granular texture. Most polymineralic rocks are inequigranular with individual grains ranging from 0.2mm. up to about 4mm. (figs. 34, 35), but such monomineralic rocks as dunite generally have an equigranular texture (fig. 32). (Note the possible euhedral grain of olivine in fig. 32. The photomicrograph (Spec. No. 10996) is of dunite in a foliated and layered feldspathic harzburgite-dunite sequence. The significance of euhedra are referred to on page 146.)

Fig. 34. Xenomorphic-granular texture in olivine-eucrite. (Specimen No.10979) (X 5)

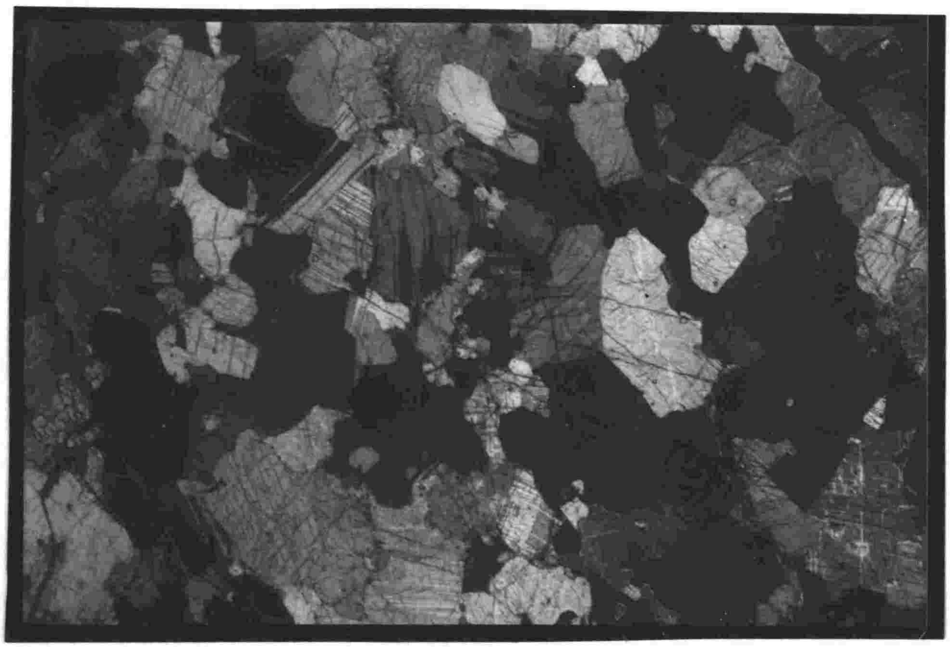


Fig.35. Xenomorphic-granular texture in eucrite. (Specimen No. 10964). (X 15).

Xenomorphic-granular rocks in the lower part of the Upper Zone frequently contain highly irregular grains (fig. 36) a texture referred to as irregular xenomorphic-granular.

<u>O t h e r   t y p e s   o f   r o c k   t e x t u r e</u>

Although the widespread formation of dunite by replacement of prexisting peridotite (and vice versa) is demonstrable (p.238) textural evidence of replacement is very rare. Apparently the replacement of olivine by orthopyroxene tended to form a large number of small discrete individual grains. This can be explained as due to later recrystallisation. If, as is to be expected, replacement caused the partial conversion of olivine grains to orthopyroxene then later recrystallisation resulted in regrowth of the partly-replaced grain as two separate grains, one of orthopyroxene, the other olivine.

Clear evidence of partial replacement of olivine by orthopyroxene is shown in one instance, however (fig. 40). A single large (6mm.) orthypyroxene grain encloses a ragged, partly absorbed grain of olivine. The olivine 'inclusions' have a single extinction position indicating optical continuity. Note the tendency for many of the olivine "inclusions" to have a long axis parallel to the orthopyroxene c-axis. This perhaps illustrates preferential replacement. Note also that the exsolution lamellae (probably of CaO-rich pyroxene) only occur near the centre of the orthopyroxene grain. As the CaO content of orthopyroxene is a function of temperature (Atlas, 1952) and is exsolved during slow cooling, it may be concluded that the outer rims of the grain (lamellae absent) developed at a lower temperature than the centre. Brown (1957 p.528) has suggested that augite blebs in the centre of orthopyroxenes of the Skaergaard Complex are due to inversion of monoclinic to orthorhombic pyroxene beginning first in the centre of the grain with subsequent material exsolved from the rim added to the centre. The example is however not wholly analagous for though blebs of augite are restricted to the centre, fine augite lamellae similar to those featured here persist to the outermost margin of the grain.

RECRYSTALLISATION

In addition to local recrystallisation that occurs in replacement veins (p.135) a gradual recrystallisation process, modifying original protoclastic texture can be identified in the rocks of the Basal and lower part of the Upper Zones.

The first stage in the process is shown by the development of irregular apophyses on the margins of porphyroclasts in protoclastic harzburgites of the Basal Zone (fig. 30). These apophyses, surrounded by the matrix, are unlikely to have persisted during protoclasis and have probably grown by post-deformational recrystallisation; the larger grains assimilating the smaller, relatively unstable grains of the matrix.

As assimilation continued the larger grains began to interfere and develop smooth unsutured crystal boundaries (figs. 37, 38). Remnants of the matrix of original protoclastic texture may still be observed interstitial to the larger grains (fig. 38). The final stage in the process was complete obliteration of the protoclastic texture and the development of xenomorphic-granular texture.

The irregular xenomorphic-granular texture described above is readily explicable in terms of the recrystallisation process. It is possible that late in the recrystallisation process some grains assimilating the interstitial finely-granulated material were constrained by adjacent porphyroclasts, and the grains developed long apophyses penetrating along the crystal boundaries with growth only completed when all the matrix had been absorbed.

Most specimens from the lower part of the Upper Zone show considerable modification of original protoclastic texture while specimens from the Basal Zone generally show only weak, if any, evidence of recrystallisation. It is therefore probable that the degree of post-deformational recrystallisation increases upwards in the Complex. Perhaps protoclastic texture was developed in all rocks at one time but due to recrystallisation it is now preserved only in the rocks near the base of the Complex.

Ragan (1963) has given several specific examples of recrystallisation in the twin Sisters Dunite - such as a thin unstrained mosaic zone cutting a single strained crystal. Similar features occur throughout the Red Hill Complex. In one example a train of small grains of

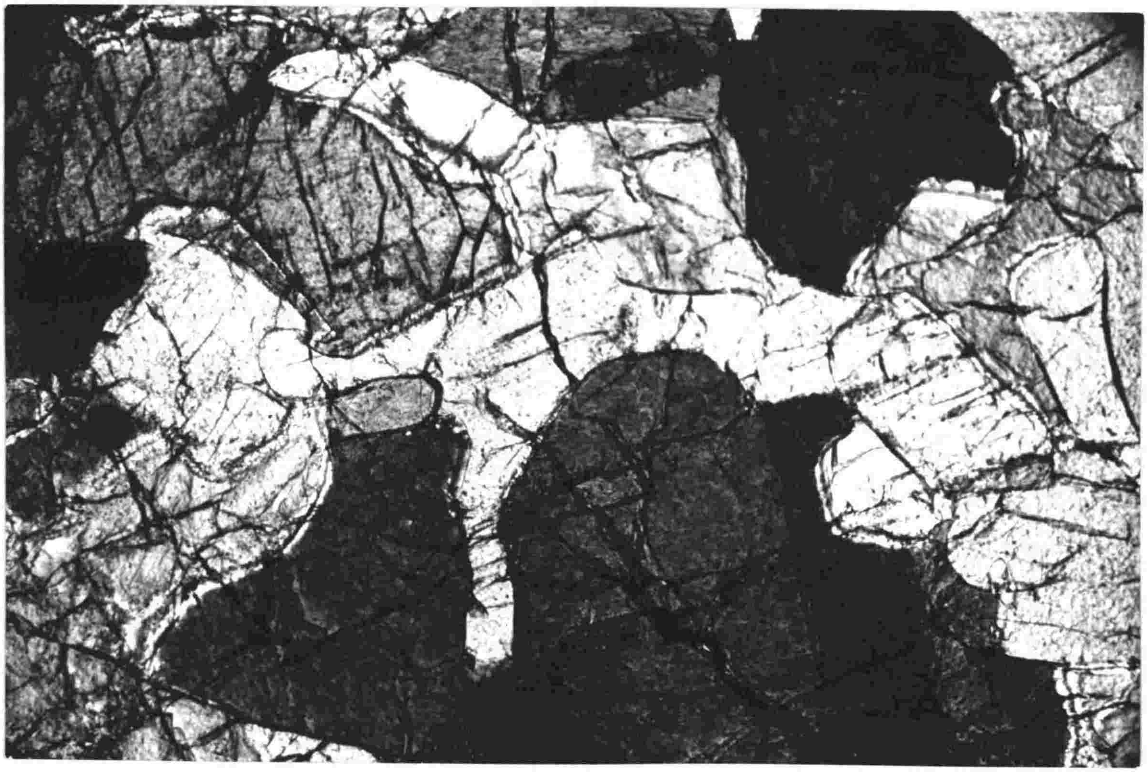


Fig. 36. Irregular xenomorphic-granular texture. Long thin apophyses from an olivine grain penetrate along crystal boundaries and into adjacent grains.
Spec. No.10974 (X 50).

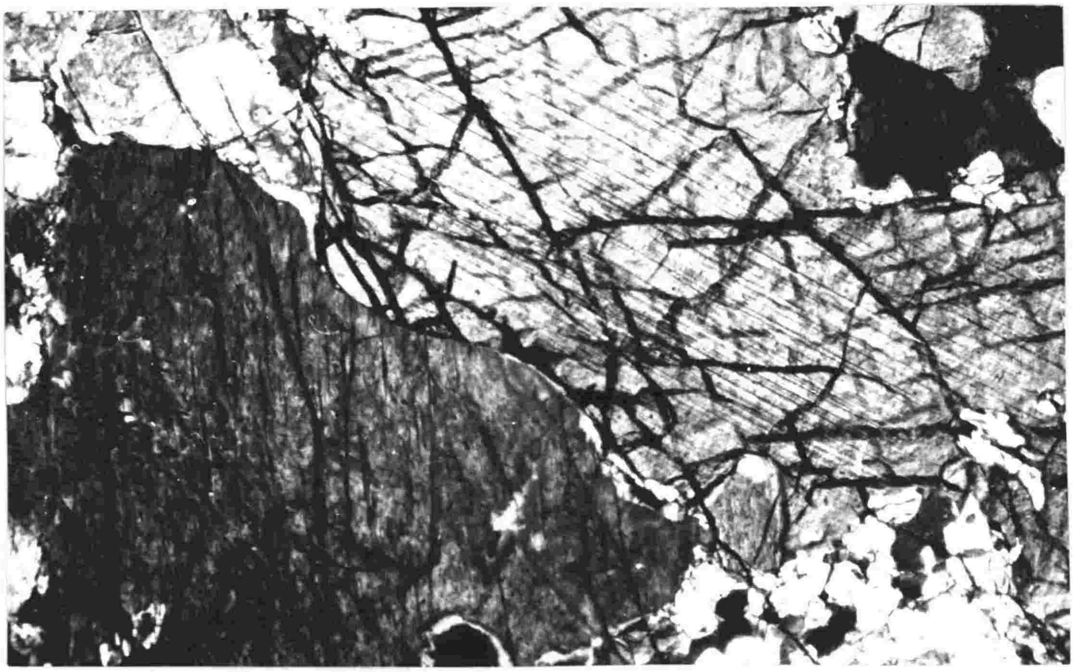


Fig. 37. Smooth unsutured boundaries between orthopyroxene grains in protoclastic harzburgite partly modified by recrystallisation. (X 50).

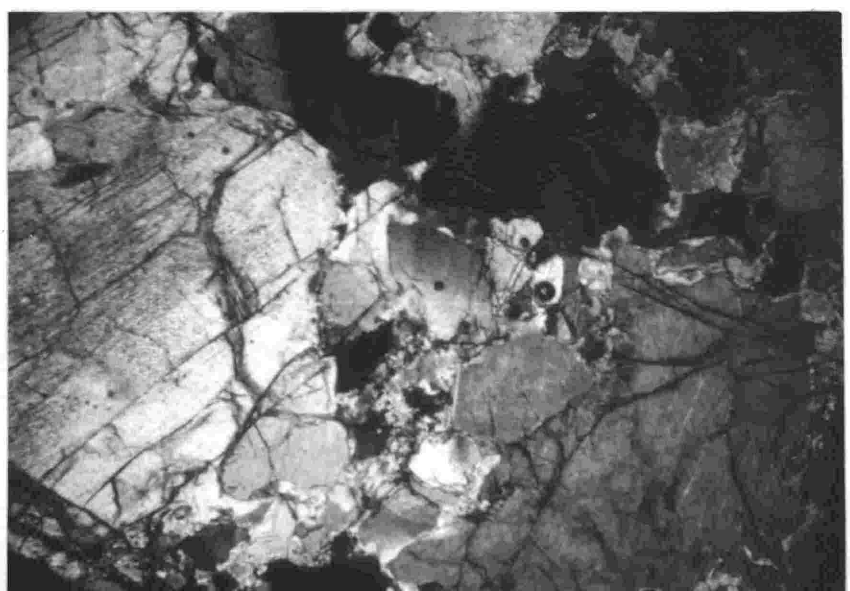


Fig. 38. Interstitial remnants of the matrix of original protoclastic texture. (X 50).

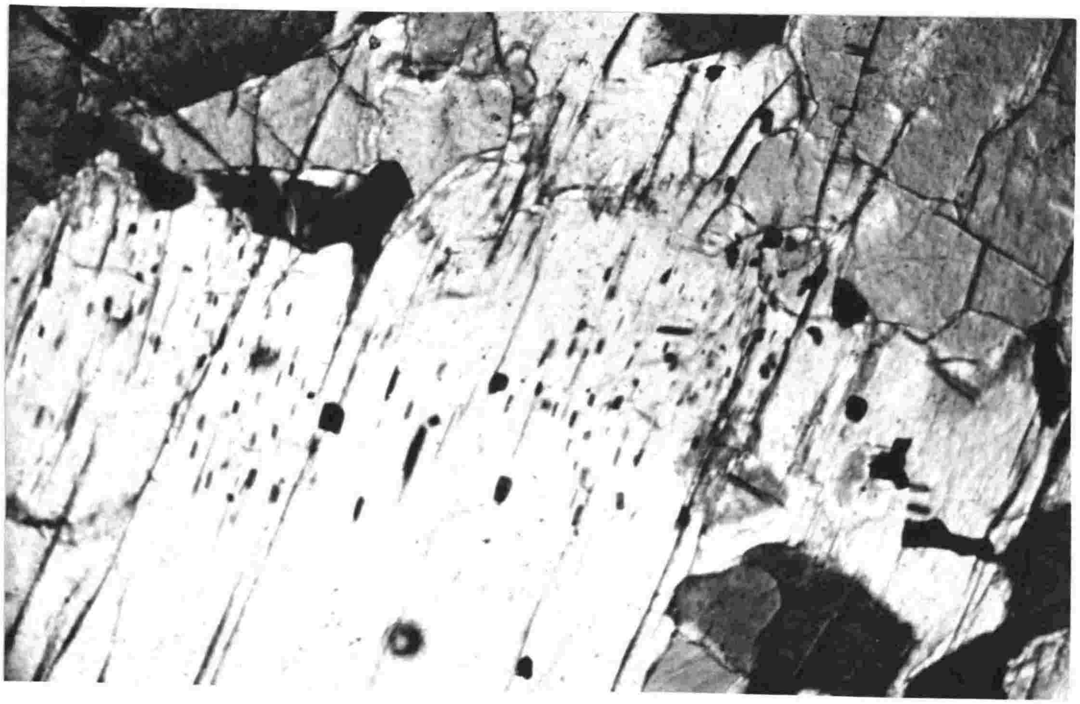


Fig. 39. Granular fragments of spinel enclosed in orthopyroxene (Opx) and olivine (Ol) grains.

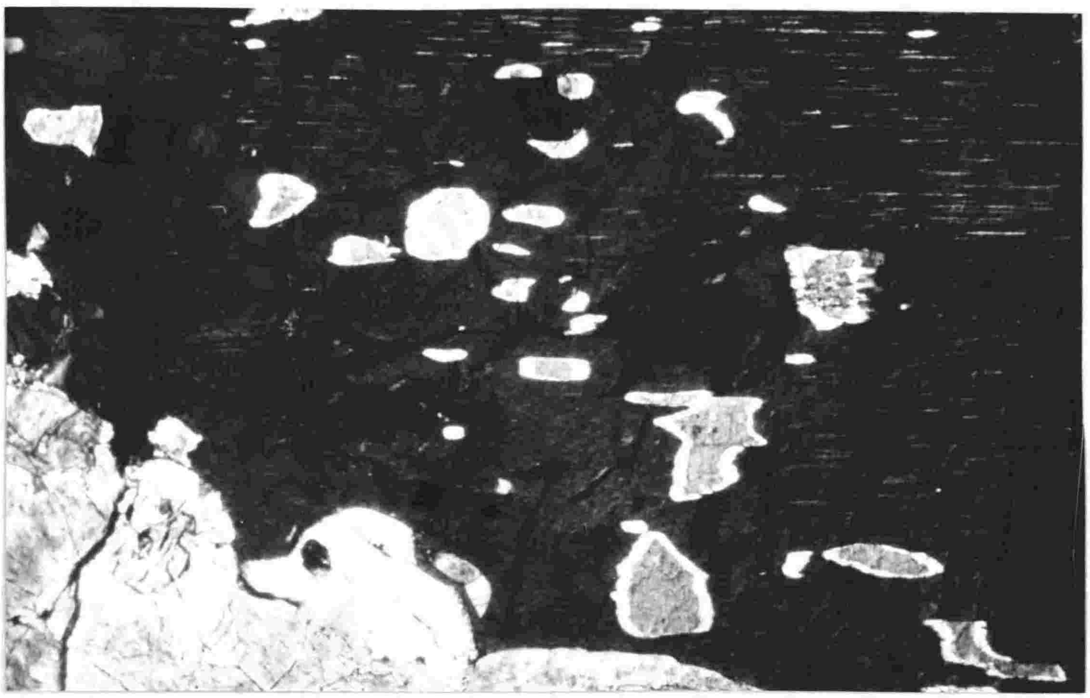


Fig. 40. Orthopyroxene grains (dark) replacing olivine (light). All olivine residuals have parallel extinction position.

spinel crosses a large otrhopyroxene grain and adjacent grains of olivine (fig. 39). The spinel was probably derived from a single spinel crystal which became granulated by protoclasis. (Several examples of small areas rich in fractured spinel grains have been observed in Basal Zone rocks.) The orthopyroxene evidently grew *in situ* surrounding and partly assimilating the grains of spinel. Note that the spinel grains within the orthopyroxene are all smaller and more rounded than those in the olivine. This probably reflects the greater absorption of alumina into the pyroxene than into the olivine lattice. The apparent alignment of the elongate grains parallel to orthopyroxene c-axis may be due to preferred absorption normal to c-axis.

Euhedral crystals and poikilitic textures.

Poikilitic orthopyroxene enclosing grains of olivine are of common occurrence. In general the enclosed grains are wholly anhedral though a few grains may show possible crystal faces. The enclosed grains are smaller and more rounded than the olivine grains which make up the remainder of the rock. The orthopyroxene host is anhedral with irregular boundaries. Such poikilitic grains may be interpreted as formed either by replacement of olivine by orthopyroxene, or by growth of an orthopyroxene porphyroclast surrounding olivine grains of an initially protoclastic matrix, or, as has been done by Challis (1965a), as indicating crystal sedimentation from a magma. In no case has the writer observed textures identical to the automorphic-granular textures described from the Stillwater Complex by Jackson (1961). Such textures have wholly euhedral grains of olivine surrounded by a poikilitic mesostasis.

Wholly euhedral crystals are exceedingly rare in the Red Hill Complex and even when present such grains make up only a very small fraction of the rock. For this reason the texture of the rocks is described as xenomorphic-granular. A possible euhedral grain has been featured earlier (fig. 32).

Euhedral olivines, determined on the universal stage, showing in some cases [010] sections with the development of [100] and [101] faces have also been observed in metamorphic layering of the Porters Knob Outcrop (p. 248). Such euhedra are probably of metamorphic origin. Although euhedra in monomineralic metamorphic rocks are not common they are not unknown, e.g. euhedral grains of glaucophane occur in almost monomineralic glaucophane-schists (pers. comm. R.H. Clark) and have been observed by the writer.

The occurrence of euhedra in metamorphic layering makes it invalid to use euhedral grains as conclusive evidence of crystal sedimentation. Niggli (1954, Rocks and Mineral Deposits, W.H. Freeman and Co. p.234) notes in this respect "attempts to use the degree of idiomorphism of crystals as a means of reconstructing the processes of crystallisation from solution, or of recrystallisation in solid aggregates, are fraught with very considerable difficulties. Too little criticism is brought to bear on such matters, and this has given rise to many problems more imaginary than real".

Challis (1965a) considers some rocks near the western boundary of the Red Hill Complex have hypautomorphic-granular textures resembling those of the Stillwater Complex and by parallel argument has interpreted the Red Hill layering as formed by crystal sedimentation from a liquid magma. This is reasonably interpreted as meaning that part or the whole of Red Hill Complex passed through a liquid magmatic stage. The writer hopes to examine at some future date the evidence of crystal sedimentation in the light of euhedra in replacement veins and the possibility that poikilitic textures may also be of replacement origin. At the present time Challis's interpretation for the rocks studied by her is tentatively accepted.

Zoning of crystals

Challis (1965a) does not record zoning in olivines from the Red Hill Complex but she describes and illustrates one case of zoning in an orthopyroxene grain. This grain has an euhedral core (En_{94}) with an anhedral rim of composition En_{86}. The writer has previously described (p. 139) zoned orthopyroxene grains from protoclastic harzburgites. The cores contain augite lammelae and the rims are free of such lamellae. It was not possible to determine any change in Mg/Fe ratio between cores and rims of these crystals.

S T R U C T U R E   O F   T H E   R O C K S

MESOSOPIC STRUCTURES

I n t r o d u c t i o n

The most conspicuous mesoscopic structures of the Red Hill Complex are compositional layering, mineralogically defined foliation and jointing. Layering and foliation, evidently formed during or before emplacement of the Complex, have great significance in determining ultramafic petrogenesis. Jointing, presumably largely developed during post-emplacement deformation, is not likely to be so important and has not therefore been studied. In the following section, the term mesoscopic structures refers implicitly to layering, foliation and allied rock structures such as lineation or lamination.

Nomenclature. Lineation is defined by parallel orientation of elongate minerals or mineral aggregates.

Foliation is a planar structure the individual folia of which cannot be traced more than a few inches and is of three kinds. It may be expressed by parallel orientation of tabular minerals (fig. 41, (i)) or aggregates (ii) or the folia may be defined by small, widely separated coplanar mineral grains (iii). The last is the usual form in which a feldspathic foliation occurs.

Lamination is produced by numerous parallel, closely spaced veins or platelike bodies of mineral grains only one or two crystal diameters thick. It differs from foliation in that individual laminae are persistent over distances of several feet.

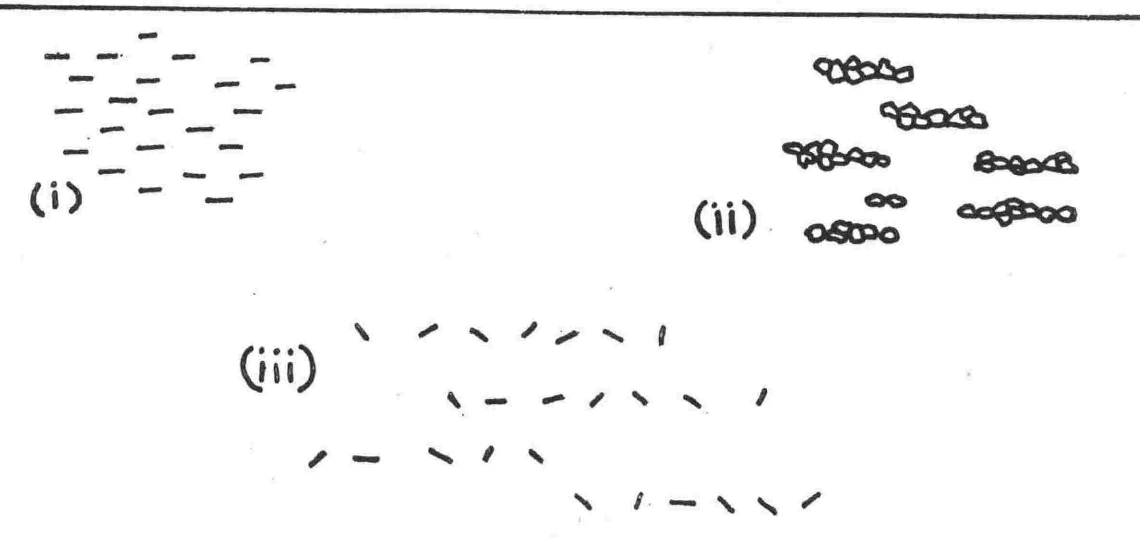


Fig. 41 FOLIATION

(i) Parallel tabular crystals
(ii) Parallel aggregates of crystals
(iii) Widely separated coplanar grains

Layering is a planar structure expressed by parallel tabular or lenticular bodies of rock of differing mineralogical composition. In this sense of the term, closely spaced parallel veins may give a layered structure. This extension of normal usage is necessary as no clear distinction can be made between multiple parallel veins (for instance dunite in harzburgite) and layering formed by some other mechanism.

In some outcrops there are two intersecting layered structures. These will be referred to as 'early' and 'late' layering depending on the cross-cutting relationship.*

D e s c r i p t i o n

Lineation is not common in the Red Hill Complex and is only observed in the lowest rocks of the Upper Zone, where it is produced by parallel orientation of elongate aggregates of orthopyroxene grains. The individual grains show strongly preferred orientation with their c-axes parallel to the lineation (p.179). Microscopically, well developed lamellae parallel to (100) can be observed (No. 10967) and usually the grains are strongly cleaved on {210}. The lamellae are distinct in that they are much more numerous, finer and closely spaced than exsolution lamellae and are probably translation lamellae (Turner and Weiss, 1963, p. 359). The texture of the rock is protoclastic with only minor recrystallisation and evidently the lineation is a deformational feature.

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* The terms 'primary' and 'secondary' are commonly used in the same sense but carry strong genetic connotations. Primary layering for instance, usually implies layering formed by crystal sedimentation (e.g.Thayer,1963). As shown in a later section there is no justification for assuming that 'early' layering has developed in a different manner from the 'later'. The terms 'primary' and 'secondary' are therefore not used.

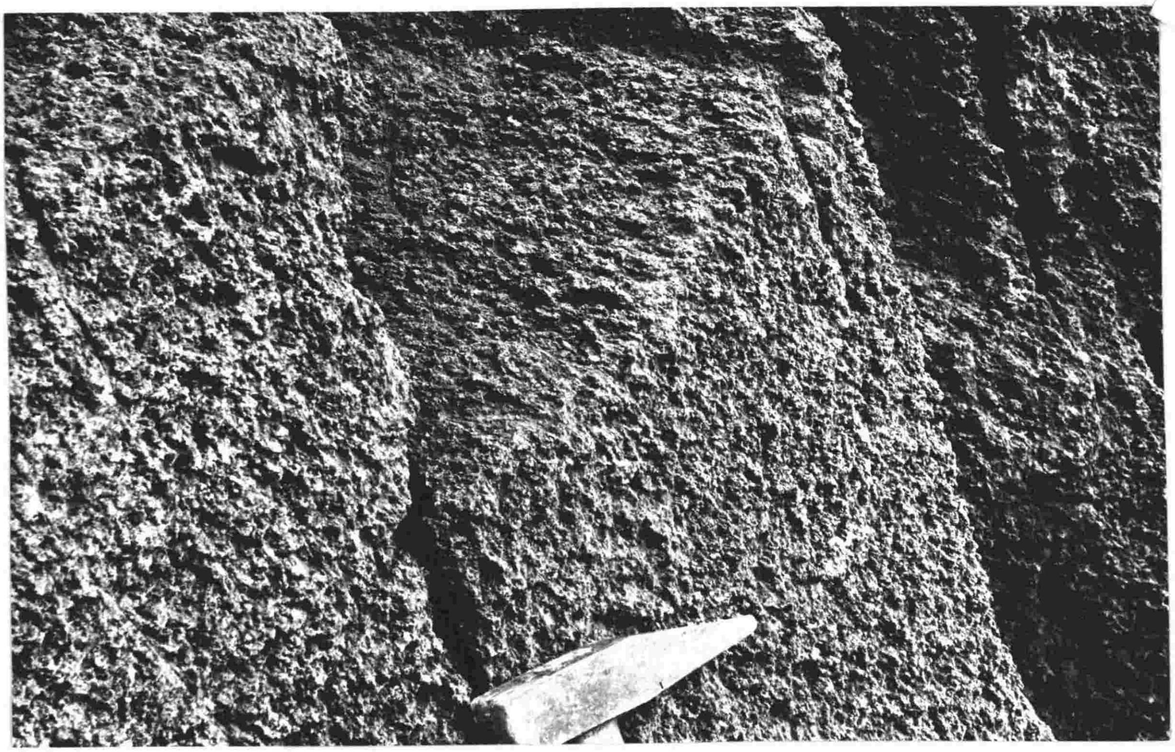


Fig. 42. Lineation (top,centre) in protoclastic harzburgite near boundary with Basal Zone. The lineation determined by elongate aggregates of orthopyroxene.

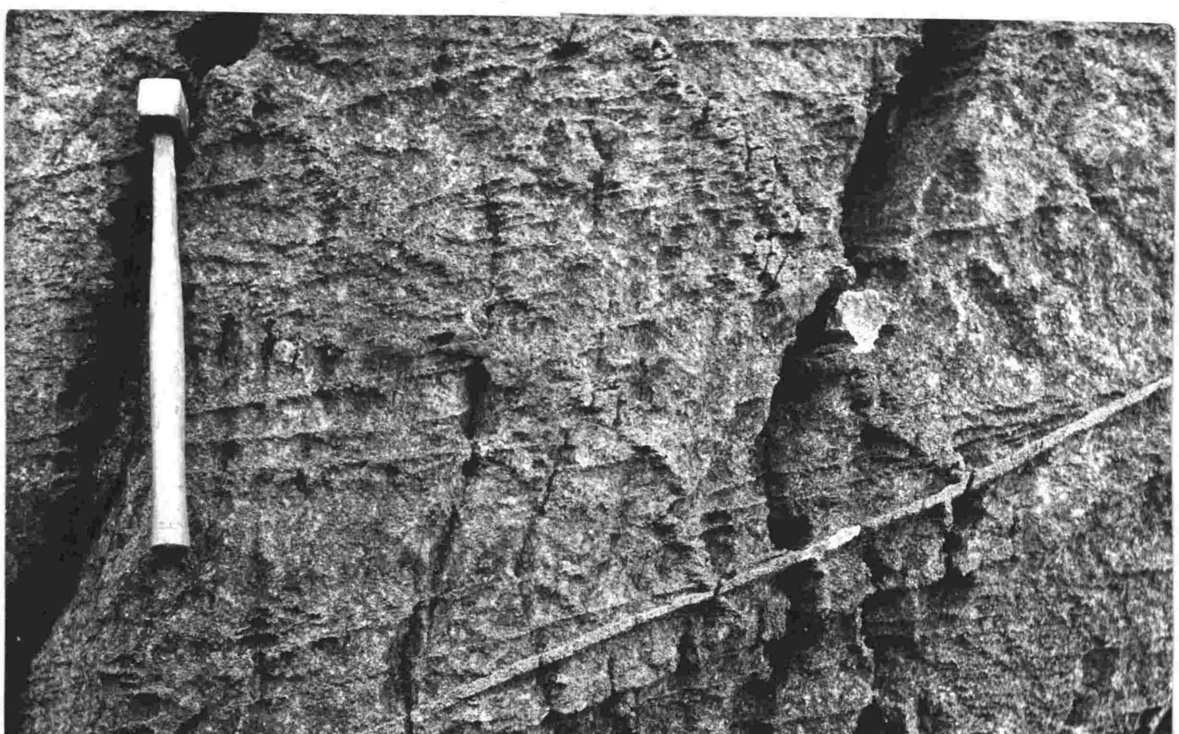


Fig. 43. Feldspathic foliation formed from earlier veins that were inclined steeply to the right. Foliation parallel to late veins and dips to the left.

Fig. 44. Feldspathic foliation associated with diopside veins. (see text)

In strongly weathered surfaces (fig. 42) the lineation is easily observed and lies in the plane of foliation developed nearby.

Foliation. A weak foliation defined by parallel aggregates of orthopyroxene is commonly observed in protoclastic harzburgite slightly higher in the Complex (i.e. further from the upper boundary of the Basal Zone than the lineation). Rarely a lineation can be obeerved in the plane of foliation but is very weak. The foliation is widespread and is usually parallel to thin layered structures (dunite and Harzburgite) developed higher still in the Complex.

The most conspicuous and best developed foliation occurs in peridotites containing a few per cent of feldspar. The feldspar is generally sparsely disseminated to the extent of 5 per cent in patches about a foot across. Surrounding the patches or pockets is non-feldspar-bearing peridotite of the same mineral composition (less the feldspar) and texture as the feldspathic-peridotite. The foliation is of type (iii) (fig. 41); widely separated small feldspar grains lying in folia throughout the rock. The feldspathic foliation is sometimes accompanied by an orthopyroxene foliation of the type described above.

Commonly a feldspathic foliation is associated with thin feldspar and/or diopside bearing veins (fig. 43). Two periods of veining can be distinguished, one partly-disrupted set dipping steeply to the right, the other dipping to the left of the photograph. The foliation is sub-parallel to the later veins and the feldspar of the foliation has probably has probably been derived from the earlier veins. Another example (fig. 44) shows more clearly the derivation of feldspar from an earlier vein. In this case, the feldspar folia are restricted to a 4 inch wide zone which is traversed down the middle by a 'vein' of diopside. The diopside grains are discontiguous, probably the result of recrystallisation of the peridotite after injection of the vein. The vein is considered to have been initially composed of feldspar and diopside. A change in orientation of the stress pattern resulted in migration of the feldspar along schistosity planes and recrystallisation of the peridotite. That recrystallisation has occurred is evident in fabric studies (p.185, No. 10987).

Lamination. As the proportion of feldspar in peridotite increases, the feldspathic foliation becomes better defined, individual folia more persistent and foliation grades into lamination. An example of a structure intermediate between foliation and lamination is shown in figure 45. In this rock the proportion of feldspar is about 8 per cent.

A laminated structure formed by closely-spaced veins composed of feldspar and diopside is illustrated in figure 46. The veins cut a thinly-layered dunite-harzburgite sequence, the residuals of which can be recognised between the laminae. (A hand specimen from the laminated rock is composed of feldspar and diopside acquired from the veins and olivine and orthopyroxene from the layered sequence, i.e. it is a eucrite.) In thin section (No. 10980) the structure evident in the field can be recognised only with difficulty; apparently recrystallisation has tended to obliterate on a microscopic scale the distinction between veins and layering.

Lamination is a common structure throughout the upper two thirds of the Upper Zone and in addition to feldspar, is commonly composed of pyroxene laminae (fig. 47). The lamination is confined to a 6 inch band

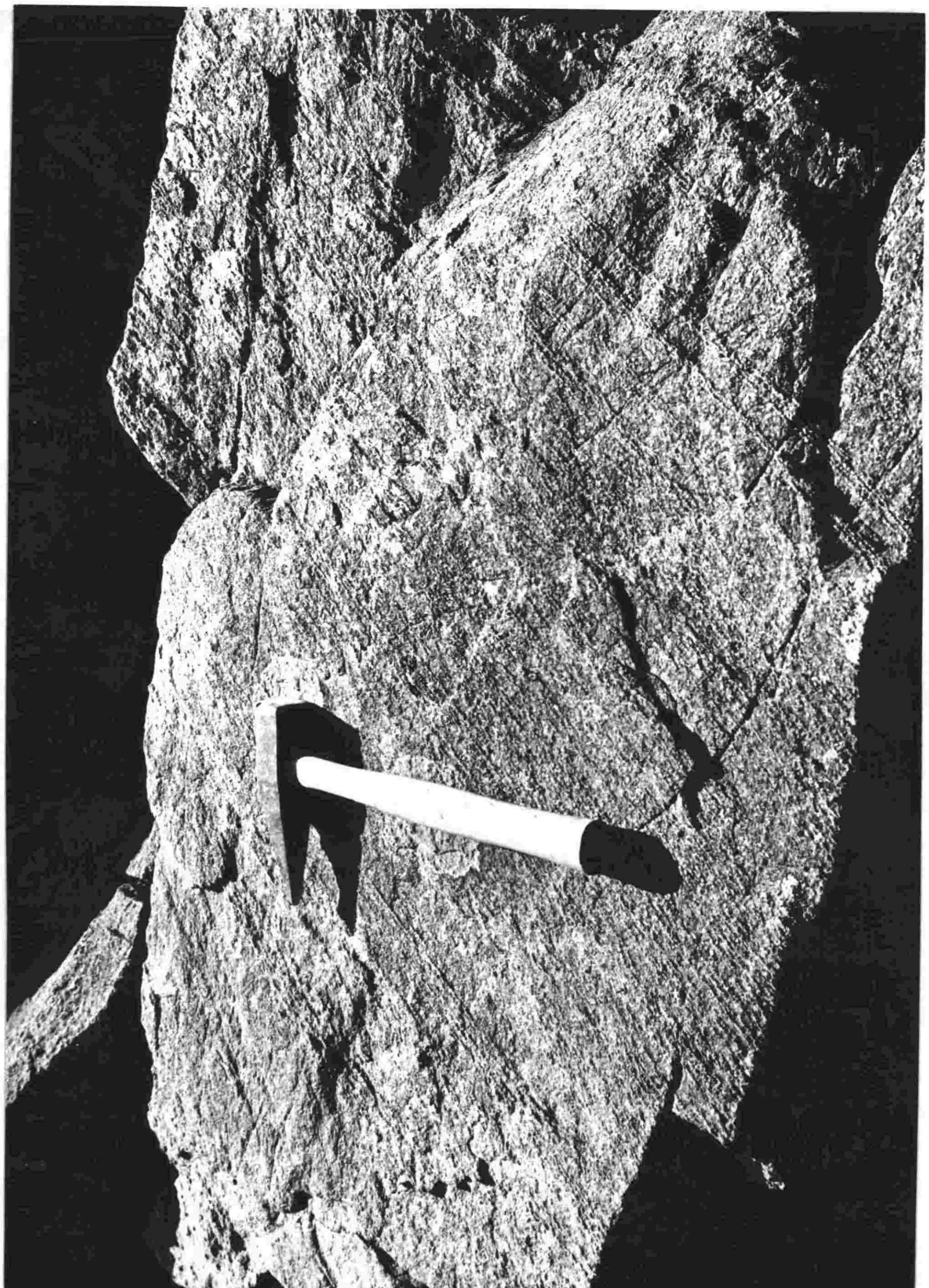


Fig. 45. Feldspathic foliation-lamination in feldspathic-harzburgite.

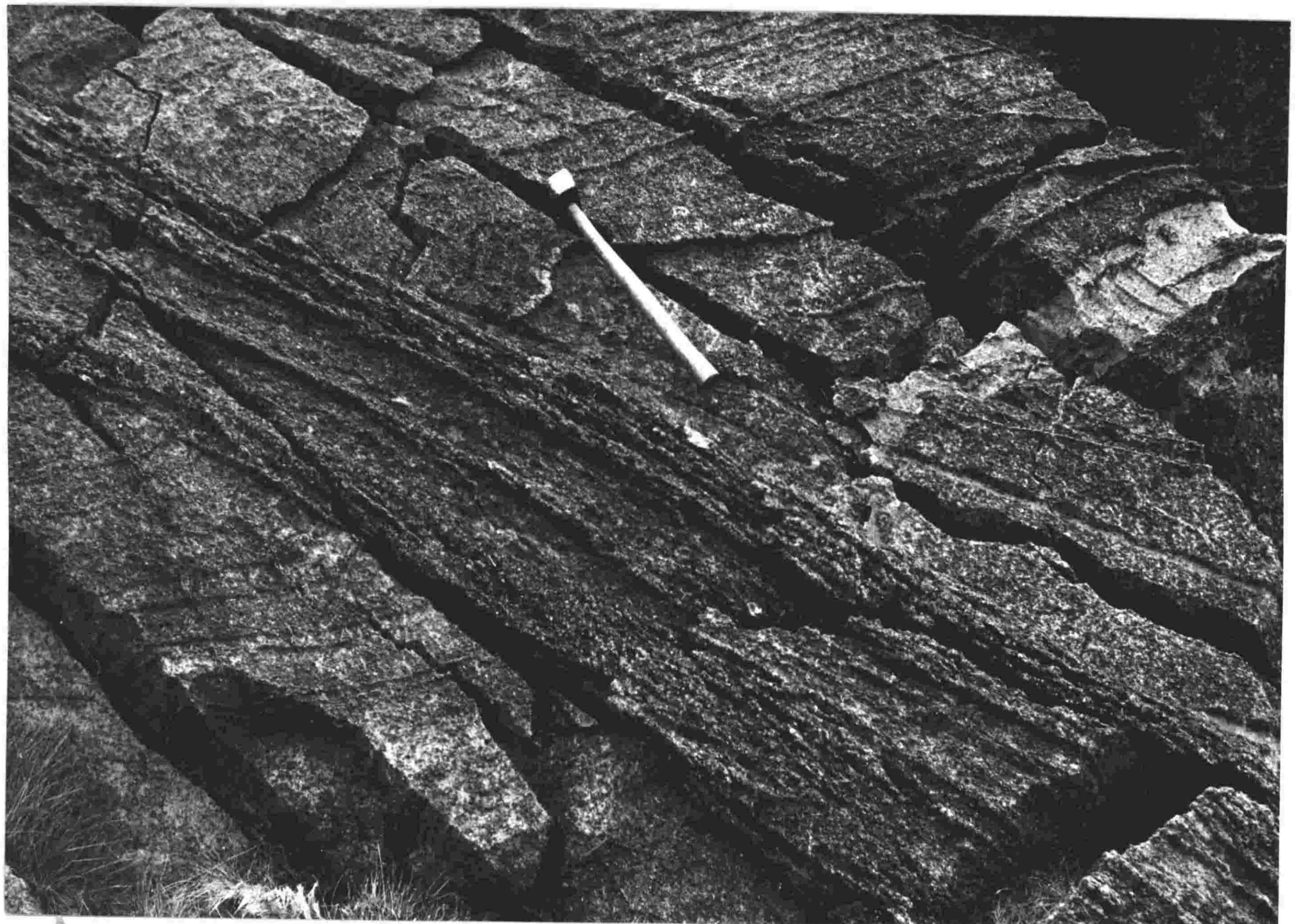


Fig. 46. Feldspar-diopside lamination cutting layered dunite-harzburgite sequence.

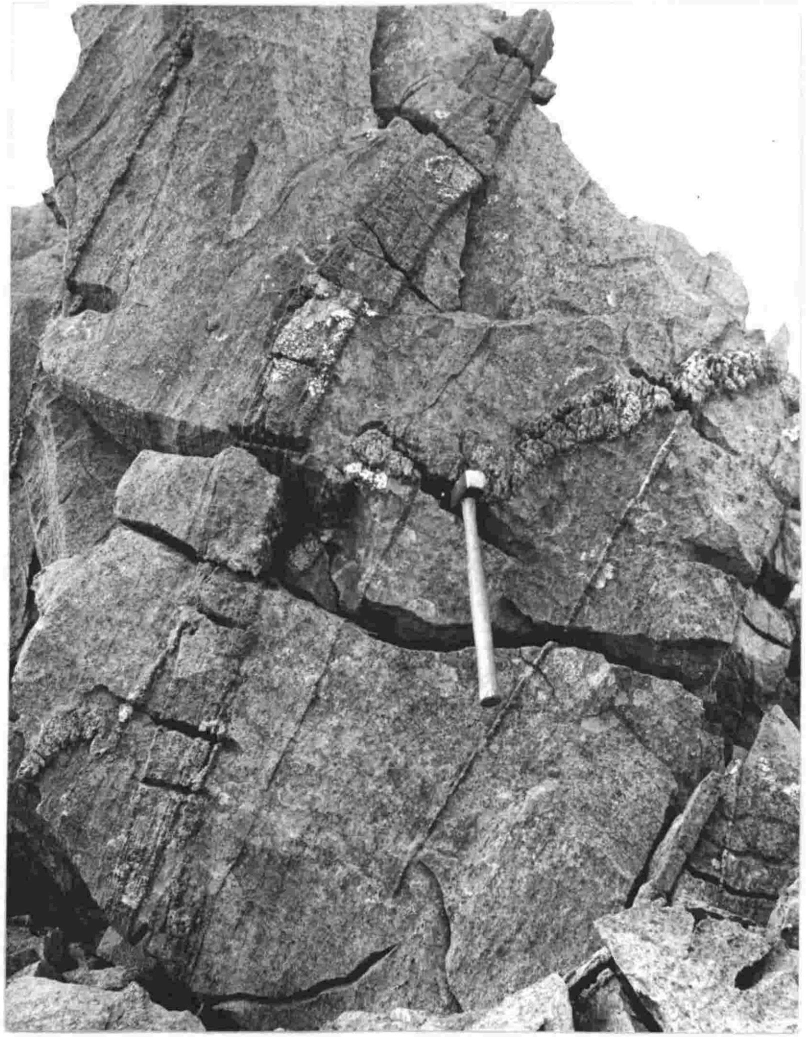


Fig. 47. Laminated chrome-diopside layer cutting dunite and cut by a pyroxenite pegmatite.

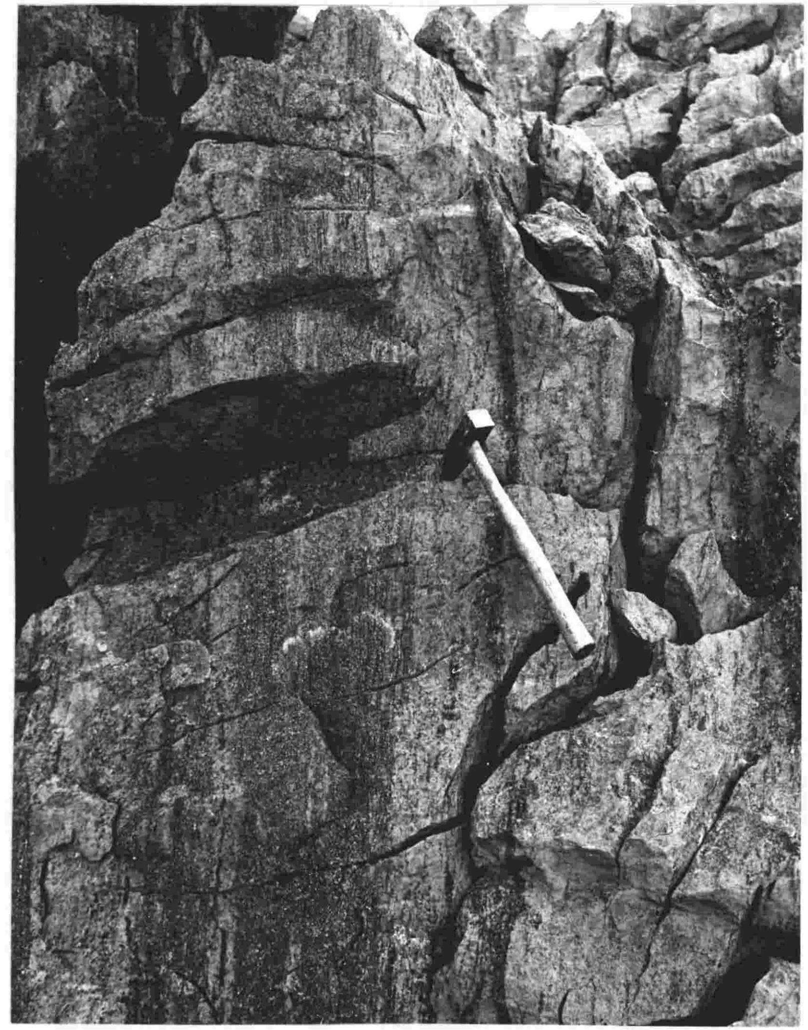


Fig. 48. Orthopyroxene lamination in dunite. The lamination grades to the right and also laterally into a pyroxene foliation.

cutting dunite and is parallel to nearby layering in harzburgite.

Orthopyroxene lamination (fig. 48) like that of feldspar, grades into pyroxene foliation with decrease in the proportion of pyroxene in the rock. However, lamination is only common in the upper part of the Upper Zone and evidently some factor other than the proportion of pyroxene in the rock is necessary for its development.

Layering. Layering is strongly developed to the west of the right branch of the Motueka River and although not very well exposed elsewhere, is also recognisable over much of the rest of the Red Hill Complex. It is absent in the rocks in the lowest part of the Upper Zone (where the mesoscopic planar structures are pyroxene or feldspathic foliation) as well as in the Basal Zone. In general, layering becomes increasingly well developed towards the stratigraphic top of the Complex.

Typically, layering is shown by alternating layers of dunite and pyroxene-bearing peridotite, usually harzburgite (fig. 49) each about two inches to two feet thick, but considerable variation in thickness and composition of layers occurs.

The pyroxene-bearing peridotite layers usually are laminated or foliated parallel to layering (fig. 50) but in some outcrops this is difficult to observe. This is because weathering of cracks and joints in the rock results in deep etching of the surface, obscuring any lamination or foliation that may be present.* The harzburgite layers of figure 51, for instance, are, on close examination, finely laminated in the manner illustrated in figure 50.

* Deep etching occurs prevalently in unserpentinised peridotites; partly serpentinised peridotites generally weather evenly and the individual mineral grains can be easily recognised. This is probably due to serpentine minerals closing surface cracks following expansion during serpentinisation.

Superficially, the layering in the Red Hill Complex resembles layering in "Stratiform Complexes" such as the Stillwater Complex, and Challis (1965a) believes both have formed in a similar manner, i.e. by crystal sedimentation from a crystallising basaltic magma. However, many features of the layering of the Red Hill Complex are not consistent with this hypothesis.

(i) Grading is only rarely observed in a layered sequence (e.g. fig. 53) and where present the direction of grading is not the same everywhere.

(ii) Although layering and foliation are generally parallel, a cross-cutting relationship is frequently observed. An example is shown in figure 53, in which a dunite layer (parallel to others nearby) cuts across, or is cut by, a weakly defined foliation in harzburgite. (Note that the fracture pattern in the right-hand part of the dunite layer is similar to a structure illustrated in Bowes, Wright and Park, 1964, plate 6 (b) thought by those writers to be a current-bedded structure formed by crystal sedimentation. It is considered here to be a deformational feature.)

Another example is shown in fig. 54 in which a combined feldspathic and pyroxene foliation cuts an earlier layering of 2 inch dunite layers in feldspathic harzburgite.

(iii) Two or more sets of layering are commonly developed in outcrops near the hinge of the macroscopic folds (see below), (figs. 55 and 56). In figure 55, fine layering (note grading) in harzburgite is cut and displaced by layers of dunite and laminated harzburgite. In figure 56, an early set (S_1) of layers of dunite and harzburgite has been cut, displaced and folded by movement parallel to the late set (S_2) of dunite layers and pyroxene veins. A pyroxene vein intruded after S_1 and before S_2 (and hence referred to as an S_1 pyroxene vein) has been folded strongly by flow parallel to S_2.

A large-scale map of the Porters Knob area (Map II in rear pocket) shows the orientation of 'early' and 'late' layering in a small area of the Red Hill Complex. Outcrops on the limbs of the macroscopic fold whose hinge-line passes near Porters Knob usually have only a single set of layering with or without a parallel foliation. Near the hinge of the fold, it is usual for outcrops to have two sets of planar structures, sometimes an S_2 foliation cutting an S_1 layering, but usually an S_2 layering cutting an S_1 layering.

In all cases the later set (S_2) is parallel to the single layering of the eastern limb, i.e. dips at about 25° to the south-east.

(iv) Schlieren-like structures (figs. 57, 58 and 59) are common in parts of the Red Hill Complex, most notably north-west of Porters Knob. Their formation is ascribed to plastic flow with resulting disruption and attenuation of original layering (most clearly evident in figs. 57 and 58) and to the complex veining of massive harzburgite by dunite (probably of replacement origin), (fig. 59).

In these respects, layering of the Red Hill Complex differs considerably from that of the "Stratiform Complexes" and is characteristic of Alpine-type peridotite-gabbro complexes (Thayer, 1960).

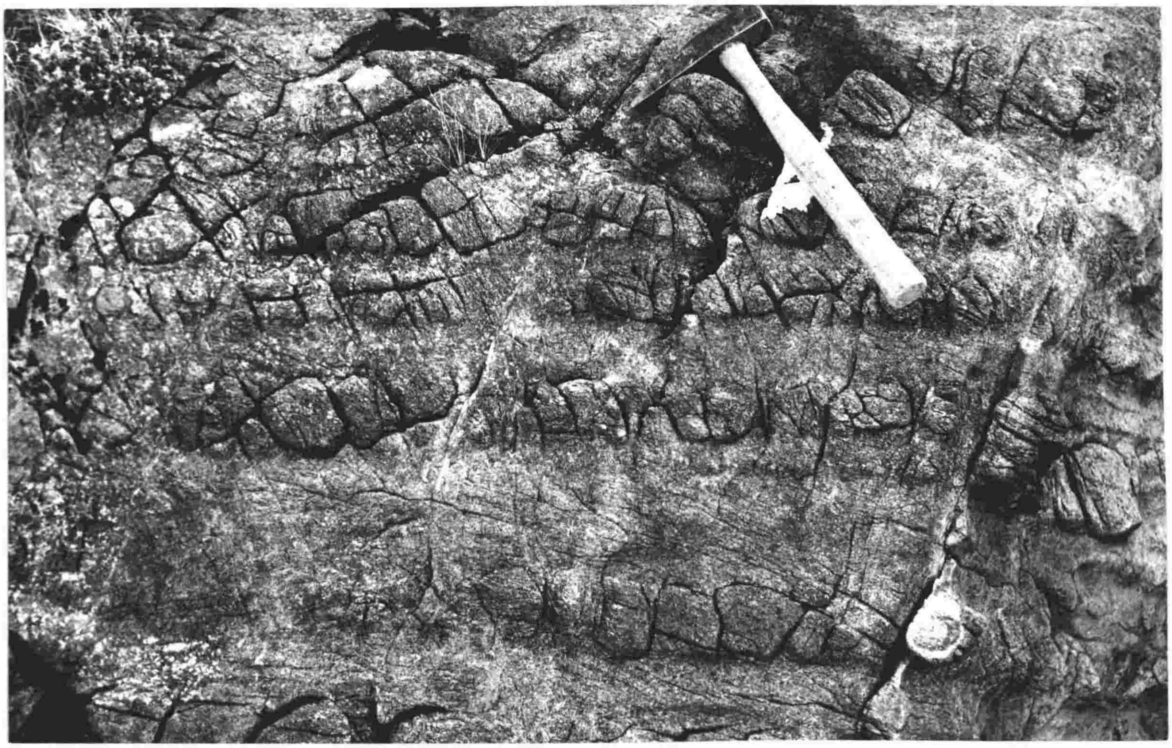


Fig. 49. Dunite-harzburgite layering. Dunite generally smooth; harzburgite strongly jointed and in relief above dunite layers.

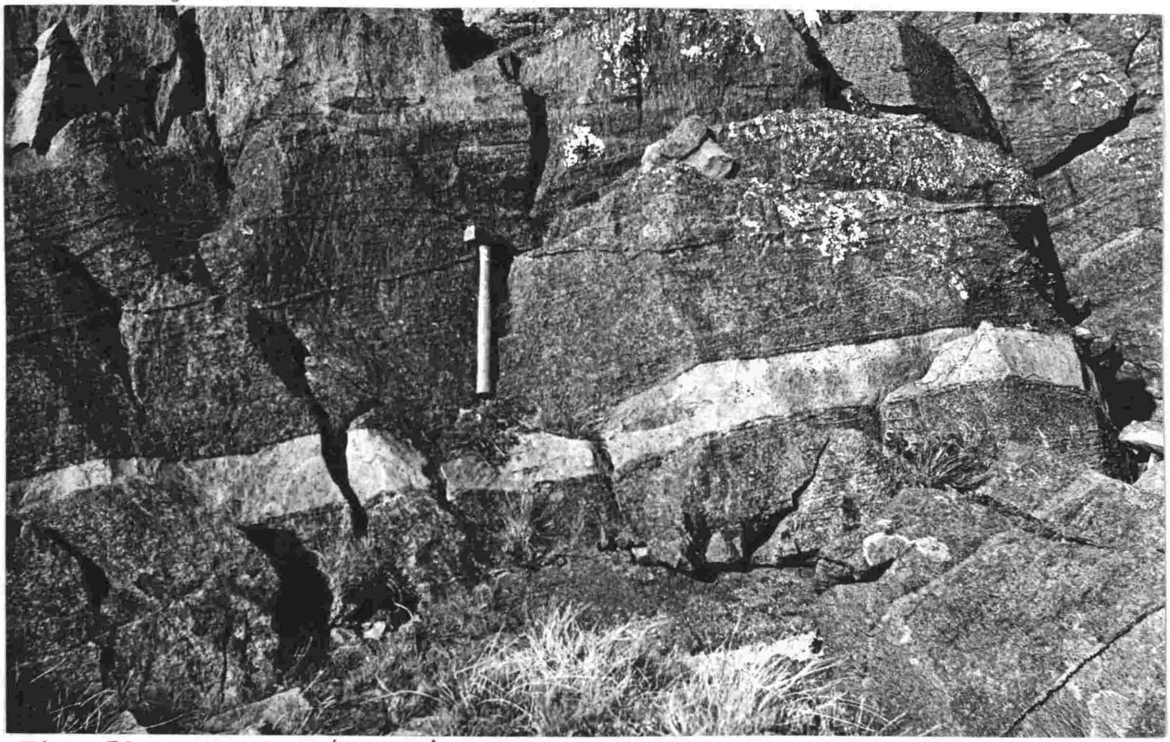


Fig. 50. Dunite (light) and harzburgite layers (dark). Harzburgite finely laminated.

Fig. 51. Layered dunite (smooth surface) and harzburgite (broken).

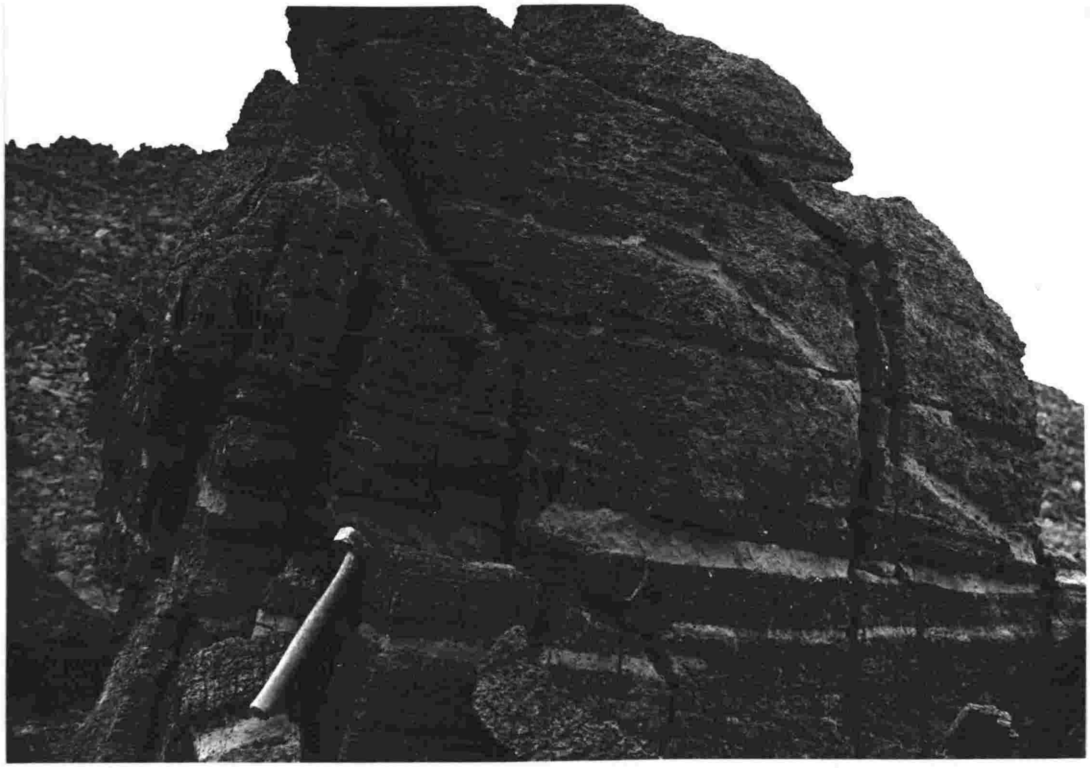


Fig. 52. Intersecting layers and veins of dunite (light) in protoclastic harzburgite (finely etched).

Fig. 53. Dunite layer (parallel to others not shown in photo.) cutting foliated harzburgite. Note fracture pattern in dunite on right of photograph.

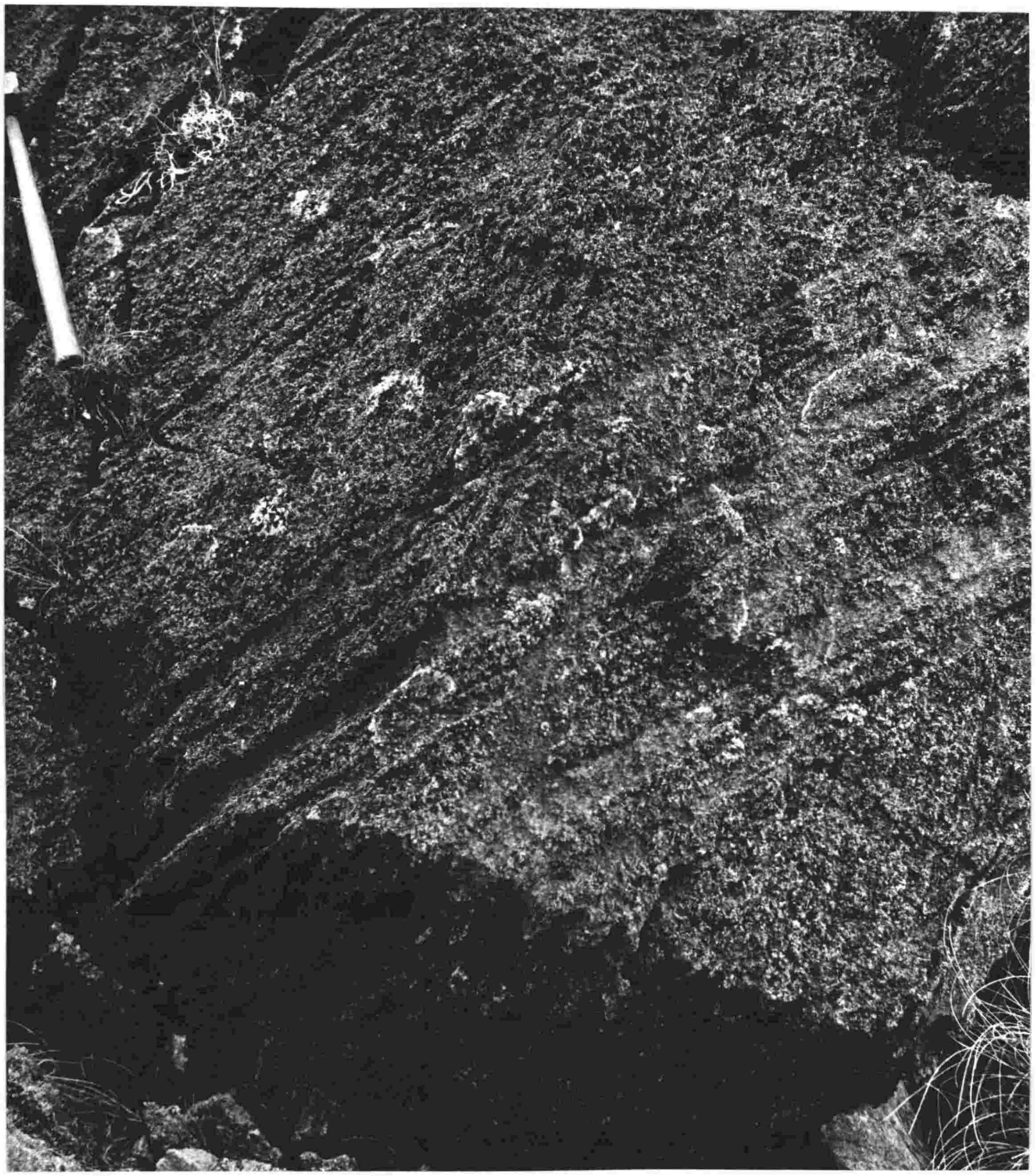


Fig. 54. Foliation cutting layered dunite (deeply etched) and feldspathic harzburgite layers. Foliation dips very steeply, and inclined about 40° to layering.

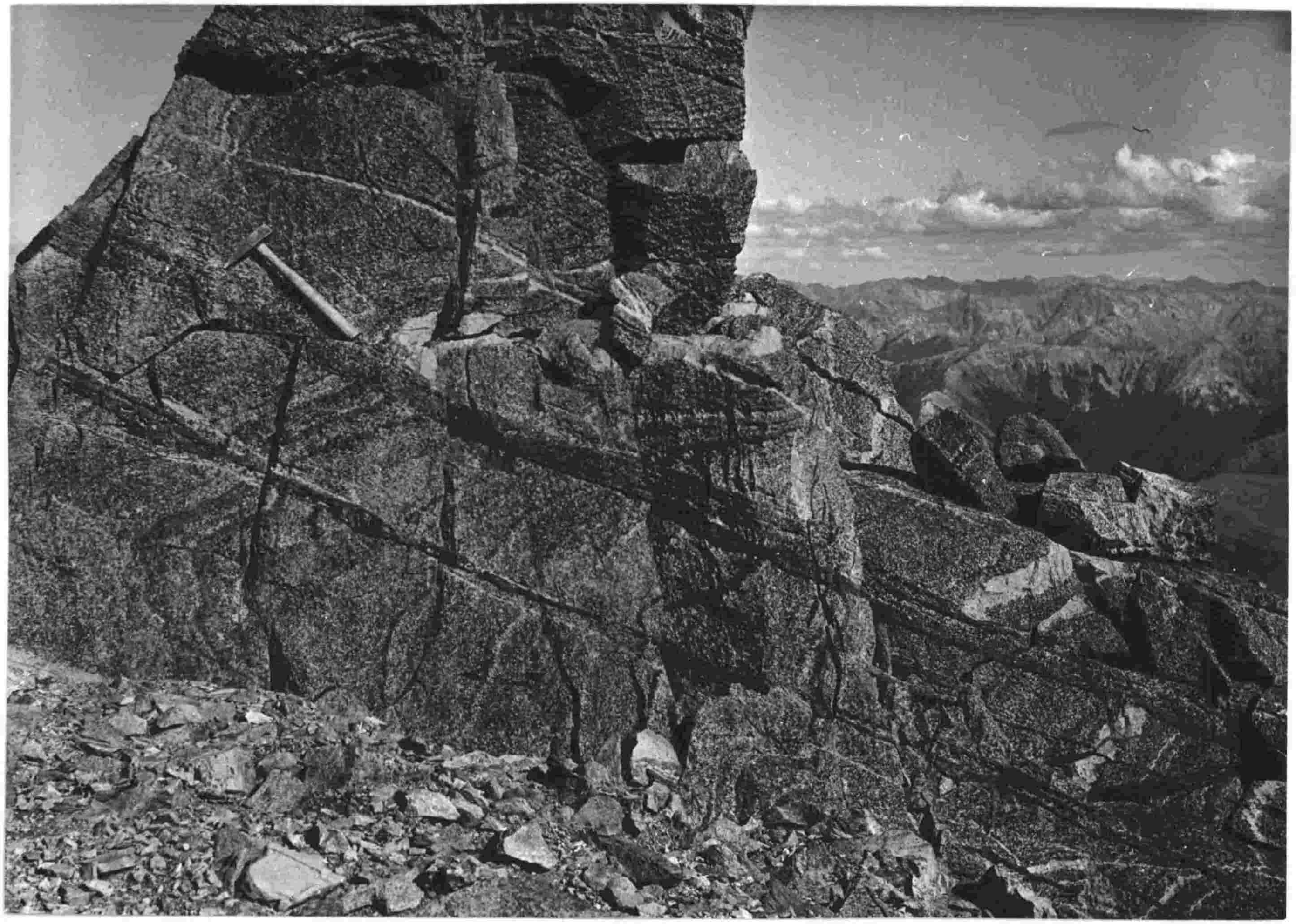


Fig. 55 'Early' folded layering offset on 'late' layering of dunite (light coloured) and laminated pyroxenite veins.

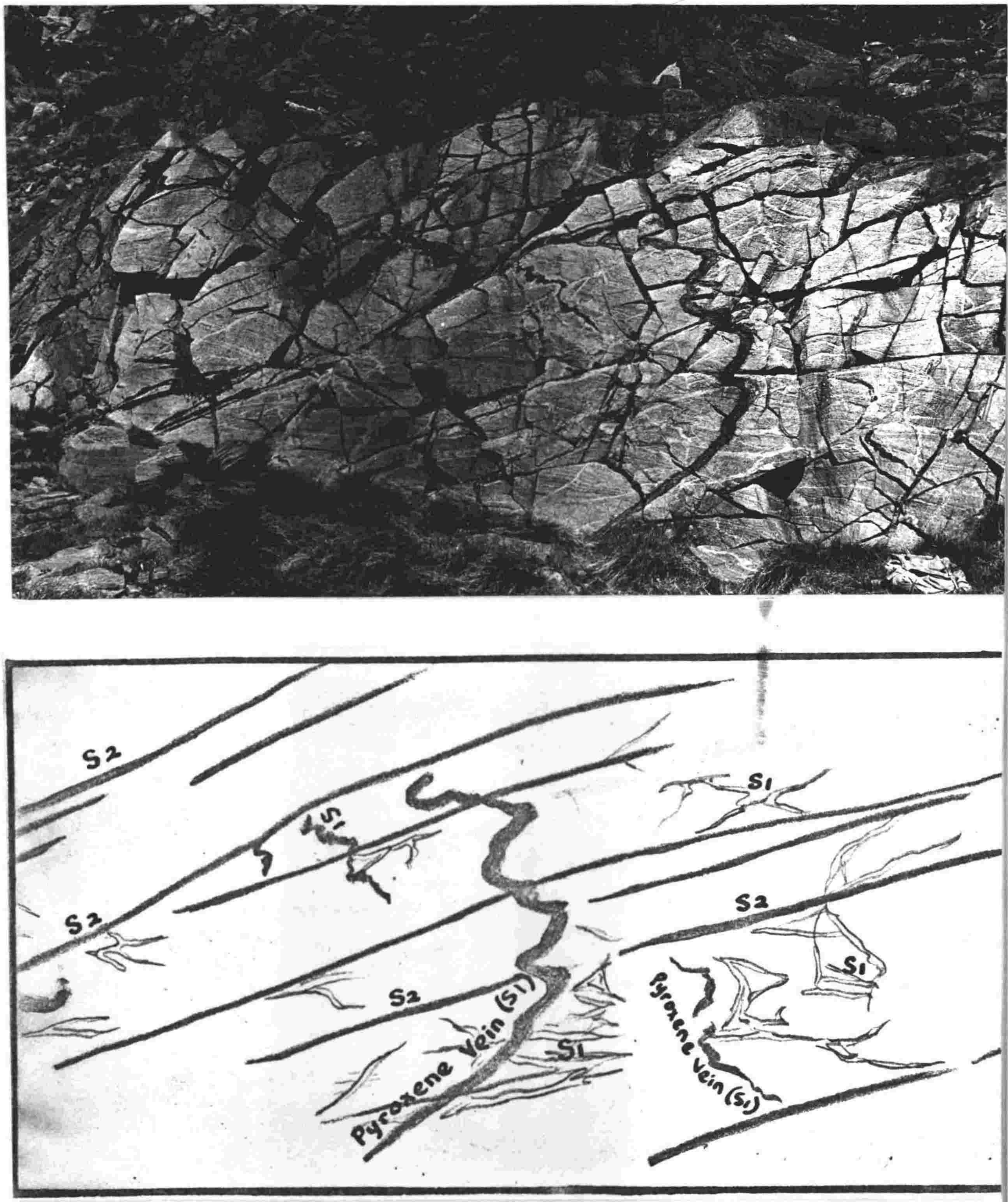
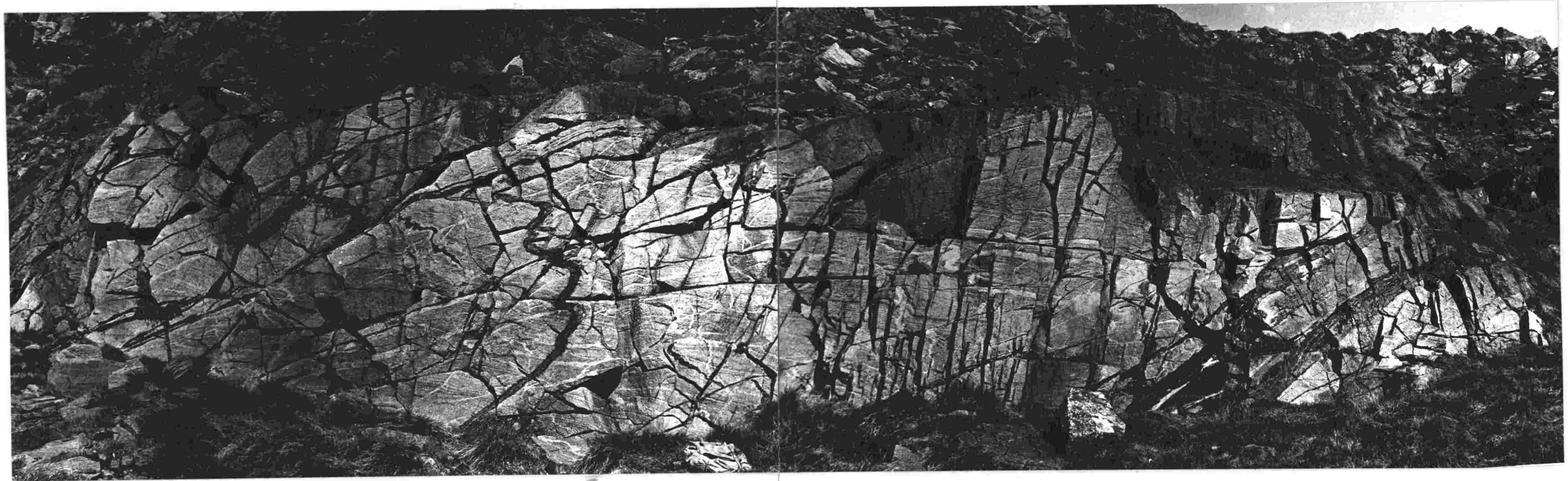


Fig. 56. Cross-cutting layers. 'Early' layering (S_1) is flat lying and is cut by 'late' layers of parallel dunite and pyroxenite veins dipping at about 30° to the left.



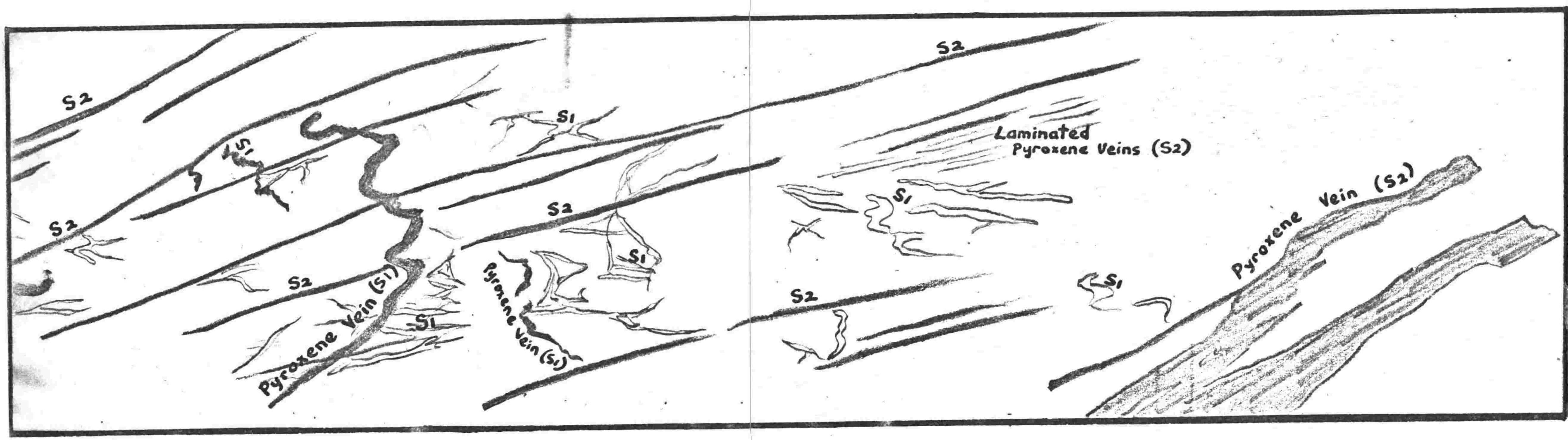


Fig. 56. Cross-cutting layers. 'Early' layering (S_1) is flat lying and is cut by 'late' layers of parallel dunite and pyroxenite veins dipping at about 30^o to the left.

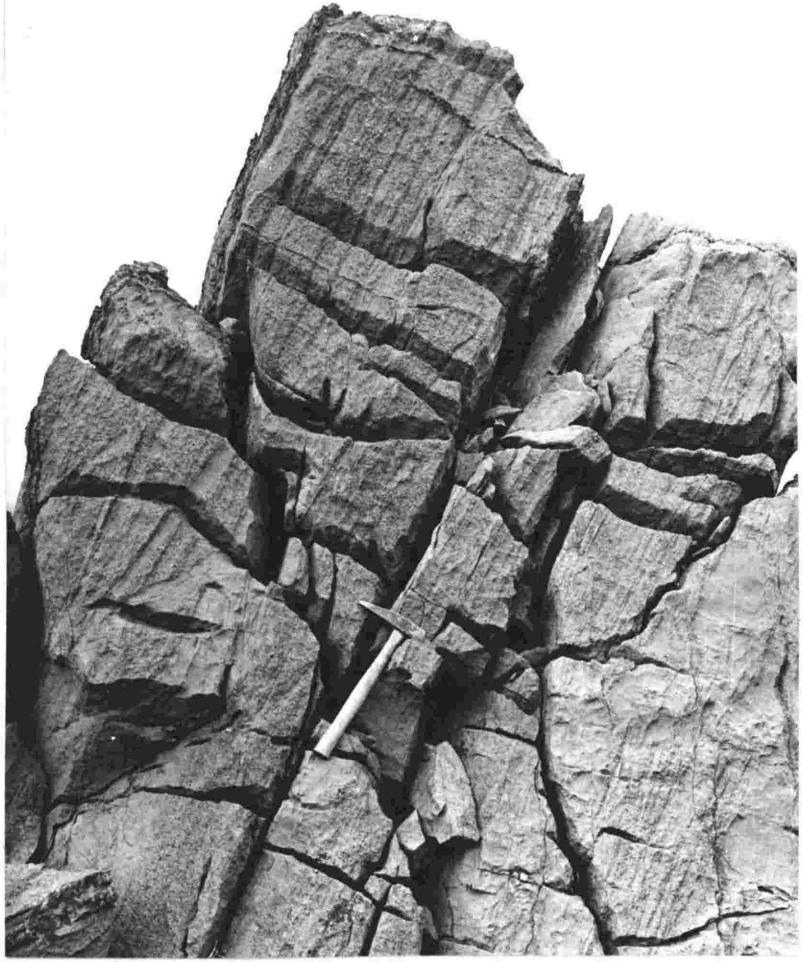


Fig. 57. Dunite-harzburgite layering.

Fig. 58. Detail of fig. 57,(lower right) Pyroxene-rich layers disrupted by flow in plane of layering. Schlieren-like structure.

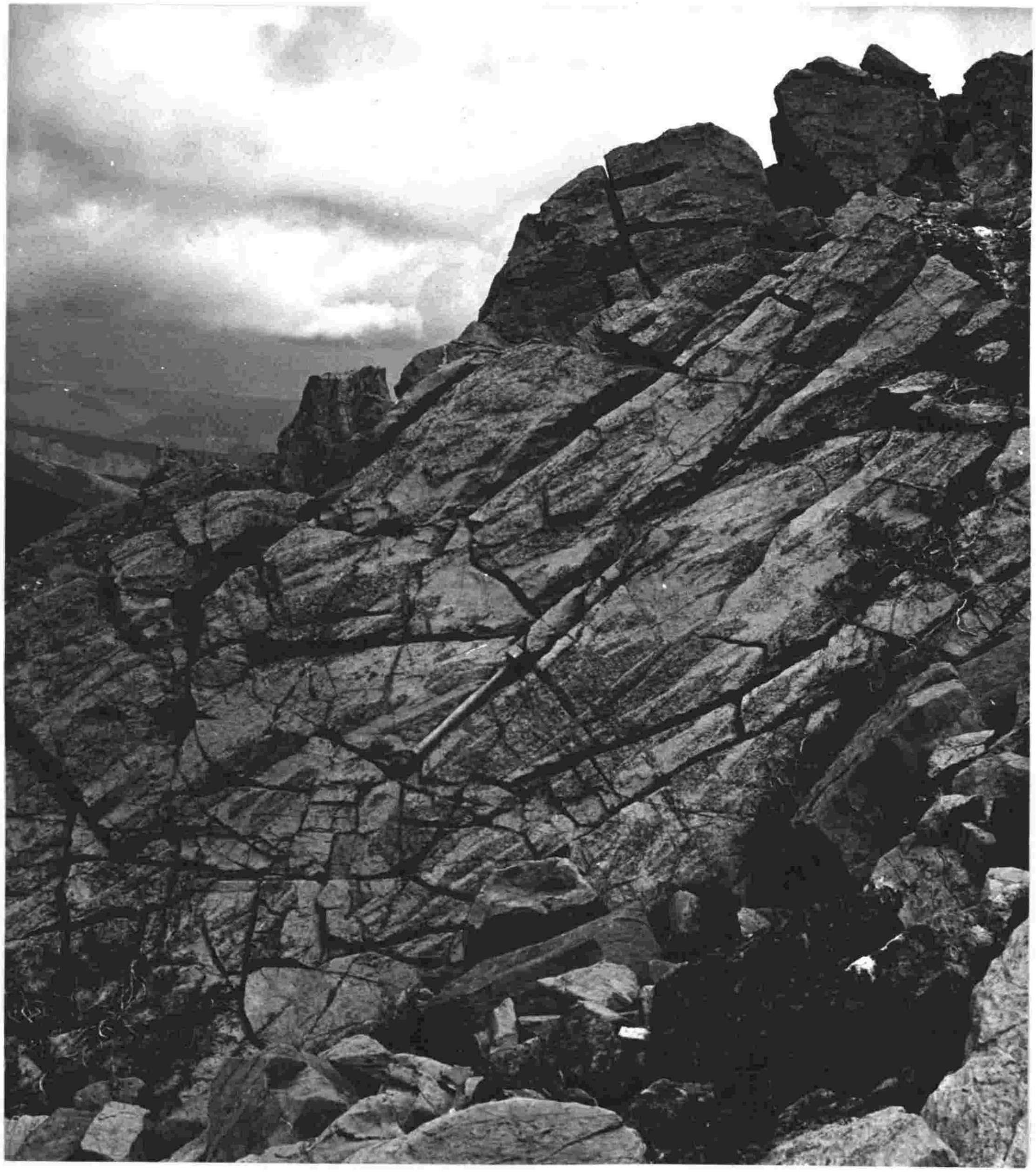


Fig. 59. Harzburgite schlieren in dunite (lighter coloured). Probably formed by intersecting network of dunite veins cutting original massive harzburgite.

Veins and Pegmatites. Planar or curvi/planar bodies of rock composed of one or more of the five constituent minerals of peridotite and gabbros (p.115) and formed by introduction of material into, or replacement of, peridotite or gabbro are referred to as veins when the grain-size is approximately the same as the country rock or pegmatite when much coarser. The distinction between pegmatites and veins is arbitrary as the grain-size varies continually between the two. For convenience, veins will generally refer to bodies with a grain-size less than 1 cm.

The criteria for recognition of veins and pegmatites, is generally a cross-cutting relationship, thus the late layering described above is ascribed to multiple parallel vein formation.

Veins and pegmatites occur throughout the Complex but are most abundant and thick near the stratigraphic top and very rare, thin and fine grained in the Basal Zone. They may be composed of dunite, pyroxenite, anorthosite or a combination of pyroxene and feldspar. Chromitite veins have not been observed but chromite occurs as euhedral grains about one centimetre across and constitute about 10 per cent of the vein. Fine-grained veins are generally planar and parallel to layering (fig. 61). Close inspection (fig. 62) shows a close relationship between layering and veins. Much of the layering can be seen to have formed by vein injection which subsequently became disrupted, in part, to form discrete, discontiguous grains. Only remnants of the veins remain which as traced laterally into layerings become less well defined

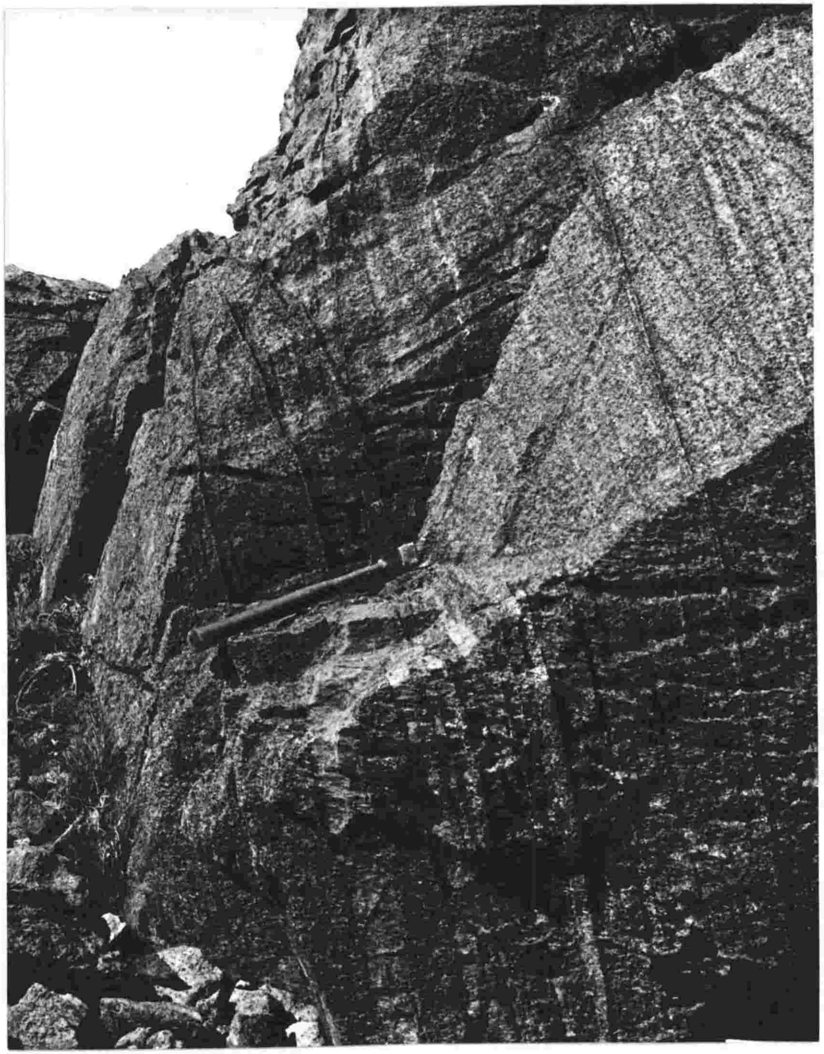


Fig. 61. Thin dunite-harzburgite layers cut by subparallel veins of orthopyroxene.

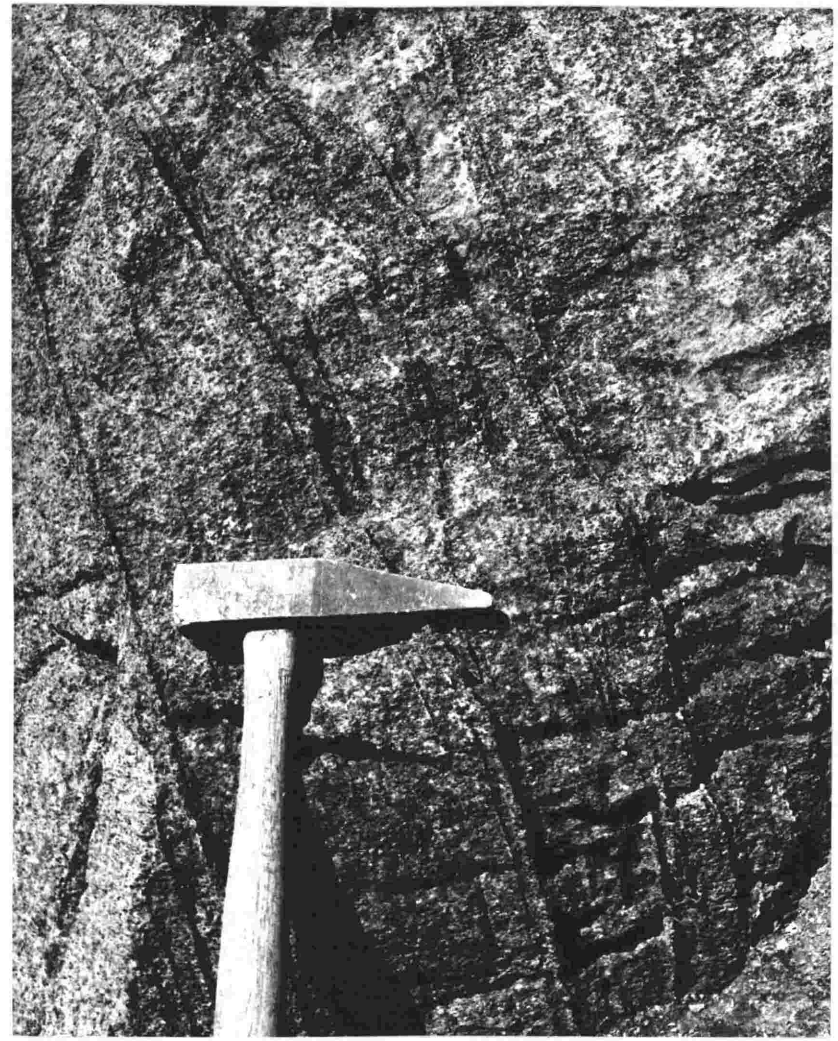


Fig. 62. Detail of fig.61
Note: Veinlets to right of hammer head merge into diffuse harzburgite layers.

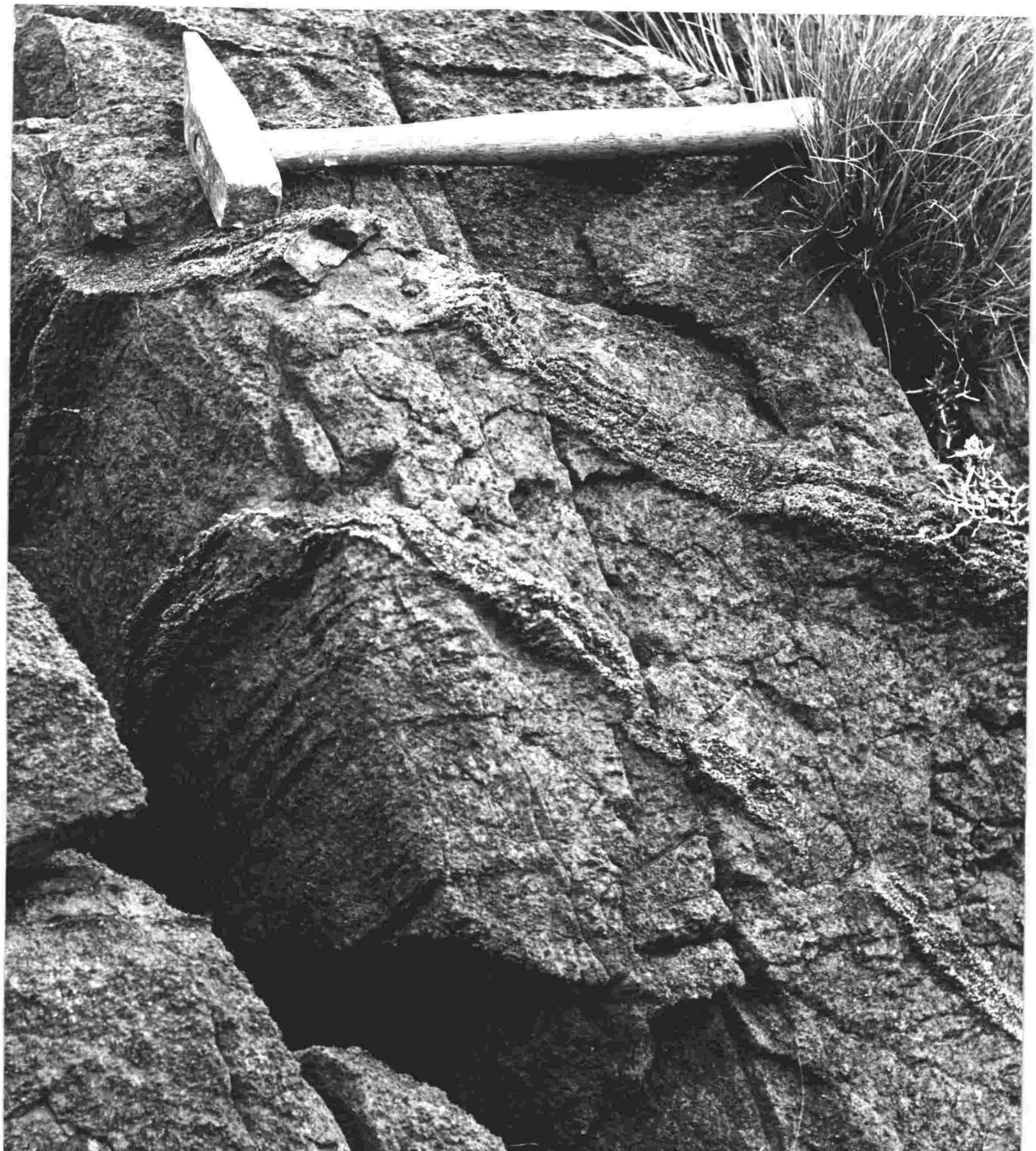


Fig. 63. Two laminated eucrite veins cutting foliated harzburgite.

Fig. 64 Orthopyroxene pegmatite cutting pyroxenite vein.

and are eventually marked only by aligned, but separate, pyroxene grains in the peridotite. The disruption process is thought to be recrystallisation.

Veins of eucrite composition are shown in figure 63. They occur as thin, tabular bodies cross-cutting an earlier pyroxene foliation in harzburgite.

Pegmatites are generally irregular in thickness and orientation, frequently strongly curved and undulatory. A pyroxene vein (fig. 64) parallel to weak foliation in pyroxene-bearing dunite is cut by a coarse-grained pegmatite of orthopyroxene. Note the change in thickness and irregular boundaries of the pegmatite.

An important observation related to the genesis of the veins and pegmatite is that wherever two veins or pegmatites cross-cut, the coarse-grained veins (or pegmatites) are invariably younger than the finer-grained veins (or pegmatites). This observation is believed to indicate that the coarser rocks formed at lower temperatures than the finer (p.258).

Temperature of formation of veins.

The 'late layering' described above has very probably formed by metamorphic processes. It is therefore referred to as metamorphic layering. If Challis's (1965a) interpretation of textural and fabric criteria is correct some layering of the Red Hill Complex is formed by crystal sedimentation. Such layering may be described as 'Primary Layering' in the sense of Thayer (1963).

Veins and pegmatites cut both the Primary Layering (Challis, 1965a) and the metamorphic layering (e.g. fig. 91). Also it is very probable that metamorphic layering itself is formed by multiple vein development (p.149). Therefore estimation of the temperature of vein formation indicates the temperature of formation of metamorphic layering.

Most of these veins are probably of replacement origin. The replacement process presumably required water to act as transporting agent, but the minerals talc and serpentine have not generally developed and therefore temperature of vein formation and hence metamorphic layering is probably in excess of 650°C. (see fig. 2, Bowen and Tuttle, 1949). Some monomineralic feldspathic veins cutting harzburgite have developed a reaction zone (including talc) against enclosing peridotites. Estimation of their temperature of emplacement is possible from a study of the reaction zone assemblage.

Anorthosite vein. Cutting massive and weakly foliated harzburgite near Chrome, is a three inch wide vein of anorthosite, at the margin of which is an inch wide reaction zone separating the feldspar from the harzburgite. A similar vein cutting layered rocks at 382885 is shown in figure 87.

The vein has a slightly sinuous form and was traced 120 feet before thinning and lensing out. The centre of the vein has a medium-graded aplitic texture but near the margin, the feldspar is coarse-grained, with some grains up to 2cm. long. The reaction zone (No. 10956) is composed of talc, chlorite and amphibole, the optical properties of which are given below:

Chlorite $N_z = 1.597$, $Nx = 1.592$ Ny estimated 1.593

Birefringence about .005

(+ve) $2V = 8^{\circ}$.

X ray diffraction pattern gives a 14Å peak

No anomalous interference colours.

The optical data indicates that the chlorite is an Fe-rich clinochlore according to the classification of Hey (1954), or a Mg-rich chlorite according to Albee (1962).

Amphibole $N_z = 1.642$ $Ny = 1.632$ $Nx = 1.623$

(-ve.) 2V about 80°.

$Z\hat{}c = 19^{\circ}$. Colourless.

The lack of colour and optical properties indicate that the amphibole is a tremolite (Tr_{80}).

In thin section the talc and tremolite are intimately mixed and replace olivine and enstatite of the country rock, and the chlorite occurs as a zone separating the feldspar from the talc-tremolite assemblage. Nowhere is feldspar in direct contact with anything other than chlorite. The necessary calcium of the amphibole and the aluminium necessary for the chlorite is probably derived from the feldspar. Thus the reaction zone is believed to have formed in the manner indicated in the following reaction.

Olivine + enstatite + anorthite + water = talc + amphibole + clinochlore.

The temperature during development of the assemblage on the right of the equation may be approximately determined, assuming a load pressure of about 1000 bars estimated from the thickness of the super-incumbent load. Serpentine is not present in the assemblage and therefore the temperature was probably greater than its upper stability limit of 500°C. (Bowen and Tuttle, 1949).

A natural assemblage of clinochlore-talc-tremolite has been described by Durrell (1940) in contact metamorphosed ultramafic rocks surrounding a quartz-monzonite intrusion in the Sierra Nevada of California. (Recorded in Turner and Verhoogen, 1960, p.516.) The ultramafic rocks are interbedded with metamorphosed basic rocks of the hornblende-hornfels facies which is correlated with temperatures between 550 and 700°C. by Turner and Verhoogen (1960).

It is probable therefore that the anorthosite vein was formed at a temperature of about 600°C. As the anorthite was presumably derived from the rocks of the Complex this is also the temperature of the country rock at the time of the vein formation.

The possibility that the veins and metamorphic layering formed by locally developed high temperatures in otherwise cold peridotite can be discounted for the following reason.

Reaction zones, such as that surrounding the anorthite vein described above are very rare. Yet if all veins and metamorphic layering were formed by intrusion of hot aqueous solution, reaction zones including talc or even serpentine should be universal. It is therefore concluded that during development of the metamorphic layering the ultramafic rocks as a whole were at a temperature higher than that of talc formation, i.e. greater than 650°C.

FABRIC

Introduction

The fabric of peridotites has been extensively studied in recent years and a review of the subject and considerable new information is given by Collee (1963). Most writers have described fabric in terms of the orientation pattern of olivine. This is the method used here.

Olivine in peridotite nodules, in layered ultramafic intrusions such as Rhum and Skye, in metamorphic olivinites as well as in Alpine-type peridotite complexes, commonly shows strongly preferred orientation of X (= [010]). In layered or fissile peridotite the X axes are usually concentrated in a maximum normal to layering, but short girdles have also been recorded. Z(= [100]) and Y (= [001]) axes do not always show strongly preferred orientation but single maxima do occur in peridotites of many different origins, e.g. Turner (1942) and Ladurner (1956) describe Y maxima parallel to weak '<u>b</u>' lineation inferred to be of tectonic origin; Raleigh (1963) showed that Z axes tend to lie parallel to fold axes in banded dunites from an Alpine type peridotite, and in peridotite nodules Z maxima are sometimes the most strongly developed features of the olivine orientation pattern (Collee, 1963 p. 13). Thus the characteristic orientation pattern of olivine - a single X maximum perpendicular to layering and the Z and Y axes concentrated in individual maxima or partly developed girdles - is common to a large variety of peridotites.

The actual process by which the preferred orientation of olivine is achieved, is the subject of much discussion. At present there are two main divisions of opinion on the orientation process which may be called the magmatic and tectonic hypotheses. Brown (1956) and Brothers (1964) consider that the preferred orientation of olivine in some peridotites (peridotite nodules and layered peridotites of Rhum) is due to dimensional orientation of suspended olivine crystals either by crystal-settling or by laminar-flow of magma. Other writers (e.g., Turner, 1942; Ladurner, 1956; and Collee, 1963) regard the preferred orientation of olivine as a tectonite fabric formed in an analagous manner to the better understood fabric of calcite or quartz bearing rocks.

The intersecting planar structures of the Red Hill Complex provide a unique opportunity for examining the development of preferred orientation of olivine, and it will be shown that the characteristic orientation pattern of olivine can, in fact, originate quite unambiguously by processes operating in the solid state.

Two writers have recently studied the olivine fabric of peridotites from the Nelson Ultramafic Belt.

Battey (1960) showed that in a number of specimens collected from Dun Mountain the Z olivine axes tended to lie parallel to the fold axes in the adjacent Maitai Group. X axes in the different diagrams tended to lie at some point in a girdle normal to Z, suggesting that the various specimens were rotated about the Z axis.

Challis (1965a) has described the fabric of layered peridotites from the Red Hill Complex. All specimens have a X maxima perpendicular to the plane of layering but Z maxima in adjacent layers do not show consistent direction of orientation within the plane of layering. Challis interprets the data in terms of a crystal sedimentation hypothesis for the origin of the layering.

Procedure. Thin sections were cut from selected, oriented specimens and the orientation of at least 50 and usually 60 olivine grains was measured for each diagram. Except for one specimen (No. 10967, and fig. 67) all diagrams are presented with the primitive circle horizontal and north at the top of the page. All diagrams are a lower hemisphere plot on an equal area net. The indication marks at the base of each diagram point to true south. The attitude of the principle mesoscopic structures is indicated in each diagram.

Collee (1963) has discussed the sampling errors involved in the measurement of orientation of olivine axes and in the construction of the orientation diagrams. He shows that in rocks with strongly developed dimensional orientation of minerals or those with inequigranular textures sampling errors may result in incorrect representation of the preferred orientation, and suggests that the errors may be reduced by avoiding double measurement of grains and selecting grains along lines rather than by random sampling. That quite significant errors may result, is evident in some examples given by him.

Although care was taken to minimise sampling errors in the present study errors are bound to arise where only 50 or 60 grains can be measured (suggested by Turner (1942) to be adequate and usually the maximum number available in most thin sections studied by the writer) and the rocks in many cases are strongly inequigranular in texture. Nevertheless in one inequigranular fine grained specimen (No. 10987) two lots of 40 grains each, were independently measured and plotted on a diagram. The two diagrams differed only in trivial detail. For this reason fine detail in the following diagrams is ignored and only the gross features of the pattern are regarded as significant.

The fabric diagrams

Fabric and layering. The olivine orientation pattern of an example (No. 10988) of layered peridotite is given in fig. 65. The outcrop from which the specimen was taken is shown in figure 49, p.160. Each olivine axis shows strongly preferred orientation and the X maximum is normal to layering.

Fabric and lineation. Preferred orientation diagrams of olivine in a harzburgite (specimen No. 10967) are given in figure 67. The rock has a partly recrystallised protoclastic texture and ortho-pyroxene grains have strongly developed (100) lamellae. (See p.149). The specimen, and the primitive circle of the fabric diagrams is approximately normal to the lineation. Figure 67d shows the orientation of 20 large grains of orthopyroxene. The lineation is parallel to c-axis of orthopyroxene and Z axes olivine. The X and Y axes of olivine are also concentrated in single maxima.

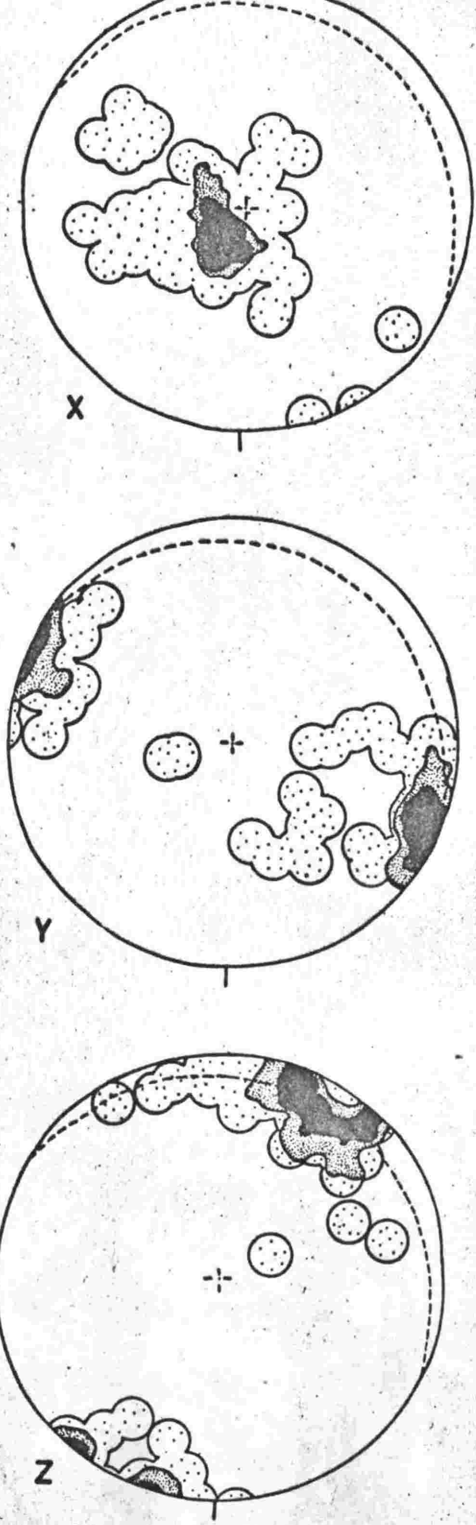


Fig. 65. Spec.no. 10988
(60 points)

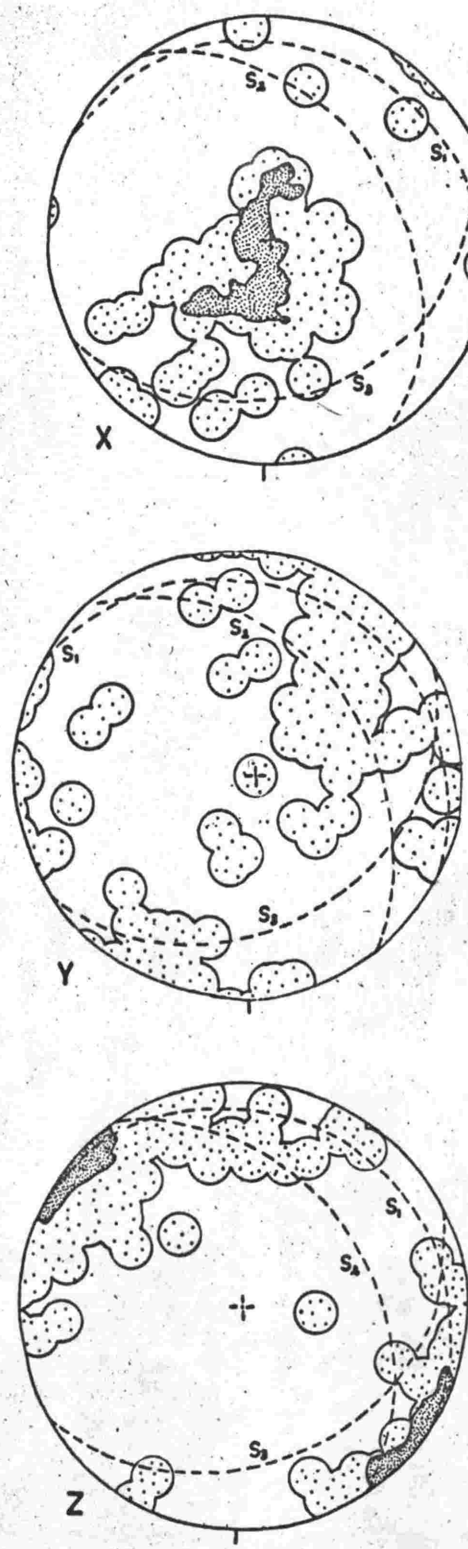


Fig. 66 Spec.no. 10989
(60 points)

Contours: 10%, 5% and 1 point per 1% area.

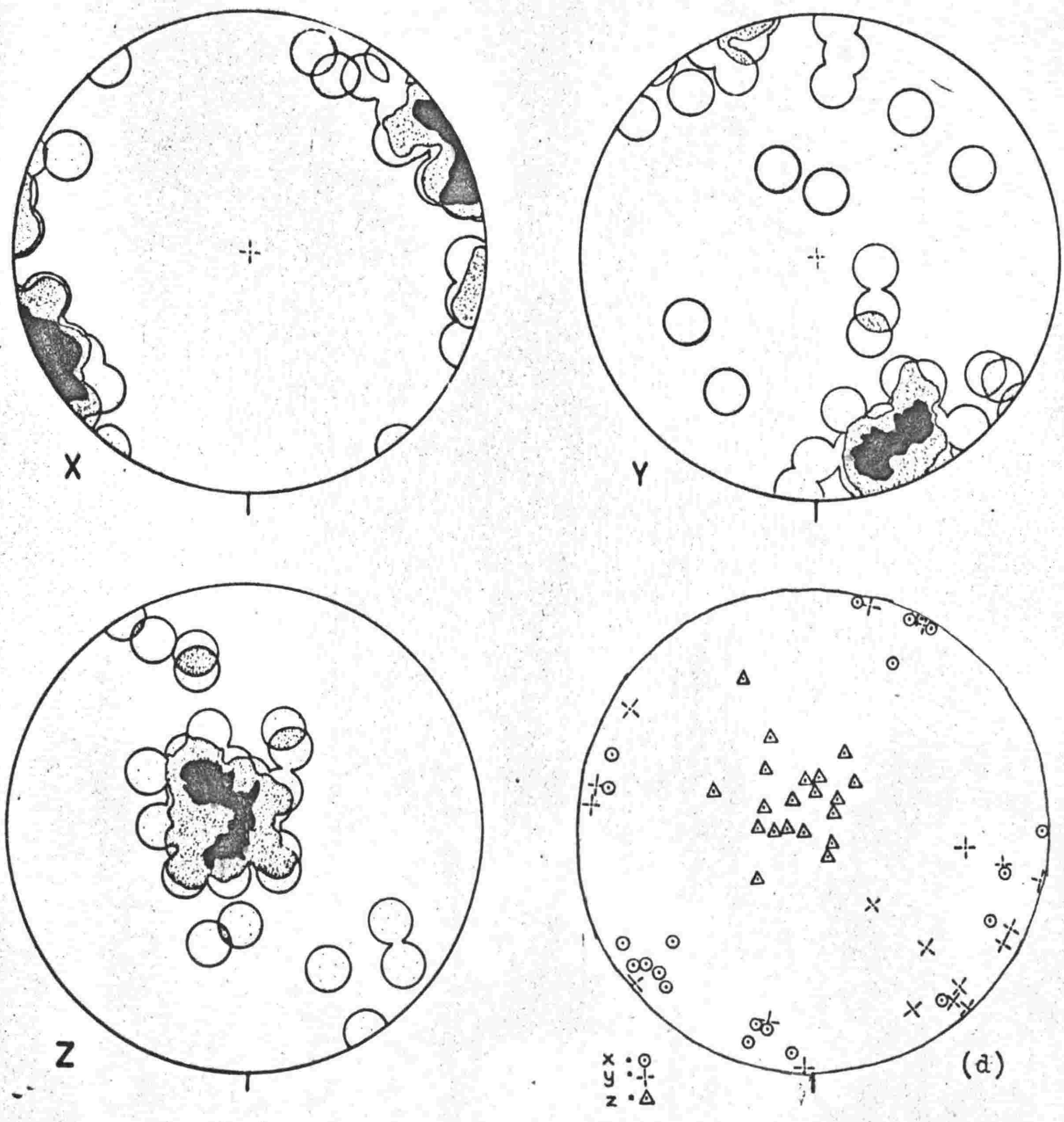


Fig. 67. Spec. no. 10967. Lineation vertical.

X, Y and Z show orientation pattern of olivine. (50 points).

(d) shows orientation of orthopyroxene. (20 points).

Contours: 10%, 4% and 1 point per 1% area.

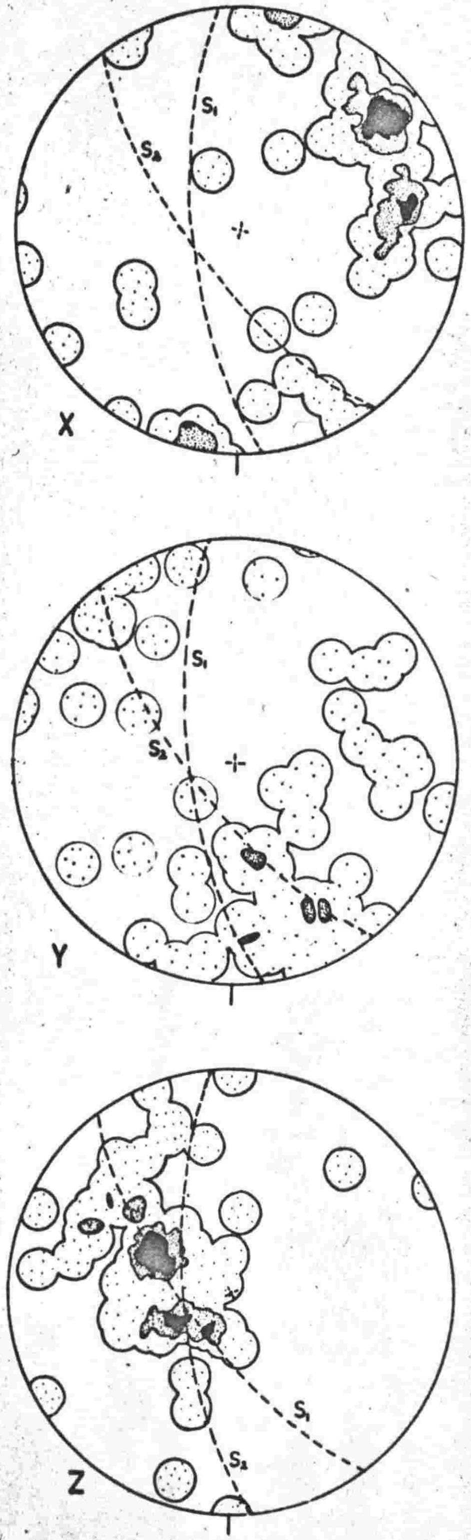


Fig. 68. Spec.no. 10987
(80 points)

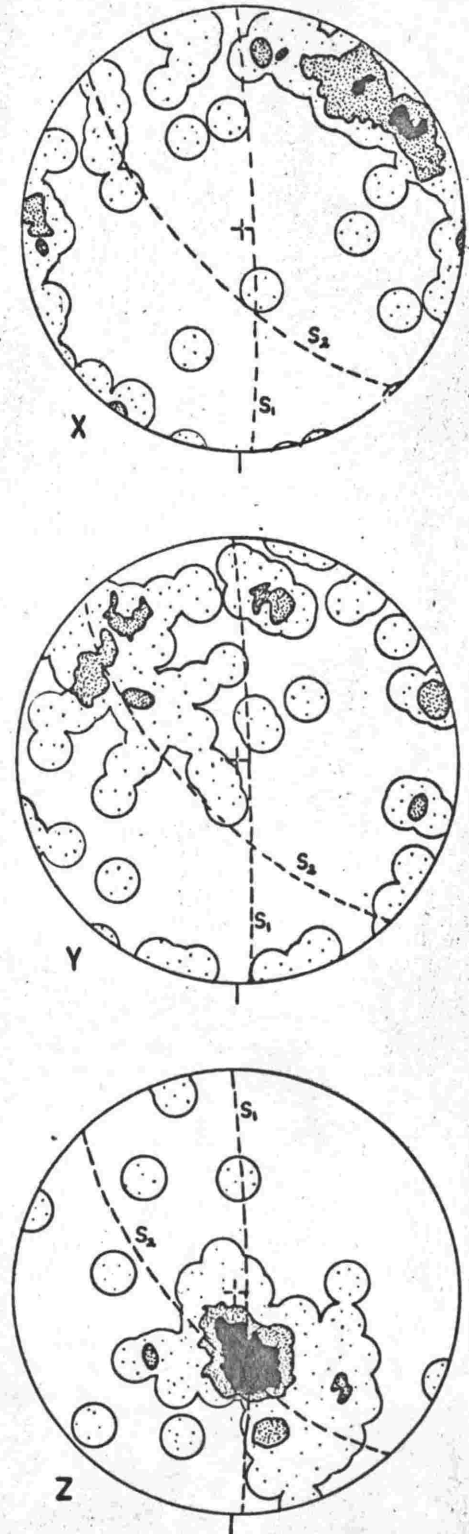


Fig. 69. Spec. no. 10990.
(60 points)

Contours: 10%, 5% and 1 point per 1% area.

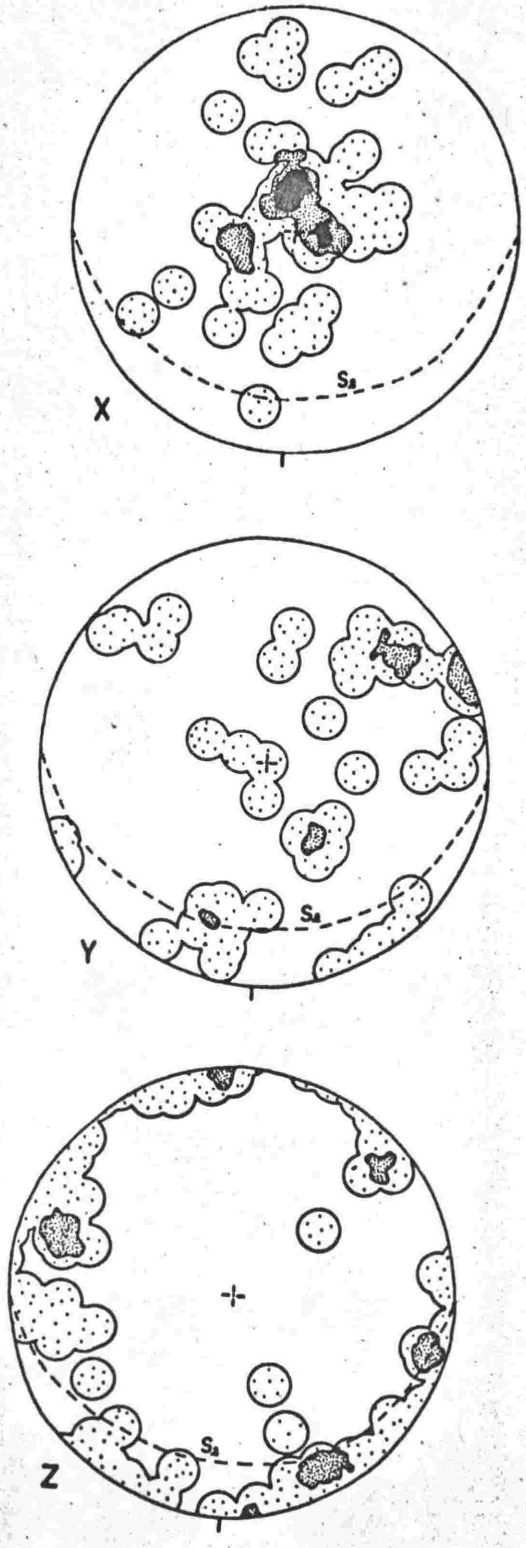


Fig. 70. Spec.no.10991
(50 points)

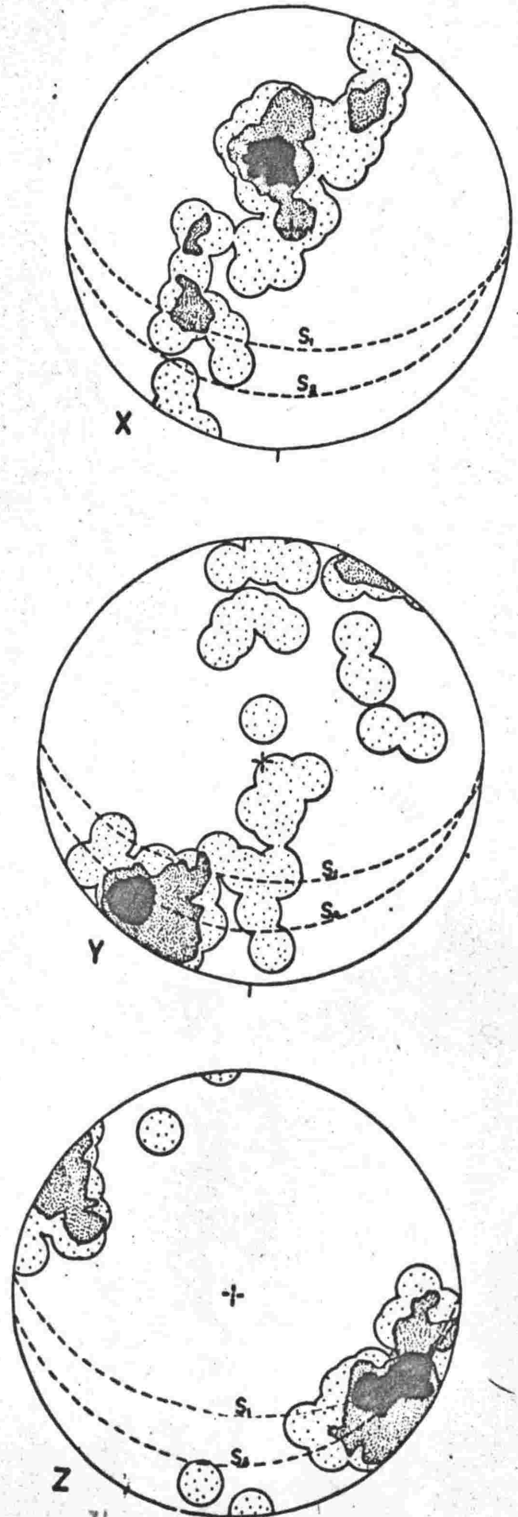


Fig. 71 Spec. no. 10992
(50 points)

Contours: 10%, 4% and 1 point per 1% area.

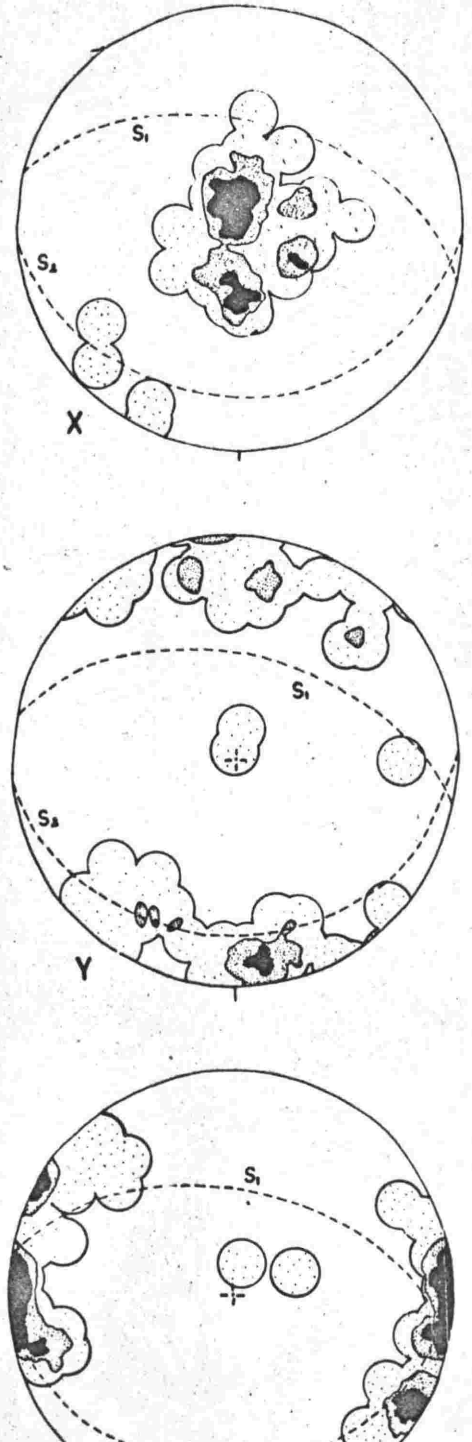


Fig. 72. Spec.no.10993
(60 points)

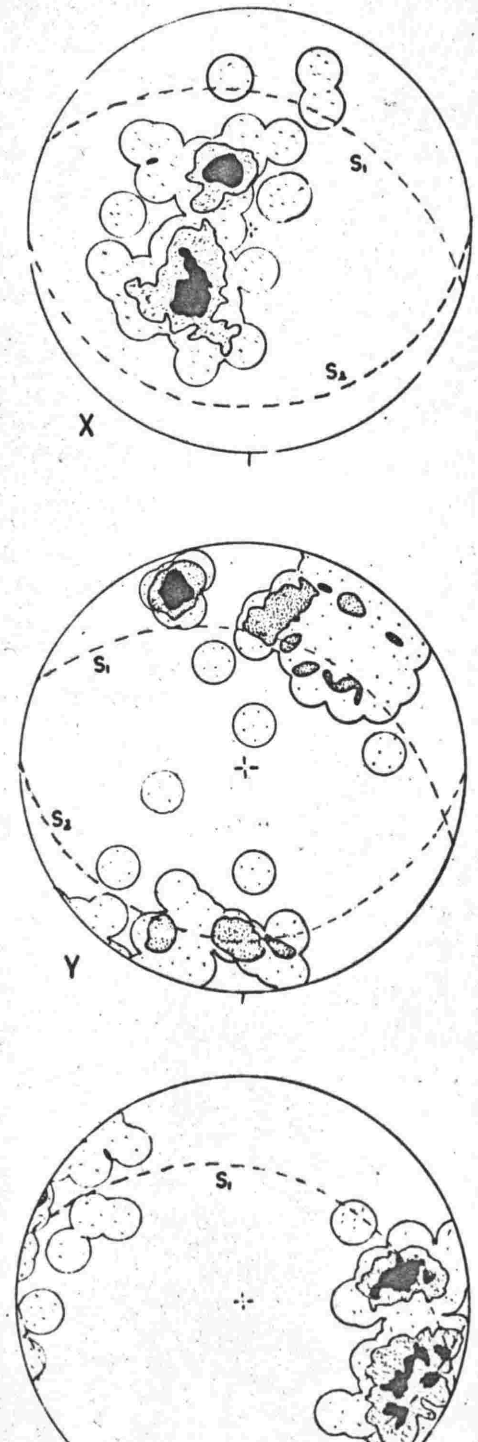


Fig.73. Spec.no.10994
(60 points)

Contours: 10%, 5% and 1 point per 1% area.

Fabric and foliation. Two specimens of foliated peridotite were studied.

(i) Specimen No. 10989 is from the outcrop shown in figure 85, and was collected from the lower left-hand corner of the exposed face, but in hand specimen only the foliation can be observed. The mesoscopic structures present in the outcrop are an early layering (S_1) foliation (S_2) and a dunite dyke (S_3). Z axes (fig. 66) are strongly concentrated near the intersection of S_1 and S_2. X axes lie in a partly developed girdle with a broad maxima approximately normal to both S_1 and S_2. The Y axes are much more scattered, but tend to be concentrated near S_1 or S_2.

(ii) The preferred orientation diagrams of a specimen (No. 10987) of peridotite containing a diopside vein (S_1) and divergent feldspathic foliation (S_2) from the outcrop illustrated in figure 44 are given in figure 68. The Z axes are concentrated in two maxima, one at the intersection of S_1 and S_2, and the other on S_1, 25° from the intersection. X axes also lie in two well defined maxima, one normal to S_1 and the other normal to S_2. Y axes show a much more diffuse concentration but in general, lie near S_1 and S_2.

Fabric of dunite dykes. Specimen No. 10990 is from a dyke of dunite cutting foliated feldspathic harzburgite. The order of dunite dyke and foliation cannot be determined unambiguously. The foliation may have developed before or after formation of the dunite dyke.

The fabric diagrams are given in figure 69. Z axes are strongly concentrated at the intersection of S_1 (dunite dyke) and S_2 (foliation) in a single maximum. X axes lie in a girdle with strongest maxima normal to foliation. A weak concentration lies normal to the dunite dyke. As before, Y axes show less strong concentration but tend to lie in a girdle.

Fabric and two sets of cross-cutting layers. Fabric diagrams of four specimens collected from the outcrop near Porter's Knob, illustrated in figure 74 were prepared (see also illustrations figs. 91-3 p. 250-52). A description and discussion of the features shown in this outcrop is given on page 248. The folded, early layering is designated S_1 and the crosscutting veins and replacement dykes constituting the second set of layering is designated S_2. Two specimens from the early layering, and two from the later dunites were studied, and their positions in the outcrop are shown in figure 74.

(i) Specimen No. 10991 is from the thick, S_2 layer near the top of the outcrop. The X axes (fig. 70) are concentrated in a maximum normal to S_z, Z axes tend to lie in a broad girdle, and Y axes show no significant concentration.

(ii) Specimen No. 10992 is from the central dunite S_2 vein and preferred orientation diagrams are given in figure 71. X axes lie in a girdle normal to the intersection of S_1 and S_2 the strongest concentration at the pole of S_2. Y axes form a less well defined girdle with the strongest concentration (25 per cent of axes fall within the maximum) on S_2. Z axes are concentrated in a single maximum at the intersection of S_1 and S_2.

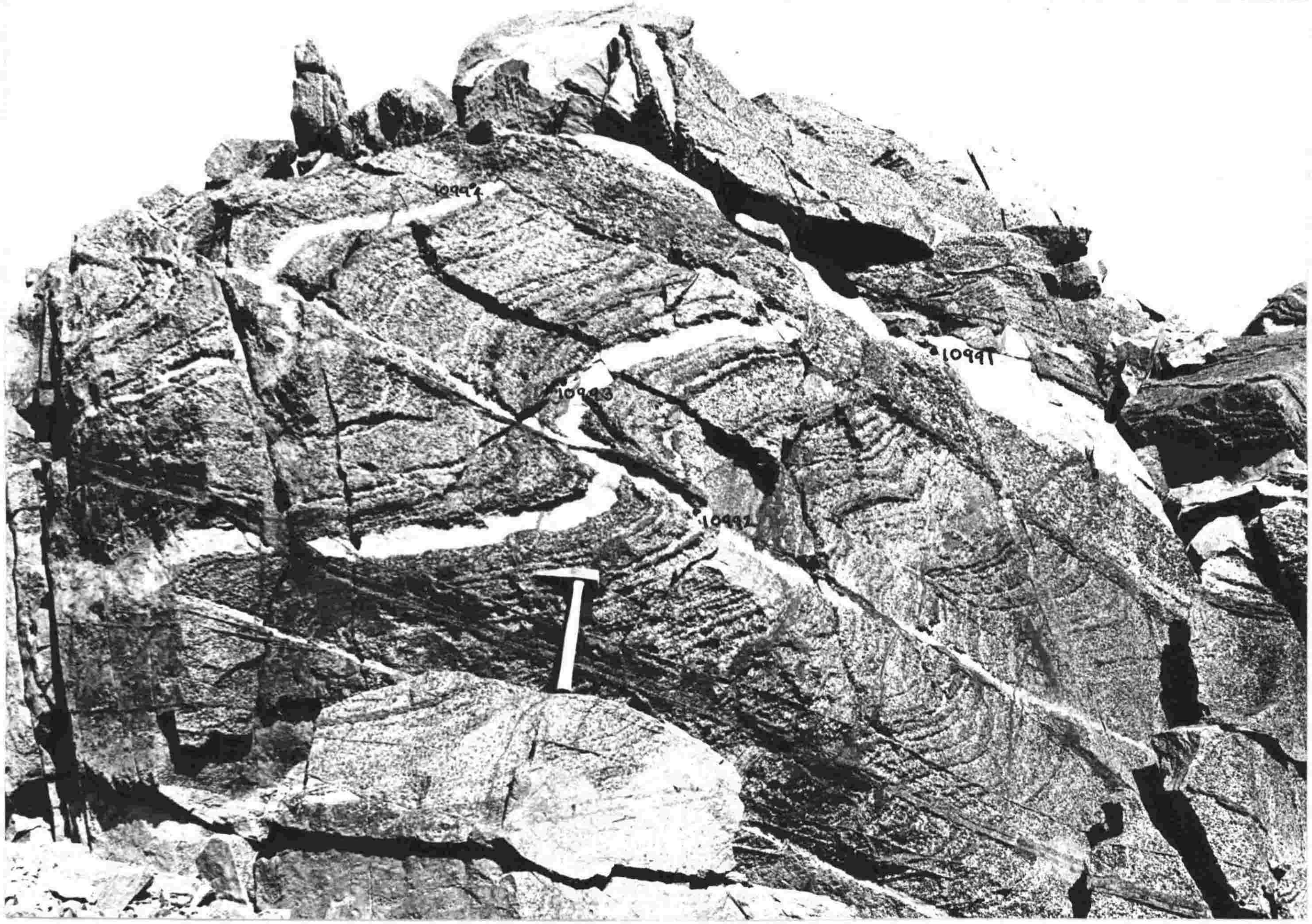


Fig. 74. Porters Knob outcrop. Position of petrofabric specimens shown. See also figures 91, 92 and 93.

(iii) Specimen No. 10993 was collected as far away from mesoscopically evident S_2 structures as was possible in the outcrop, and in hand specimen only S_1 layering can be seen. Fabric diagrams (fig. 72) show Z axes lie in a part girdle on S_2, X axes from two maxima, one near the pole of S_2, and the other about 45° from S_2 and 60° from S_1, Y axes tend to be concentrated near the north and south edges of the diagram.

(iv) Specimen No. 10994 was collected from within 6 inches of the thick S_2 dunite-harzburgite band near the top of the outcrop. Z axes (fig. 73) are concentrated near the intersection of S_1 and S_2 and form two maxima, one on S_1 and the other on S_2. X axes form a short girdle containing two maxima, one normal to S_1 and the other normal to S_2. Y also forms two maxima; one a well-defined point maximum on S_1 and the other a broad band between S_1 and S_2.

Interpretation of Fabric Diagrams

Tectonic origin of preferred orientation of olivine is clearly indicated in the following specimens.

(i) Specimen Nos. 10991 and 10992 from the Porters Knob outcrop. The central dunite vein (No. 10992) is very probably of replacement origin as remnants of S_1 can be traced across it showing neither dilation nor offset but the orientation of the olivine shows a well defined X maximum normal to S_2 and a weak concentration normal to S_1. Such preferred orientation of X normal to S_2 must be related to processes operating in the solid state. Specimen No. 10991 from the thick layer near the top of the outcrop also shows preferred orientation of olivine, X normal to S_2. As this layer is also probably of replacement origin (see section on petrogenesis, structural development) this fabric too, has developed in the solid state.

(ii) The lineated harzburgite (specimen No. 10967) also shows strongly preferred orientation of olivine. As the texture is protoclastic and bent lamellae etc. (p.132) indicates that the rock has suffered considerable deformation it is most unlikely that the fabric can have been inherited from a pre-deformational period and has therefore developed during or after deformation, again presumably in the solid state.

<u>The orientation process of olivine.</u> The close relationship between mesoscopic structures and fabric is clearly indicated in the diagrams. Specimens from outcrops containing a single mesoscopic structure, layering or lineation, have a fabric in which the three axes of olivine are concentrated in mutually perpendicular maxima, with X perpendicular to layering and Z lying parallel to lineation. A similar but more complicated fabric is shown in specimens from outcrops containing two intersecting S planes. Double X maxima are commonly present (Specimens Nos. 10989, 10987, 10990 and 10994) each near the pole of one S plane. Some specimens with double X maxima also have double Y and Z maxima and it is possible that the fabric of these specimens is controlled by two independent groups of olivine grains of which one is related to S_1 and the other to S_2.

To examine this possibility more closely two diagrams of specimen No. 10994 have been prepared showing the orientation of those olivine grains whose X axes lie near the pole of S_1 in figure 75a and near the pole of S_2 in figure 75_b. In the first diagram Z axes are evenly distributed between both maxima and Y axes show two concentrations marked Y_1 and Y_{1a}. In the second diagram the Z axes do not lie in only one of the maxima but are evenly distributed between both. All Y axes however lie on or near S_2. Evidently the fabric diagrams cannot be explained as due to two independent groups of olivine grains; the change in orientation of X from the pole of S_1 to the pole of S_2 has not involved the complete reorientation of all axes of individual grains, but rather partial reorientation involving one or two axes only of a large number of grains. This feature can best be explained in terms of rotation of grains about one crystallographic axis so that one other axis moves to the new position.

On this hypothesis an explanation can be made for the features of the fabric of Specimen No. 10994. The positions marked X_1 Y_1 Z_1 and X_2 Y_2 Z_2 are considered to be the initial (related to S_1) and final (related to S_2) positions of the axes of olivine. Y_{1a} is an intermediate position between Y_1 and Y_2. Some grains rotated about X_1 so that their Z axes fell near Z_2 and Y axes near Y_{1a}. Others rotated about Z_1 and X and Y axes fell near X_2 and Y_2. No rotation about Y_1 appears to have occurred. Following this initial rotation of say Z to Z_2 further rotation has occurred about Z_2 so that X moves from X_1 to X_2 completing the reorientation of the grain, but this complete reorientation has only occurred in a few of the grains.

This type of reorientation is considered to have occurred also in other specimens, explaining the much greater scatter of Y than X or Z axes. Also, it can be inferred that rotation involved almost complete transition rather than gradual rotation from the initial to final

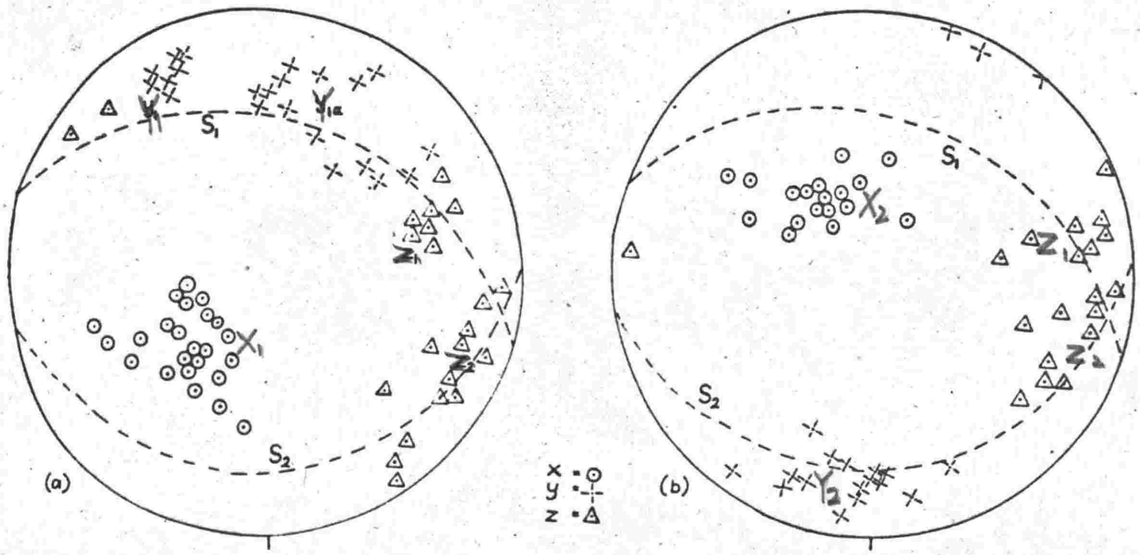


Fig. 75. Spec. no. 10994.

(a) Orientation of olivine grains whose X-axes lie near the pole of S_1.

(b) Grains whose X-axes lie near the pole of S_2.

position of axes for otherwise a part girdle rather than individual maxima would be apparent in the fabric diagram. It should be noted that the rotation is apparent only; The concept is merely a semantic convenience. Non-componental movements such as recrystallisation may give the observed features of the fabric diagrams as much as componental movements. It is possible for instance that grains with a suitably oriented axis may, under conditions of anisotropic stress, grow at the expense of less favoured crystals. In fact the strongly developed fabric of the S_2 dunite layer in the middle of the Porters Knob outcrop is better explained by recrystallisation than strain. Although it is evident that the grains in that part of the outcrop have been rotated there is no evidence of translation. Yet considerable translation would be required to cause rotation by discrete amounts of slip along intragranular slip planes.

Crystal sedimentation hypothesis and fabric.

A number of writers have interpreted the olivine fabric in peridotites in terms of the crystal sedimentation hypothesis. X maxima are ascribed to crystal settling on a broad (010) face. Z maxima, when present, are thought to be due to alignment of elongate crystals by flow of the magma. Challis (1965a) has cited as evidence of crystal sedimentation the fact that Z maxima in adjacent layers do not show parallel orientation. Such a fact is explained by the crystal sedimentation hypothesis as due to change in direction of magma flow. Challis, argues that hypothesis of tectonic deformation for the origin of fabric does not explain this observation.

The fabric of specimens from the S_2 layers of the Porters Knob outcrop do not support Challis's conclusion. Both show X maxima perpendicular to S_2 layering (No. 10991 and 10992) but the orientation of Z in both differs considerably. In 10992 there is a clearly defined Z maxima approximating E - W but in 10991 there is no clearly defined Z maxima and certainly no significant orientation E - W. The two diagrams cited here are comparable to those given by Challis (1965a). Only if the S_2 layers in the Porters Knob outcrop can be interpreted as formed by crystal sedimentation is it possible to interpret Challis's observation as evidence of crystal sedimentation. The evidence clearly shown in the photographs of the S_2 layering (p.250) does not support the hypothesis of crystal sedimentation for these layers.

Relationship between layering, foliation and veins. An important feature of the relationship between fabric and structure is the close similarity of preferred orientation of olivine in specimens from outcrops containing intersecting planar structures, regardless of whether these are foliation, veins or layering. It is inferred from this that the structures are closely related in origin and probably formed, like fabric, by anisotropic stress.

Intersection of S_1 and S_2. The line of intersection of S_1 and S_2 in outcrops containing two intersecting planar structures does not show consistent orientation over the whole of the Complex or even in a small area of it. This is clearly shown in the map of Porters Knob (Map II).

The early layering shows no consistent pattern of orientation but the late (S_2) layering, east of the hinge of the Porters Knob Antiform, is parallel throughout the area. The folding of the early layering therefore appears to be due to some earlier and unrelated period(s) of deformation. From this is can be seen that little tectonic significance can be attached to the direction of the line of intersection.

SERPENTINITES.

TECTONIC INCLUSIONS AND RODINGITES

Post-emplacement hydrothermal activity is an important feature of many Alpine-type complexes manifested by the development of serpentinite from peridotite, metasomatism of sedimentary and igneous rocks in contact with the complexes and the development of rodingites. Several other hydrothermal effects also occur but these three are the most conspicuous.

SERPENTINITES

DESCRIPTION

The relative amounts of serpentine and unaltered olivine and pyroxene in ultramafic rocks can be approximately determined in the field from their density, colour and appearance in hand specimen. Consequently, it is possible to determine the direction of increase in degree of serpentinization. This is towards the margin as well as towards mafic dykes and shear zones of the peridotite. The ultramafic rocks of the margin of the Complex are generally entirely serpentinized. Completely serpentinized rocks occur within 200 feet of the contact, and from there the degree of serpentinization decreases rapidly inwards, falling to zero at about 500 feet to 1000 feet from the contact. The width of this partly serpentinized zone is irregular. It is thicker on the western (1000 feet) than on the eastern and northern contacts (500 feet).

The more basic mafic dykes cut unserpentinized peridotites, but the hornblende dykes invariably cut at least partly serpentinized rocks, and commonly the contact peridotite is completely serpentinized. Where hornblende-bearing dykes are widely separated, the degree of serpentinization increases towards the dykes and the width of partial-serpentinization is only about 20 feet for a 5 foot dyke, and correspondingly greater for thicker dykes. Within the northwesterly trending belt which contains most of the dykes (p.95) all the peridotite is partly serpentinized, and the degree of serpentinization increases towards the dykes.

The relationship of serpentinization to shear zones is very obvious. All rocks within 10 feet to 50 feet on either side of the shear zone are at least partly serpentinized. Many of the shear zones are evidently faults developed during regional faulting (p.227) long after intrusion of the peridotite.

SERPENTINIZATION

As the assemblage olivine-enstatite-water is unstable at temperatures below 650°C. (Bowen and Tuttle, 1949) it may be inferred that since emplacement of the Complex, wherever water was available for reaction, the olivine and enstatite will have altered to talc or serpentine, and the persistence of unaltered peridotite probably means that for some reason water was not available for reaction.

As a considerable proportion of the ultramafic rocks of the Red Hill Complex are partly or wholly serpentinized a large amount of water must have been required. Hess (1933) suggested that the water causing serpentinization in many Alpine-type peridotites was derived from within the peridotite itself, but other writers consider that much of the water was derived from the surrounding sediments (Turner and Verhoogen, 1960, p. 320). The serpentinization of ultramafic rocks in the Red Hill Complex is believed to have been effected by water derived from both sources as well as from the intrusive mafic dykes.

(a) Water derived from within the Complex. The development of veins, replacement bodies and metamorphic layering requires a large amount of water. Much of the water may have escaped during emplacement, but some would be retained and with falling temperature would have reacted with the anhydrous minerals to form serpentine, talc etc. Most of the serpentinization due to this water would probably occur at, or near the contacts of the Complex. This follows from consideration of the temperature gradient during cooling. As soon as temperatures dropped below a critical value, i.e. around 650°C. water would react to form hydrous phases. The critical temperature would be reached first near the contact of the Complex and water would flow to the contact to be converted into the hydrous minerals serpentine or talc. If talc had formed first, further reaction at lower temperatures between olivine, talc, and water would have produced serpentine. Some water may have remained within the body of the Complex, but it is not likely to have been much. The few examples of development of small amounts of talc and serpentine in otherwise fresh peridotite well away from the contact, dykes or shear zone, may be explained as caused by this residual water.

(b) Externally derived water. The very large amounts of water required to completely serpentinize the minor ultramafic bodies of the Goat Formation, was probably derived from outside the body. There is no reason to suppose that these ultramafic bodies contained a greater proportion of water than the Complex and there was certainly not enough water in the latter to cause complete serpentinization. Also, the serpentinization along the shear zones in the Complex must be derived from an external source. Assuming an unlimited supply of water in the rocks surrounding the Complex, it is surprising that all ultramafic rocks have not been serpentinized. In fact, the width of the serpentinized zone is very narrow in comparison with the size of the Complex. This is especially surprising because the peridotites are very well jointed and in many cases, strongly fractured. The probable explanation is that any crack wide enough to permit ingress of water would be closed because of the volume increase of serpentinization. For this reason, peridotites are probably highly impervious rocks, and water may only get inside by fracturing the serpentinite sheath or by slow diffusion.

(c) Water from the Mafic Dykes. The mafic dykes are believed to have intruded the Complex over a long period commencing when it was still fairly hot, i.e. at about 500 - 600°C. (p.259). Some evidently contained a fairly high proportion of water, because of the crystallization of amphibole rather than pyroxene, but the amount of serpentinization is much greater than can be expected if water was derived solely from the dykes. Probably the intrusion of the dykes allowed additional water from surrounding strata to get inside the protective sheath surrounding the Complex. The wide distribution of partly serpentinized rocks related to the mafic dykes is explained by the high temperature of the peridotites permitting rapid diffusion of water away from the vicinity of the dykes.

O t h e r h y d r o u s p h a s e s

Serpentine is not the only mineral to form by hydration of the peridotite though it is by far the most common. Reference has already been made to talc, which occurs only in pyroxene-bearing peridotites, presumably because of the higher silica content of those rocks. Talc is found only in thin section, and does not form any mesoscopically visible deposits.

Two hydration products of feldspar have been noted. Feldspar is often the only mineral to alter in some peridotites. In some specimens (e.g. No. 10995) the feldspar grains have altered to a turbid, anisotropic material, identified by X-ray diffraction studies as hydrogrossular, which contains slightly less than 1 molecule of water in the lattice (a = 11.90Å), (Yoder, 1950).

Another alteration product is zoisite. A small part of the feldspar grain in specimen No. 10996 is replaced by a finely-granular material which has weak to moderate birefringence and refractive index higher than olivine. Only a very small amount could be separated from the rock and by the oil immersion method gave a refractive index of 1.695. Consideration of likely chemical compositions, refractive index and birefringence strongly suggests that the mineral is zoisite.

Zoisite is found only in a few thin sections of peridotites as very small grains and is commonly associated with hydrogrossular.

T E C T O N I C   I N C L U S I O N S

Numerous disorientated exotic blocks of rock are found in the crushed serpentinite at the margin of the Complex. These are derived from the adjacent strata or mafic dykes and following Brothers (1954) and Coleman (1961) are referred to as 'tectonic inclusions'. Coleman (in press) described in detail the mineralogy and petrology of tectonic inclusions from N.Z. ultramafic bodies including Dun Mountain and Red Hill. Because of Coleman's detailed work only brief mention of the tectonic inclusions is made here.

Coleman (in press) considered that the tectonic inclusions are metasomatised exogenous blocks derived from three different types of rocks; gabbro-basalt, argillite-greywackes, and limestone. The tectonic inclusions of each type have distinctive mineral assemblages developed in zones surrounding a central unaltered or partly altered core. The gabbro-basalt type have an assemblage hydrogrossular-chlorite surrounded by a skin of chlorite; the argillite-greywacke type, an assemblage of albite-actinolite surrounded by an outer zone of actinolite; and the limestone type have outer zones of the assemblage hydrogrossular-prehnite-wollastonite.

As the cores of the zoned tectonic inclusions are commonly

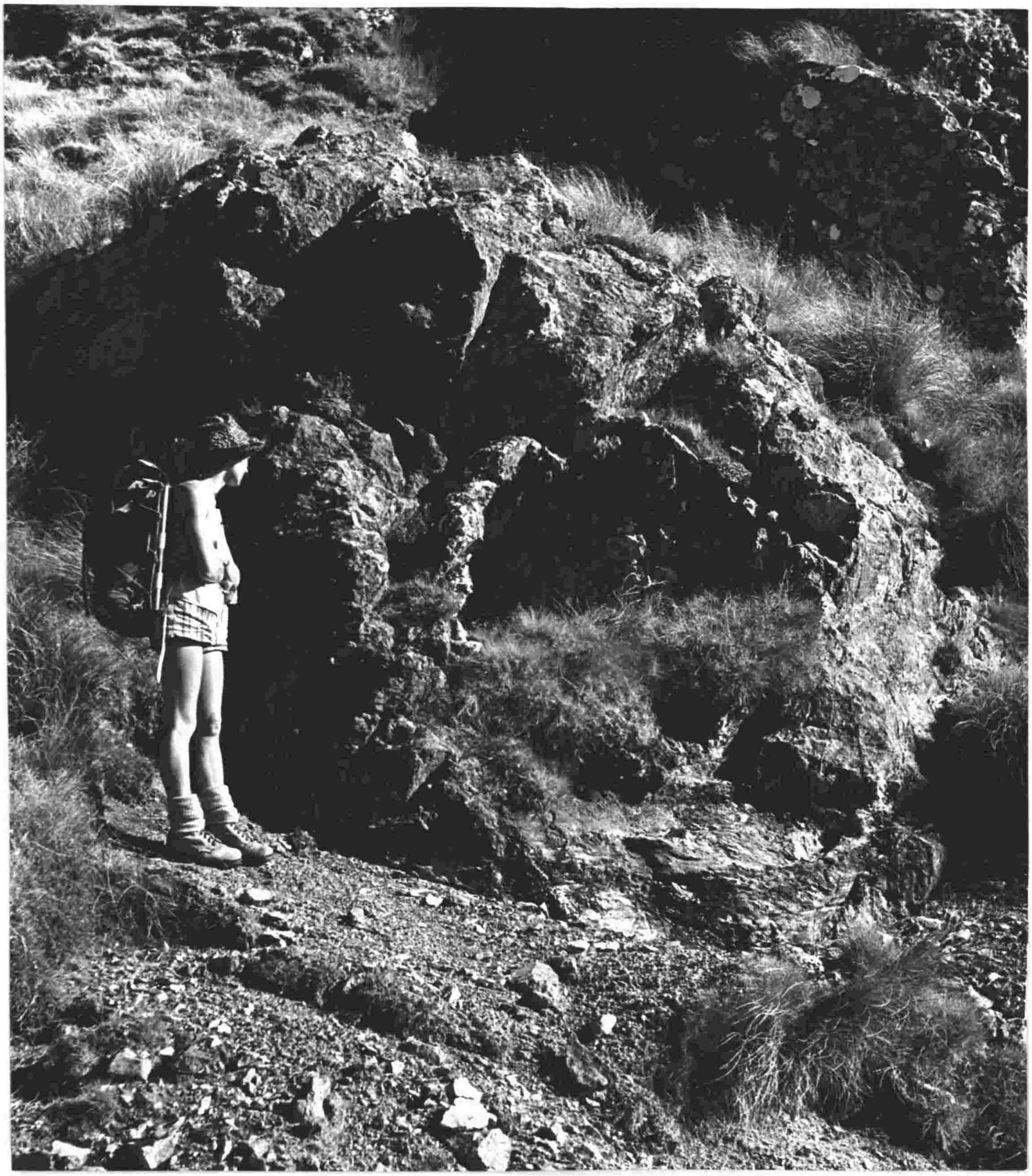


Fig. 76. Type A Tectonic Inclusion. The inclusion is cylindrical with a diameter of about 10 feet, and is surrounded by crushed serpentinite at the margin of the Complex at 420898.

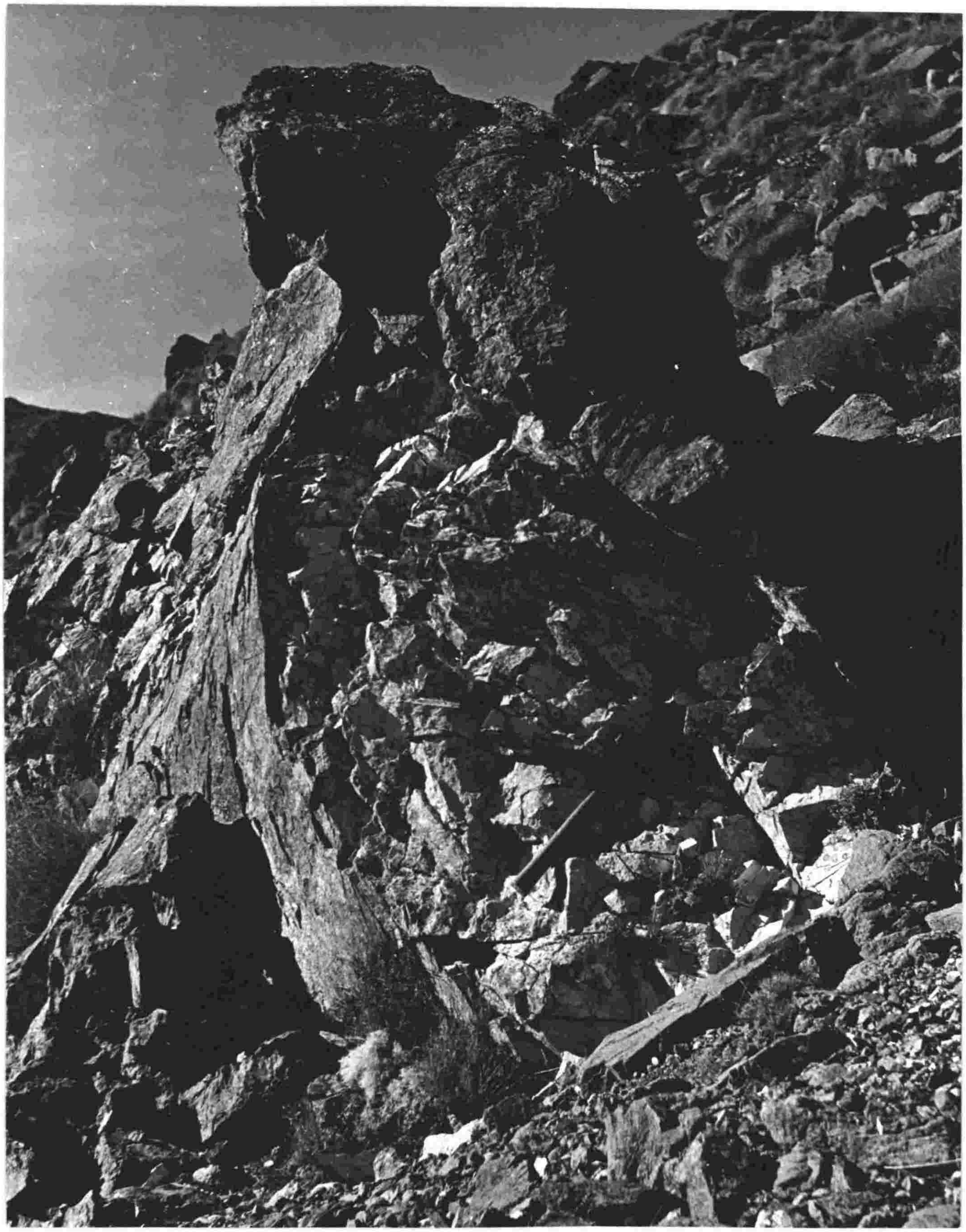


Fig. 77. Type B Tectonic Inclusion. The inclusion is unzoned and is composed of a hard and brittle leucocratic rock. It is surrounded by crushed serpentinite of the margin of the Complex (at 418898).

unaltered, Coleman pointed out the mineral assemblages cannot have been formed by metamorphism but rather are metasomatic, formed by reaction between the core and serpentinite surrounding the inclusion.

In the Red Hill area, a number of tectonic inclusions were examined by the writer; all fall into one of three types. Limestone blocks were not found as tectonic inclusions.

A. Core of altered gabbro, surrounded by a sheath of actinolite-prehnite (fig. 76).

B. Core of hydrogrossular - chlorite - diopsidic-pyroxene (fig. 77).

C. Core of albite - actinolite.

The mineral zoning of Type A. has a skin of dark-green, almost translucent antigorite ($N_z = 1.570$) surrounding a leucocratic sheath composed of prehnite and actinolite. Thin section No. 10997 is from the leucocratic sheath of a tectonic inclusion (Type A) at 419899. Prehnite occurs as sieved, xenoblastic grains about 1mm. across and was identified from x-ray diffraction pattern. It contains inclusions of actinolite (pale green amphibole, $Z \wedge c = 18^{\circ}$, 2V large and negative sign) and a little sphene. The core of the tectonic inclusion (section No. 10998) has a planar fabric of flattened grains of brown-green hornblende. These grains are rimmed by strongly pleochroic fibrous actinolite (iron rich) and are separated by prehnite, albite and hydrogrossular. Altered, zoned plagioclase occurs as grains with hydrogrossular cores and albite rims, and the subhedral habit is still well shown. The rock is petrographically similar to hornblende microgabbroic dykes which cut the ultramafic rocks of the Complex.

Type B. occurs as leucocratic blocks that show little mineral zoning. The core is made up of an assemblage of hydrogrossular-diopside and usually some chlorite. Granular hydrogrossular, at the margin of the inclusion, is disseminated through serpentine (section No. 10999), and increases in amount towards the core. Locally developed thin veins of fine-grained, radiating and parallel, acicular aggregates of prehnite cut the serpentinite.

In the core of the inclusion (thin section No. 11000), hydrogrossular occurs as a turbid matrix enclosing elongate, randomly orientated prisms of colourless pyroxene (fig. 78). Neither chlorite nor serpentine are present. No. 11001 is a thin section from another Type B. tectonic inclusion. Hydrogrossular and pyroxene make up the major part of the section, but chlorite (pale green, very weakly birefringent) replacing pyroxene, is present in minor amount. The pyroxene has a fibrous habit is diopsidic in composition and is weakly pleochroic, with a brownish tint in Z. ($X = 1.691$ $Y = 1.698$ $Z = 1.713$ (all $\pm .003$) $2V = 57^{\circ}$ Z c about 46°). The hydrogrossular has: $a = 11.94$ Å and ri = 1.718 and according to the data of Yoder (1950) is approximately of the composition $3CaO.Al_2O_3$ $2.5SiO_x$ $1\ H_2O$,

Coleman also noted the assemblage hydrogrossular-chlorite-diopside but it is not clear whether he considered the diopside to be relict or authigenic. The fibrous pyroxene in section No. 11000 is thought to be authigenic, and low temperature, authigenic diopside associated with prehnite has been observed by Mr D. Palmer in the Collins River serpentinites (pers. comm.). The alteration of pyroxene to chlorite is thought to be a retrogressive effect, the pyroxene being the

Fig. 78. Fibrous and prismatic authigenic pyroxene (light). surrounded by hydrogrossular (black). Crossed nicols. ($\times$ 400).

Fig. 79. Rodingite veins cutting serpentinised peridotite.

stable mineral developed during low temperature metasomatism of the inclusion.

Type C. tectonic inclusions are widespread on all contacts and are also found in the crush zones that traverse serpentinized peridotite in the north-west of the Complex. A hard tough very fine-grained, blue-coloured argillite outcrop at 388917 is a typical example. In thin section (No. 11002) a mosaic of very fine-grained albite crystals can be seen within which are randomly orientated, acicular prisms of pale green amphibole. This, because of its association with albite and the few optical properties obtainable, (pale green and pleochroic, extinction inclined at between 10° and 20°) is identified as actinolite. Reed (1959) has described several examples of dove-grey argillite which occur as tectonic inclusions and contact rocks at Dun Mountain. He identified albite and tremolite-actinolite as a common assemblage, and gave chemical analyses indicating about 5 to 7 per cent soda in the rock. Reed concluded that the rocks are soda-metasomatised argillaceous rocks.

STRUCTURE AND PETROLOGY

OF THE

RED HILL COMPLEX, NELSON

PART III

STRUCTURE

D E S C R I P T I O N

The Red Hill area comprises part of the steeply dipping eastern limb of the Nelson Syncline, the major structural feature of the Upper Paleozoic rocks of the northern part of the South Island (Wellman, 1956). In the area three separate tectonic belts are recognised, named the Eastern, Central and Western Belts. The Eastern and Western Belts are composed of strata which strike parallel to the regional trend of the Nelson Syncline and dip steeply. The Eastern Belt, which has been briefly described on page 19, comprises rocks of the Pelorus Group. The Western Belt is composed of the Maitai Group and Glennie Formation of the Lee River Group (see geological map). The Central Belt is composed of the Red Hill Ultramafic Complex and the folded rocks of the Ben Nevis and Ellis Anticlines.

The major structural features of each belt are described separately below.

CENTRAL BELT.

W a r d s   P a s s   F a u l t

This previously undescribed fault separates Pelorus Group rocks of the Eastern Belt from those of the Ben Nevis Anticline. The trace of the fault is shown on air photographs as an intermittent linear feature in the heavily bush-covered country north-east of Red Hill. The fault is exposed in the tributary streams of the left branch of the Wairoa River as a brecciated zone at least 250 yards wide with vertical dipping sheets of soft fault-pug up to 50 feet wide concentrated near the centre. Fragments of hard, brittle mylonite occur in the brecciated rocks.

From Wards Pass it can be traced northwards in an almost straight line to the limits of the mapped area and if it continues, should intersect the Lee River near Anslow Stream. South of Wards Pass aligned streams and low saddles indicate the trace of the fault, toward the eastern margin of the Red Hill Complex. A major crush zone near the north-eastern contact of the Red Hill Complex can be traced for several hundreds of yards northwards towards Wards Pass and southwards along the eastern contact of the Complex, and is probably the major continuation of the Wards Pass Fault. The fault dips very steeply and strikes 025°. The trace is straight and the fault-pug unconsolidated, suggesting Tertiary or later movement. The mylonite in the crush breccia of the fault zone suggests that the fault may be older and was only reactivated in Tertiary times.

Ben Nevis Anticline

The Ben Nevis Anticline is separated from the Eastern Belt by the Wards Pass Fault. The Wether Formation in the Ben Nevis Anticline provides a good marker bed and can be traced from the north of Ben Nevis around the nose of the anticline as far as 442932 and from Wards Pass Fault to 458910. Between these points, the position of the formation is inferred from the general trend of bedding planes. The fold plunges southwestward at about 30°, and has a greatly thickened, steeply dipping southern limb. In the lower reaches of the Wairoa River, downstream from 480934, no younging directions or marker beds were obtained and the structure is not well known, but is consistent with the anticline defined by the Wether Formation.

The three dimensional form of the Ben Nevis Anticline cannot be wholly deduced from its shape north of the Red Hill Complex. The Pelorus Group very probably underlie the ultramafic rocks of the Complex and, as elsewhere, are likely to be essentially concordant with the basal contact of the Complex. Since on gravity data (p.91) it is established that the base of the Complex is approximately horizontal the Pelorus Group are likely to have the same attitude. An alternative, and preferred view (p.92) is that the basal contact dips west varying from 70° beneath the Western contact to 50° beneath the eastern contact of the Complex and is repeated by faulting (p.90). If so the Pelorus Group are likely to have the same attitude. This interpretation is illustrated in the geological cross-section. The strike of the Pelorus Group, and of the basal contact of the Complex therefore varies from approximately north-south beneath the Complex to approximately E-W north of Red Hill. This interpretation means that as the Pelorus Group dip under the Complex the Ben Nevis Anticline unrolls to form a north-south trending monoclinal structure.

The anticline is cut by numerous faults which, where observed as crush zones, are marked by the letter 'f' on the geological map (Map 1). Exposure below the bushline is poor but the larger faults are evident on aerial photographs as lineations defined by aligned streams and low saddles. The displacement of the Wether Formation is small and the faults appear to play a structurally subordinate role to folding. The faults in the north-east of the mapped area are identified largely from crush zones but displacement is unknown because no clear marker bed is available. The crush zones are about 20 yards wide and fault displacement may be large.

Small faults with displacement of a few inches or a few feet were observed in the field. They are generally parallel to the regional trend, steeply dipping and displace strata down to the east but are less common than similar faults in the Eastern Belt.

The Ellis Fault. The name Ellis Fault is given to the curved fault which defines the boundary between the Goat Formation and the Pelorus Group north of the Red Hill Complex. About 800 yards east of Mt. Ellis the fault is exposed in a small saddle between the left and right branches of the Wairoa River. It has been traced southwards to the Red Hill Complex and northwards for about a mile into the headwaters of the right branch of the Wairoa River. North of the saddle exposure is poor and the contact cannot be easily distinguished because the strata of the two Groups are parallel. South of the saddle, the contact is marked by a fault, which becomes increasingly obvious toward the Complex, where the strata of the Lee River Group strike into the fault at an angle of about 40°.

The fault dips about 50° to the south west near Mt. Ellis but about a quarter of a mile from the contact of the Red Hill Complex steepens to almost vertical. The fault arcs away from the Complex

Fig. 80. Northern contact of the Red Hill Complex.

striking initially north-west and swings to strike approximately north-south. The pronounced recurvature on the geological map (Map 1) is due to considerable change in height of topography combined with dips of about 50°. (Elevation of the fault near Mt. Ellis, 1500 feet higher than 1 mile north and south of Mt. Ellis).

The curvature on the Ellis Fault shows a conformity to the general structure of the Ben Nevis Anticline and it is probable that this reflects a close structural relationship (p.229).

Ellis Anticline

The Goat Formation immediately north of the Complex is folded in an asymmetrical, steeply plunging (60° to 80°) anticline which is named the Ellis Anticline. The southern limb strikes parallel to its contact with the Complex and the northern limb is folded into a broad synclinal flexure. The plunge of the fold is suggested by the steep dips of strata at the nose of the anticline in the headwaters of the Motueka River (fig. 17). A panorama of the area is given in figure 80.

The anticline is cut by faults of at least two ages. The earliest faults strike east-west and are cut and displaced by north-east striking faults which also displace the northern contact of the Complex. Many of the faults have been intruded by serpentinite (p. 66). The north-east faults are the continuation of a major set of faults which cut the Complex. These are described below.

Macroscopic structures in the Red Hill Complex.

Folds. The ultramafic rocks of the Upper Zone are finely layered. Macroscopically the layering defines fold-like structures. The most conspicuous is the antiform the hinge of which passes near Porters Knob (the Porters Knob Antiform). West of the hinge the layering generally dips steeply to the west or south west, and east of the hinge as far as the Motueka River layering dips at about 30° to the south-east. The fold is an asymmetric antiform with an axial plunge of about 25° to the south and with an axial plane dipping in an easterly direction.

Mafic dykes cut across both limbs of the fold with parallel orientation and therefore post-date folding (p. 95).

South of the Motueka River, on the Plateau, layering dips at about 40° to the south-east although some flat-lying layering occurs near the south-western contact of the Complex.

Very broad similarity of the Porters Knob Antiform and other fold structures in the Complex with the Ben Nevis Anticline suggest a possible structural relationship.

North-easterly striking faults. The mafic dykes in the Red Hill Complex are cut by a number of north-easterly striking faults. Some faults are however intruded by dykes, e.g. near Porters Knob. There were therefore two periods of faulting; one before and one after dyke intrusion unless faulting and dyke intrusion were contemporaneous. The latest north-easterly faults cut and displace the northern contact of the Red Hill Complex.

The faults are marked by a narrow zone of crushed and sheared serpentinite bounded by a zone about 25 feet wide of serpentinised peridotite. They are usually identifiable as lineations in aerial photographs but all those marked on the geological map have also been observed in the field. There are a large number of small faults with individual displacement of only a few inches or feet which are not shown on the map but their cumulative displacement may be very large. The smaller faults are downthrown to the east and movement is largely vertical. The displacement of the larger faults is known only for those which cut the northern contact. Most show apparent downthrow to the east. Like the smaller faults, movement is probably vertical. All faults dip steeply.

Lowther Stream Fault. The name Lowther Stream Fault is given to a major fault which cuts the western part of the Complex. It is well exposed in Lowther Stream as a 50 yard wide zone of crushed and sheared serpentinite. The fault has been traced northward from the south-western contact as far as the Motueka River and was also observed on the ridges and streams west of Porters Knob. The fault plane dips at about 70° east. The displacement is not known but is possibly the same as the north-easterly striking faults since it is subparallel to these.

North-westerly striking faults shown on the geological map cut the mafic dykes and are marked by zones of crushed serpentinite. The fault at 390850 gives an apparent sinistral displacement of the hinge, of the Porters Knob Antiform but this is the only known evidence of displacement. The faults are well shown on aerial photographs as aligned streams and saddles. The fault planes dip at about 65° to the south-west.

The north-easterly and north-westerly striking faults in the Red Hill Complex are probably of the same period of deformation as those of similar orientation which cut the Ben Nevis Anticline.

EASTERN BELT

Rocks of the Eastern Belt have a regional strike which swings from 360° in the south to 030° in the north, and dip steeply east. Local variation in strike and dip are common and rare reversals in younging direction show that some of the rocks are isoclinally folded. Most rocks young westwards and are overturned.

The hinge line of two isoclinal folds are exposed in Boulder Stream at 478832 and 472831. Axes of the folds plunge south at 20 to 30° and the eastern limbs are faulted against the western limb of inferred adjacent folds.

Numerous strike faults cut the rocks but in the heavily bush-covered country it is not possible to trace them beyond their exposure in streams. The larger faults have crush zones up to a few feet wide and dip at between 60° east and 80° west. The sense of displacement is probably similar to that of numerous smaller faults which have a displacement of only a few inches or feet. These are downthrown to the east, with dominantly vertical movement.

Despite the faulting and folding, of the Pelorus Group rocks, the Eastern Belt shows a broadly simple structural pattern of three parallel formations that can be traced for at least 15 miles along the strike (p. 19). In general, the rocks young westwards. The folding and faulting are therefore comparatively minor features superimposed on the regional structure of steeply dipping rocks that young to the west.

The inferred structure across the Eastern Belt is diagrammatically shown in figure 81. It is assumed that folding and faulting are closely related and the sense of deformation is the same over the whole width of the belt. Although locally the amount of displacement is comparatively small, the cumulative effect may be large, and the formations may be considerably wider than their true stratigraphic thickness (p. 18).

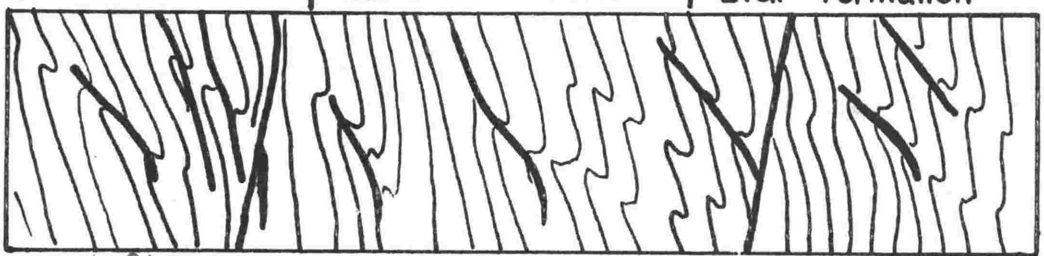


Fig ~~23~~ 81. Generallised cross-section of the Eastern Belt.

C h r o m e   S t r e a m   F a u l t

A 50 yard wide crush zone is exposed in Chrome Stream at 474797 but the fault has not been traced beyond the stream. The fault displaces a river-cut surface (with upthrow to the east) but not overlying gravels (fig. 82). It is therefore a young fault, but without very recent movement. It is probably a splinter of the Wairau Fault.

WESTERN BELT

Four north-easterly striking faults cut the Maitai Group and Glennie Formation, and displace formation boundaries.

One fault is exposed in Beebys Stream at 335796 as a 27 yard wide

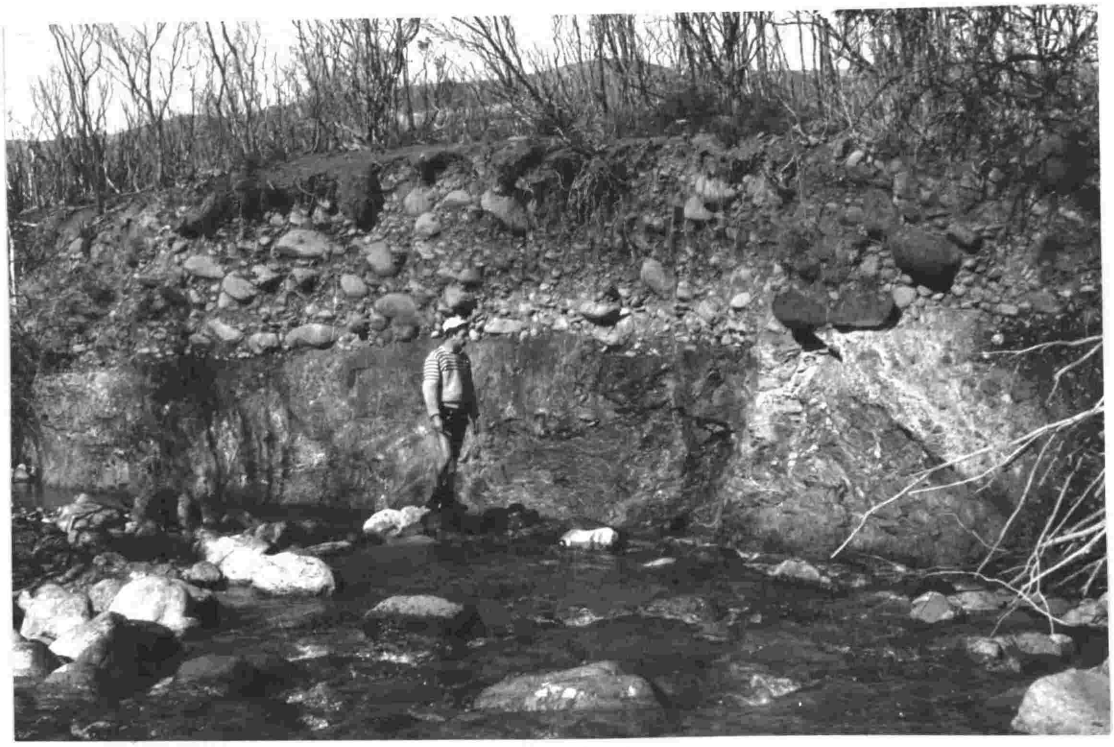


Fig. 82. Crush zone of Chrome Stream fault. The wedge shaped body on the right of the photograph consists of tilted gravels containing peridotite pebbles and crushed rock. Recent gravels lie on a platform cut on the crush zone and older gravels.

crush zone and contains small cold intrusions of serpentinite. It has been traced northwards over a step in the ridge between the Motueka River and Beebys Stream to join the marginal fault about 300 yards south of the gorge in the right branch of the Motueka River. Assuming only vertical movement the estimated horizontal displacement of 1000 feet between the Tramway/Greville boundary must be due to a vertical displacement of about 3000 feet.

Crush zones about 80 yards are exposed in the right branch of the Motueka River at 330826 and in Porters Stream at 342848, and a well defined trace joins the two. A crush zone in Ellis Stream at 360879 (fig. 83) may mark the continuation of the fault into the marginal fault, but no trace has been observed on the ground or on aerial photographs between Ellis and Porters Streams. The amount of displacement is perhaps about 6000 feet but the position of the displaced formation boundaries are not accurately known.

Crush zones are also exposed in the left branch of the Motueka River, Ellis Stream and at 370915 and 380950, and are related to a fault recognisable from aerial photographs. Although the sense of displacement is the same as that of faults described above, the amount of displacement is not accurately known.

In contrast to the north-easterly faults cutting the Red Hill Complex and Ben Nevis Anticline, the above faults of the Western Belt are downthrown to the west. They are generally straight and dip steeply, and their crush zones are soft and uncemented suggesting Tertiary or later age. Faults with similar trend and displacement near Nelson, cut Tertiary sedimentary rocks (Bruce, 1962).

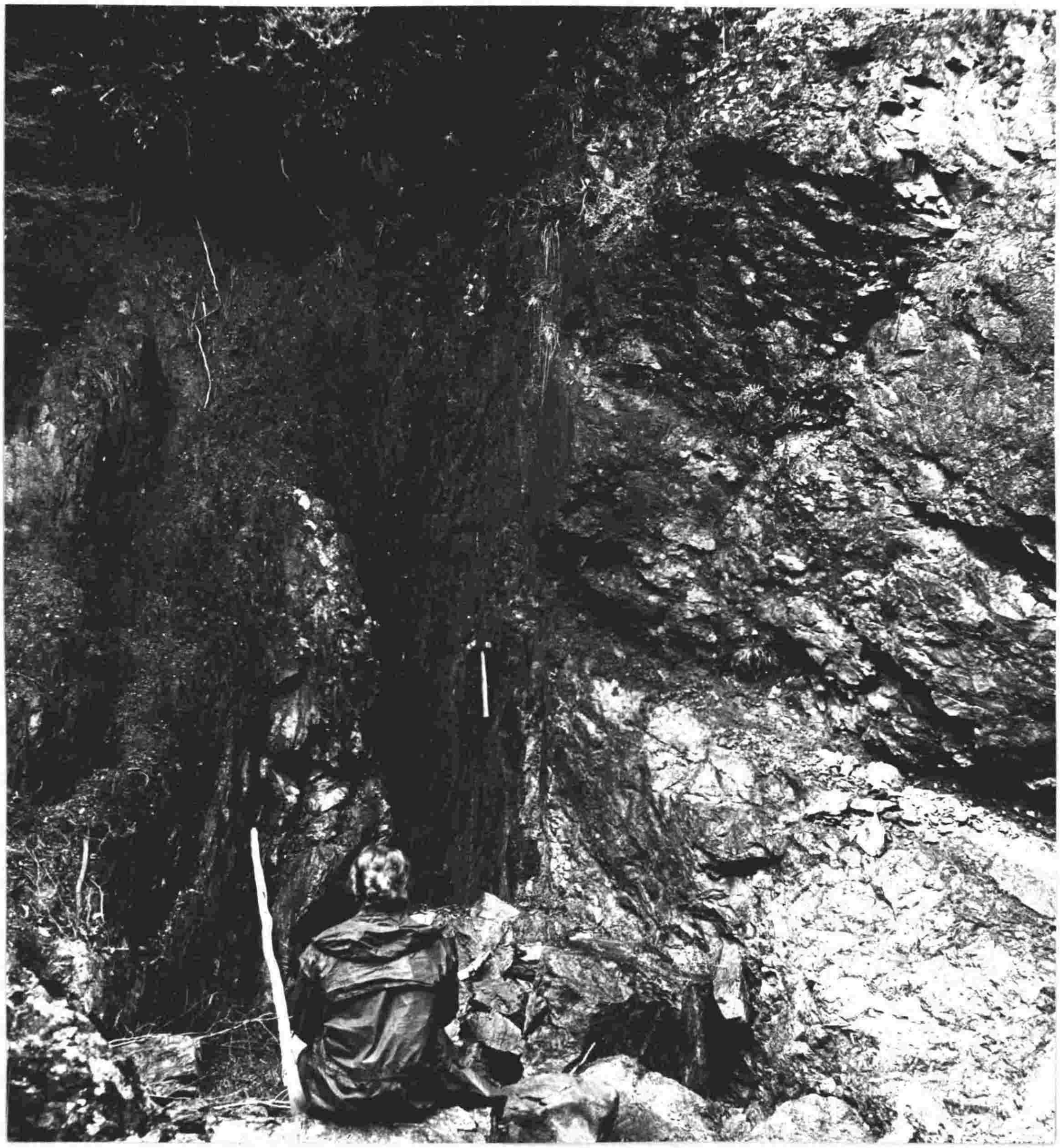


Fig. 83. Crush zone exposed in Ellis Stream. Enclosed in the crushed rock is a lensoid mass of serpentinite, probably a'cold intrusion'.

A number of small steeply-dipping strike faults with displacement of only a few feet are found in the Greville Formation. They are similar to the minor faults of the Eastern Belt in that movement has been dominantly vertical with downthrown side to the east. Bedding plane shear, and slaty cleavage are also common, but the rocks are much less deformed than the Eastern Belt rocks.

The most distinctive minor structures of the Maitai Group are zig-zag folds with an amplitude of about 100 feet. The axial planes dip at about 60° to the east, parallel to slaty cleavage.

WAIRAU FAULT

The Wairau Fault separates the rocks of the Upper Paleozoic from the Mesozoic Greywackes of the Torlesse Group. Recent movement is shown by stepped terraces and the fault trace is clearly seen in aerial photographs. Five miles east of the mapped area, the Wairau Fault cuts a group of river terraces and displacement indicates that horizontal movement was greater than the vertical, with upthrow to the east (Wellman, 1953). Transcurrent movement on the fault in the Red Hill area is not demonstrable, but a faulted surface on river gravels has a 10 foot vertical displacement with upthrow to the west. The Wairau Fault is the major northward continuation of the Alpine Fault (Suggate, 1963).

STRUCTURAL EVOLUTION OF THE RED HILL AREA.

SUMMARY OF DATA

(1) The stratigraphic succession is

Maitai Group.	Basal limestone followed by sandstone and laminated argillites.	Age; Late-Permian
Glennie Formation.	Sheet volcanics and minor clastic sediments.	Age; Probably Mid-Permian.
Goat Formation.	Pillow lavas and volcanic breccias and minor clastic sediments.	Age; Probably Early-Permian.
Pelorus Group.	Volcanic derived wackes.	Age; Probably Carboniferous.

All formations are concordant.

(2) Ultramafic rocks.

The Red Hill Complex is a lens-like body about 12,000 feet thick at its thickest part, which thins to 3000 feet.

The Complex is part of an extensive sheet which extends at least as far as Dun Mountain. The dimensions of the sheet are :- length 40 miles, breadth 12 miles or more, thickness 2000 to 12,000 feet.

Ultramafic rocks form an almost continuous outcrop from D'Urville Island to the Wairau Fault.

The Ultramafic rocks intrude the Lee River Group throughout their length.

In Otago and Southland a similar belt exist in precisely the same stratigraphic horizon. (See Grindley, 1956, Waterhouse, 1964).

In places the ultramafic rocks have metamorphosed the Lee River Group to high temperatures (Challis, 1965b). They have possibly metamorphosed the Maitai Group at Dun Mountain (p. 77) though the evidence is limited and not unequivocal.

(3) Structure within the Red Hill Complex.

The structure of the Complex is that of a broad sheet cut and duplicated by a number of faults.

The ultramafics are finely layered near the stratigraphic top of the sheet.

Layering is of at least two generations. At least one generation is of metamorphic origin, and is accompanied by foliation and recrystallisation.

Layering is inclined and defines macroscopic and mesoscopic fold-like structures.

The metamorphic layering formed at temperatures in excess of 650°C.

(4) Dykes.

A large number of mafic dykes cut the ultramafic rocks of the Red Hill Complex.

They cut the limbs of the macroscopic folds with parallel orientation and therefore post-date folding.

The dykes are cut by north-easterly (and north-westerly faults) which also cut the northern contact of the Complex.

Rare dykes intrude along an earlier set of north-easterly faults.

The dykes were intruded as a liquid magma of temperature about 1000° to 1200°C.

After intrusion some dykes were metamorphosed to hornblende microgabbros and leucocratic veins.

The mafic dykes show chemical affinities (based on two analyses) with the volcanic rocks of the Glennie Formation. Both volcanics and dykes are broadly tholeiitic in character.

(5) Metamorphism.

In general the Maitai, Lee River and Pelorus Groups are only slightly altered. The widespread occurrence of authigenic pumpellyite indicates only low grade metamorphism - lower than green schist.

Some Glennie Volcanics are thermally metamorphosed adjacent to the ultramafic rocks. Pyroxene hornfels indicate a temperature of emplacement of the ultramafic rocks of about 1200°C according to Challis (1965b).

(6) Structures outside the ultramafics.

The structure of the Eastern and Western Belts is simple, consisting of parallel steeply dipping strata. These define the eastern limb of the Nelson Syncline.

The Nelson Syncline is very probably of Upper Mesozoic age because it is similar to the Southland Syncline (Grindley, Harrington and Wood, 1959) which contains folded Jurassic strata.

The Ben Nevis Anticline, the Ellis Fault, the Ellis Anticline, and north-easterly and north-westerly striking faults of the Red Hill Complex can be regarded as belonging to a familiar structural pattern.

This structural pattern is consistent with an axis of main compression approximately E.S.E.-W.N.W.. This agrees with the likely stress pattern during development of the Nelson Syncline.

The major faults in the western belt differ from others of similar orientation in being down-thrown to the west instead of the east. They are probably of the same age as Tertiary faults of similar displacement and orientation in the Nelson Area.

(7) Comparison of fold structures.

The Porters Knob Antiform and other fold structures within the Red Hill Complex may also be regarded as developing under conditions in which the major axis of compression was approximately E - W. It is possible therefore that folds within the Complex developed during the same period as the Ben Nevis Anticline and allied structures.

PRINCIPLE DEDUCTIONS

From some of the above data two principle deductions can be made regarding the age of the ultramafic rocks and some of the structures of the area.

Age of the Ultramafic Rocks.

The ultramafic rocks are clearly younger than the metamorphosed rocks of the Lee River Group. No definite information is available on an upper limit of age, however, the following points are of importance in this respect.

The ultramafic rocks intrude the Lee River Group wherever they outcrop. For more than 80 miles in the Nelson Ultramafic Belt and for a similar distance in the Otago Ultramafic Belt the ultramafic rocks nowhere (with one possible exception, see page 77) cut younger strata. Also the association of spilitic volcanics and ultramafic rocks is a common one in orogenic zones. Several writers have considered this to indicate a distinct suite to which the term Alpine-type has been given by Benson (1926) and the term 'ophiolite' by many European writers (e.g. Kundig, 1956). The writer considers that the constant association of ultramafic rocks with spilitic volcanics in the New Zealand Upper Paleozoic Belts strongly indicates a relationship both in space and time.

The time relationship is further supported by the chemical similarity of the mafic dykes and the Lee River Group volcanics. Also Grindley (1956) has shown that albite-dolerites intrude the ultramafic rocks in the Otago Ultramafic Belt again suggesting ~~an~~ approximate contemporaneity.

The weight of evidence is therefore considered to indicate that the Red Hill Complex is Permian in age.

The ultramafic rocks were probably intruded as a sill into the volcanic rocks or possibly emplaced as submarine flows intruding superficial sediments and overlain by later volcanics. (See later page 265).

Age of formation of the Ben Nevis Anticline and allied structures.

The conformity of the structural pattern of these structures to the Nelson Syncline strongly suggests the same period of development, i.e. Upper Mesozoic. The Ben Nevis Anticline is therefore regarded as a subordinate fold of Upper Mesozoic age. Most faults in the Central Belt are probably of the same age and their formation therefore post-dates emplacement and cooling of the ultramafic rocks.

Alternative hypotheses for the structural Development of the Red Hill area.

Two alternative hypotheses can be put forward to account for the structural development of the Red Hill area. The first of these assumes structural unity throughout the Central Belt. i.e. the folding of the ultramafic rocks was contemporaneous with folding of the Pelorus Group. The second assumes that folding of the ultramafic rocks occurred at an earlier period and bears no direct relationship to the Upper Mesozoic folding.

Hypothesis I.

The ultramafic rocks were emplaced as a sill in Permian times and folded by Upper Mesozoic deformation.

Two alternatives within the general bounds of this hypothesis must be considered. (i) The metamorphic layering, foliation, and recrystallisation occurred during regional deformation in the Upper Mesozoic. (ii) These structural and textural features were inherited from an earlier period of deformation.

Alternative (i) may be discounted on the following grounds.

It was shown that the metamorphic layering could not have formed at temperatures of less than 650°C (p.175). If it is postulated that the generation of such structures occurred after emplacement and cooling of the ultramafic rocks it must be further postulated to sustain the argument that the Complex was reheated during folding. Such reheating if it occurred did not, however, affect the enclosing sediments which show only low grade metamorphism. It must therefore have been restricted to the ultramafics. The only agency to cause reheating would appear to be the deformation caused by the folding itself. Yet Turner and Verhoogen (1960) state (p. 661).

> "The work of nonelastic deformation is dissipated as heat. It can be shown, however, that even the most drastic deformation in rocks under stresses reaching their breaking strength is accompanied by a rise in temperature of no more than 10° or so. Similarly, heat generated by viscous flow under stresses such as are likely to develop within rock masses is negligible. Goguel has computed the heat equivalent of the mechanical energy involved in the deformation of some sections of the Jura Mountains and of the Alps, and finds it to be of the order of a few calories per gram, quite insufficient to produce any metamorphic effect. Nor are such effects visible in some of the intensely deformed sediments of these mountains. There is, in general, very little detailed correlation between grade of metamorphism and degree of folding, or between times of deformation and of recrystallisation."

They go on to point out that the, "association of pseudotachylite veins in mylonites suggests that frictional heat, when rapidly generated can indeed bring about fusion of the rocks affected." But such "high temperatures have not been maintained sufficiently long to permit chemical reconstitution of the deformed rock". (p. 661).

It thus appears highly unlikely that temperatures of even 100°C higher than the enclosing sediments could be produced let alone the required minimum of 350°C (assuming the low grade metamorphism indicates temperatures no higher than 300°C).

Alternative (ii) Stated in its simplest terms the hypothesis requires that during the development of the Nelson Syncline the Red Hill Complex and the enclosing strata were subjected to regional triaxial stress in which the Greatest Principle Stress (G.P.S.) was at right angles to the synclinal axis. In the Red Hill area the direction of the G.P.S. may be taken as at right angles to the strike of strata of the Eastern and Western Belts i.e. in the north of the area W.N.W.-E.S.E. and in the south of the area approximately E-W (Map IV).
With the G.P.S. acting in these directions subordinate folds may develop with axes parallel to the synclinal axis. The orientations of the Porters Knob Antiform and the Ben Nevis Anticline agree fairly well with this orientation. Furthermore it may be reasonably postulated that the mass of the heavy Red Hill Complex had an influence on the adjacent strata during folding sufficient to cause a general plunge of fold axes under the ultramafic mass. Such a postulate could account for the southward plunging Ben Nevis Anticline and also the southward plunge of folds within the Complex itself.

The strike faults which are down-thrown to the east could be interpreted on this hypothesis as high angle thrusts; this accords reasonably well with observed dips. The north-westerly striking faults, such as the Ellis Fault and the faults within the Complex could be interpreted as sinistral wrench faults. This again agrees with the few observed displacements. The Ellis Anticline could also be interpreted as formed during the same period of deformation. Its E.N.E- W.S.W. striking axial plane and steep plunge accounted for by arguing that due to its peculiar position between the Complex and the northwards extension of the ultramafic sheet it suffered local compression due to the influence of the Complex to regional stress.

The relationship of the mafic dykes to Upper Mesozoic deformation is not clear. The dykes were presumably intruded along tensional cracks and therefore their orientation should indicate possible lines of tension. Such lines of tension may develop regionally in the directions indicated on Map IV. A relationship begins to be feasible for those dykes near Red Hill but is not shown at all on those dykes further south. Therefore the orientation of the dykes cannot be due entirely to regional compression. Local tensional cracks may be expected however parallel to the hinge of folds. The dykes in the far north-west of the Complex (about 390920) and also the north-easterly striking dykes at Porters Knob may be so interpreted, but the vast majority of dykes occur in a very gently arcuate wedge shaped zone, concave westwards, that extends diagonally across the Complex. The curvature of this zone is the opposite to that expected if the dykes were related to regional folding. Also the dykes near Red Hill, and in the headwaters of Diorite and Ellis Streams, are almost at right angles to the axes of folding.

No single postulate relating the dykes to regional compression therefore accounts satisfactorily for their orientation and it is therefore considered that the dykes were not intruded during Upper Mesozoic deformation. They probably represent a separate defomational episode altogether. Since the dykes post-date folding of the ultramafics they also post-date Upper Mesozoic deformation according to Hypothesis I.

Using the data given earlier and the consequences that follow from Hypothesis I the chronology below may be constructed.

7. Faulting: (North-easterly and north-westerly faults which cut the mafic dyke).
6. Metamorphism of the dykes.
5. Intrusion of the dykes.
4. Faulting of the Porters Knob Antiform, later intruded by dykes.
3. Folding of the ultramafics and enclosing strata.
2. Deformation of the ultramafics to give metamorphic layering.
1. Emplacement of the ultramafics.

(Note: This chronology ignores the north-easterly striking faults in the Western Belt which are down-thrown west. These are presumed to be Upper Tertiary in age (p.219) and not related to the Upper Mesozoic defomation.)

1 and 2 may follow immediately or be contemporaneous. Both 3 and 4 occur very much later but may occur together. 5 must be separated by a considerable time lapse to account for reorientation of G.P.S. but may be followed immediately by 6. 7 would probably have to follow 5 after a long interval to allow reorientation of the G.P.S. This hypothesis therefore probably requires four periods of deformation and at least two separate phases of igneous activity.

Hypothesis II

The ultramafic rocks were emplaced as a sill in Permian times and layering was folded during emplacement. The folded sill was subjected to regional stress during Upper Mesozoic folding during which the Ben Nevis Anticline and allied structures developed but did not itself fold.

The major advantage of this hypothesis is that the dykes could have been intruded before Upper Mesozoic folding. Apart from this the NE, NW faults and the Ben Nevis Anticlines may be ascribed to Upper Mesozoic deformation.

The chronology would now be:-

7. Faulting of the Ultramafic Complex and Ben Nevis Anticline.

6. Folding of the Ben Nevis and Ellis Anticlines.

5. Metamorphism of the mafic dykes.

4. Dyke Intrusion.

3. Folding and faulting of the ultramafic rocks.

2. Deformation of the ultramafic rocks to give metamorphic layering etc.

1. Emplacementof the Ultramafic rocks.

1, 2 and 3 may follow immediately or be contemporaneous. 4 may follow very shortly after 3 since deformation of ultramafics was not regional and only local re-orientation of stress directions is required. 5 may follow 4 immediately or even occur contemporaneously (see below). 6 and 7 occur in the Upper Mesozoic. This hypothesis requires two separate periods of deformation and one period of igneous activity.

Preferred Hypothesis.

Hypothesis II is preferred because

(i) Four periods of deformation and two periods of igneous activity must be postulated for hypothesis I and only two separate periods of deformation and one phase of igneous activity are necessary for hypothesis II.

(ii) There is no evidence of metamorphism to account for the alteration of the dykes in post-Upper Mesozoic times apart from the dykes themselves. Hypothesis II accounts for the metamorphism by postulating their intrusion into hot, but cooling ultramafic rocks. (p.259).

The main difficulty of hypothesis II in comparison with Hypothesis I is the similarity of the folds within the Complex with those of the Ben Nevis Anticline. Such coincidence is not however unlikely. The controlling factor of most orogenic deformation is likely to be the orientation of the original geosyncline.

Wellman (1956) has shown that the geosynclinal axis is parallel to the axis of the later Nelson Syncline and hence to the fold axis of the Ben Nevis Anticline. Any intrusion of the ultramafic rocks is also quite likely to bear a close relationship to the geosynclinal axis. Any folding that may have then occurred is likely to conform to the orientation of the geosynclinal axis and hence be parallel to the later Nelson Syncline and the subordinate Ben Nevis Anticline.

Consequent upon hypothesis II it is postulated that during the Upper Mesozoic deformation no folding occurred within the Red Hill Complex but the rocks were faulted and tilted. This resulted in tilting of the base of the Complex down to the west by slip along the strike faults.

Some tilting of the complex to the south may also have occurred during Upper Mesozoic deformation resulting in the plunge on the fold axes of the Porters Knob Antiform. This is however speculative and the precise orientation of the folds within the Complex before the Upper Mesozoic is not known. The general parallelism of the layering on the western limb of the Porters Knob Antiform to the strata of the Western Belt suggests that that limb was horizontal.

SUMMARY.

The age of the Ultramafic rocks is very probably Permian. The age of the Ben Nevis Anticline and related faults is probably Upper Mesozoic.

If both these deductions are correct then folding of the Ultramafic layering and dyke intrusion probably occurred before the Upper Mesozoic. The most likely period of deformation and dyke intrusion is during emplacement of the Ultramafic rocks for it is only at this period that it is possible to reasonably postulate deformation accompanied by high temperatures.

STRUCTURE AND PETROLOGY

OF THE

RED HILL COMPLEX, NELSON

PART IV

PETROGENESIS.

Petrogenesis

In this section the origin and extent of metamorphic layering, the mode of intrusion of the mafic dykes and the mode of emplacement of the Dun Mountain Ultramafics are discussed.

Origin of Metamorphic Layering

Metamorphic layering, such as that featured in figures 52,55, 56 and 63 is produced by closely spaced, multiple veins. This is clearly demonstrable only in those outcrops in which the metamorphic layering can be seen to cut across an earlier structure, but as will be shown later there are reasons to suppose that the single layering visible in many outcrops has formed in the same manner, with earlier layering completely obliterated. In discussing the origin of metamorphic layering, therefore, close attention is given to the origin of those veins which cut earlier structures.

The origin of the metamorphic layering given below is discussed in two parts, firstly the role of replacement reactions in the formation of veins and secondly the role of deformation and accompanying recrystallisation. The dual roles of deformation and replacement in the formation of metamorphic layering is illustrated by describing in detail the features of the Porters Knob outcrop. Finally the extent of metamorphic layering and its possible mode of development are discussed.

Role of Replacement Reactions

Bowen and Tuttle (1949) suggested that many of the apparently intrusive features within peridotite complexes such as dunite veins and dykes cutting peridotite may be due to "hydrothermal (pneumatolytic) rearrangement of material taking place largely within the mass itself..." (op. cit. p. 460). Their experimental studies on the system $MgO-SiO_2-H_2O$ led them to consider that ~~that~~ at temperatures above 650°C water undersaturated in silica may desilicate harzburgite and form dunite, and silica saturated water may result in the formation of enstatite from olivine with the development of pyroxenite veins cutting dunite.

The reaction involved is

$$\text{olivine} + \text{silica} \rightleftharpoons \text{enstatite}$$

and given an aqueous phase with which to transport silica then by change in pressure-temperature conditions it may be possible to drive the reaction either to the right or left. Similarly if CaO and Al_2O_3 are soluble in a water phase and can be readily transported then feldspar and diopside may form by replacement of olivine or by direct deposition from solution. However, the olivine/enstatite reaction is of greatest importance.

The 'rearrangement' reactions described by Bowen and Tuttle are essentially 'replacement reactions'. Hess (1938) described in detail dykes, veins and irregular masses of dunite cutting pyroxene-bearing peridotites in the Stillwater Complex and he later (Hess 1960) interpreted these as formed by replacement. Lipman (1964) also describes dykes of dunite of probable replacement origin. Three examples from the Red Hill Complex are cited below:

(i) In the outcrop illustrated in figure 85 three intersecting tabular dunite bodies are shown near the top of the outcrop. The earliest is part of a layered sequence of dunite and feldspathic harzburgite which dips gently to the left of the photograph, the second is a steeply dipping vein of dunite which is parallel to foliation cutting the sequence and the last formed is a dunite dyke which dips to the right of the photograph. The dyke must have formed by replacement as there is no displacement of the earlier structures. Probably the second vein of dunite is also of replacement origin.

(ii) The original planar structure of laminated harzburgite in figure 86 does not show the degree of displacement necessary if the dunite was wholly of intrusive origin.

(iii) Large bodies of coarse-grained dunite, surrounded by layered or foliated harzburgite were mentioned earlier (p.125). They are strikingly similar to the bodies described by Hess (1938) from the Stillwater Complex and like those have probably formed by replacement.

The replacement origin of some dunite is obvious in these outcrops and numerous other examples have been observed; veins of dunite cutting protoclastic harzburgite (p. 135) and foliated peridotite (p.185) have probably formed in the same manner.

The replacement origin of pyroxenite veins and pegmatites is not so readily apparent as they seldom form very thick tabular bodies. But pyroxenite formation is the complementary process to dunite formation, involving the addition rather than the removal of silica, and it is therefore probable that these veins too, have formed by replacement.

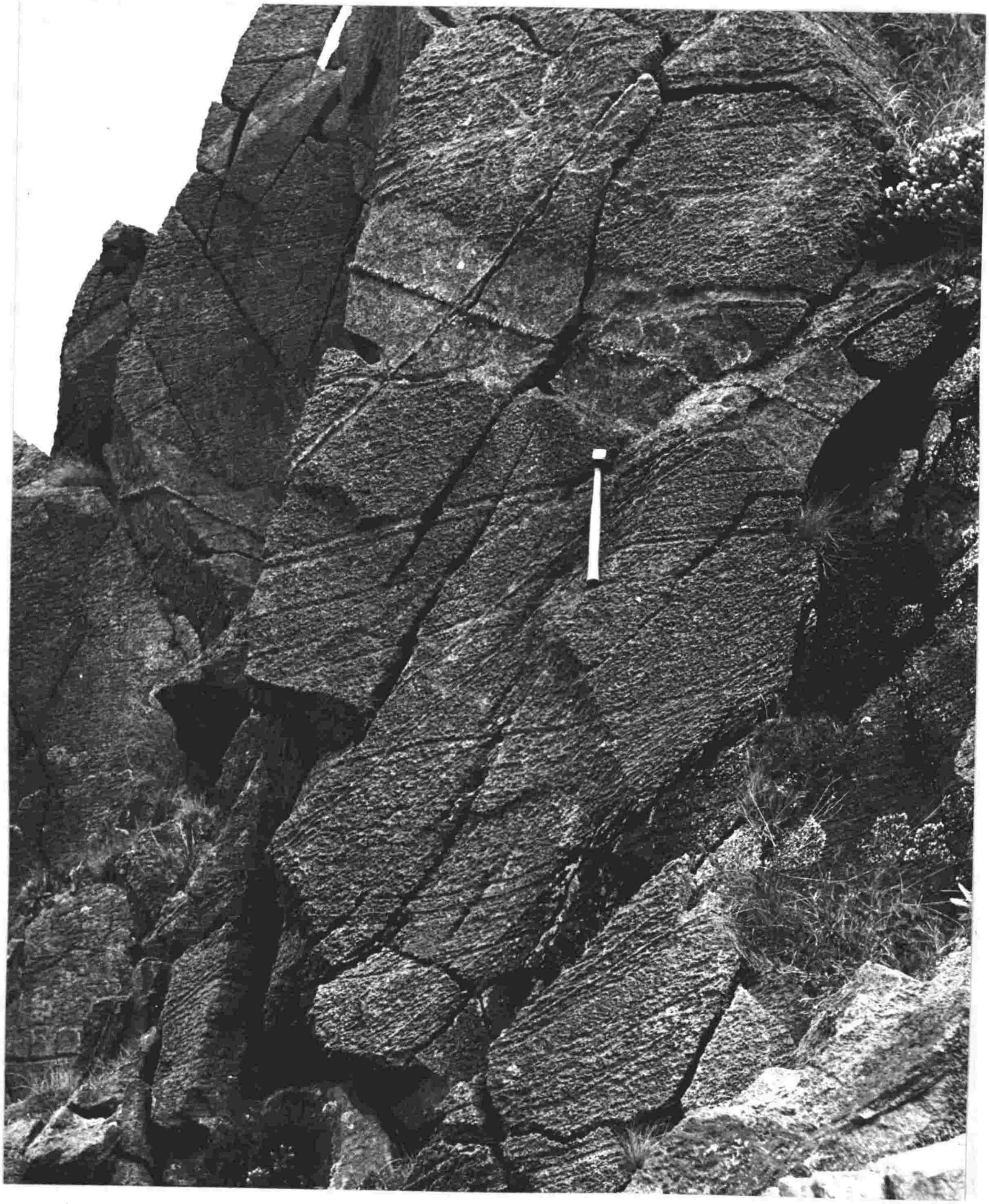


Fig. 85. Replacement of feldspathic harzburgite (finely etched) by dunite (smooth). Above hammer three tabular bodies of dunite intersect. As there is no displacement of the earliest of the dunite bodies (flat lying) no dilation can have occurred. The last formed dyke (dipping to the right of the photograph) must have formed by replacement and probably the others as well.
Note: (i) Pyroxenite veins traversing the centre of the dunite dykes. (ii) Foliation dipping parallel to one of the dunite dykes.

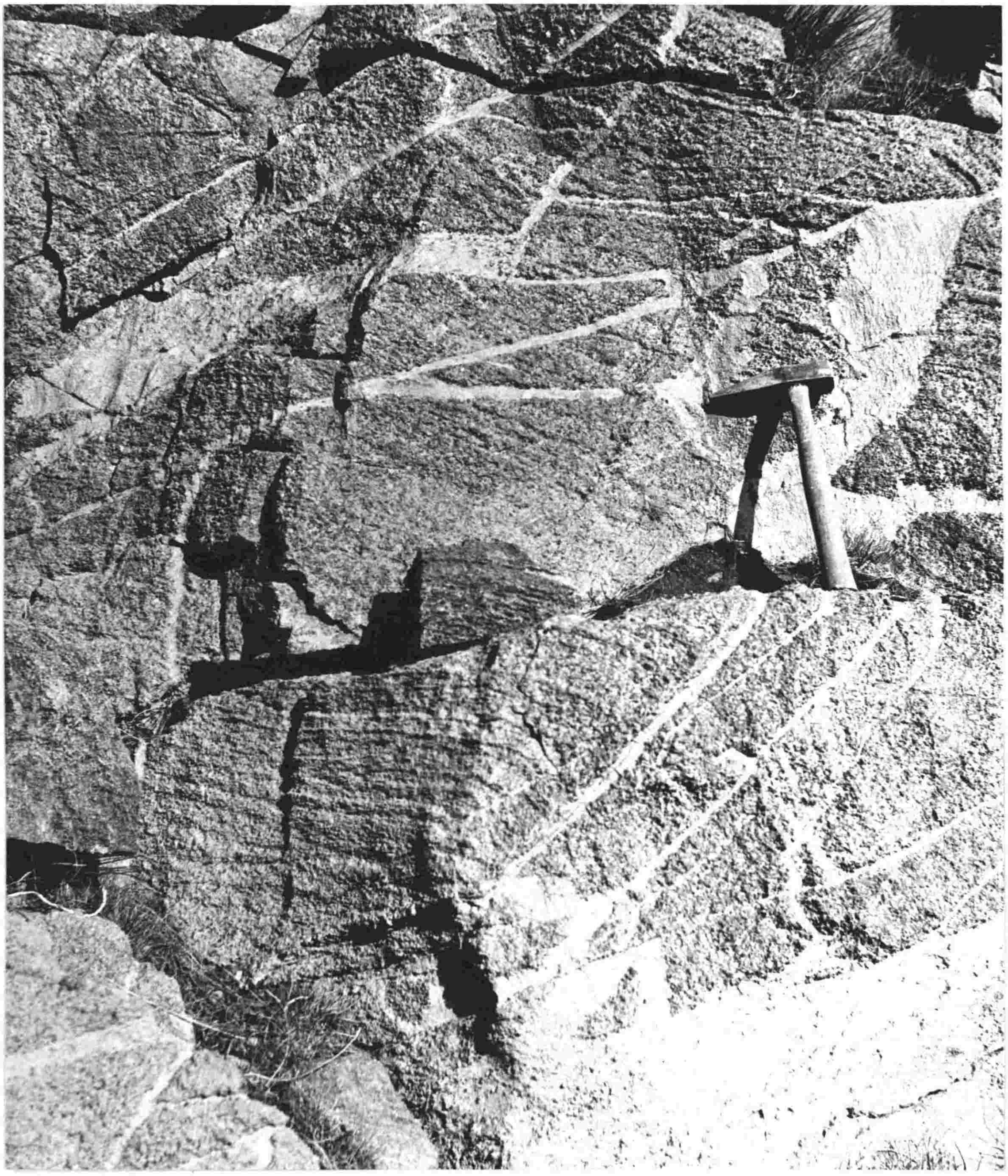


Fig. 86. Replacement of laminated harzburgite (dark) by dunite (light coloured and smooth).

Order of replacement

Where veins of dunite and pyroxenite occur together, the pyroxenite veins cut, and are later than the dunite. This is shown in figure 55 where dunite veins dipping to the right are cut by subparallel laminae of orthopyroxene, and also in figure 85 in which each of the dunite dykes are cut by central veins of pyroxenite. In some outcrops the sequence dunite-pyroxenite is repeated, with early-formed dunite-pyroxenite veins cut by later dunite-pyroxenite veins. An example can be seen in figure 56, and has been widely observed throughout the Complex.

CONTROLLING FACTORS OF REPLACEMENT

As suggested by Bowen and Tuttle (1949) the principal agent of replacement is probably a waterphase. Water vapour streaming through cracks and fissures in the virtually crystalline rocks will redistribute the rock components, in some regions forming dunite and in others orthopyroxene, clinopyroxene and feldspar. In the reaction suggested by Bowen and Tuttle (1949) silica is the mobile component, and the main problem is to determine under what conditions the reaction proceeds in a specific direction.

Bennington (1956) suggested on thermodynamic grounds, that the reaction is susceptible to pressure; olivine tended to form at high load pressure at the expense of enstatite if silica could be transported out of the system. The necessary activation energy of the reaction as well as local pressure gradient he suggested, would be provided by local shear stress, with olivine developing along the shear plane and silica migrating into regions of low stress. There are many examples of dunite occurring along zones which, because of displacement of intersecting structures, may have been formed at high shear stress, but there is abundant evidence that shear is not a necessary prerequisite (e.g. fig. 85).

Smith (1958) suggested that dunite formed at high temperatures by percolating water vapour undersaturated in silica, and enstatite formed as the temperature fell and water vapour became oversaturated in silica. Thus, in a cooling mass of ultramafic rock dunite may be expected to form during early stages of cooling and pyroxenite veins at a later stage. This explanation can only account for a single set of dunite and cross-cutting pyroxenite veins - it does not explain the repetition of the sequence (see above).

The main factor controlling the direction in which the reaction proceeds is considered by the writer to be pressure, although it is agreed that temperature also plays an important part.

The solubility of silica in water increases with temperature and pressure, and Kennedy (1950) remarked that at high temperatures, changes in pressure are marked by major changes in solubility. Small fluctuations of pressure therefore, such as may be induced in a free water vapour phase within an actively deforming body of crystalline rock, may result in quite large changes in silica solubility and hence control the direction in which the replacement reaction will proceed. It is suggested that during active deformation, regions of high pressure developed in the rocks, and water vapour escaped through available fracture zones to regions of low pressure. The first rush of water is likely to be at higher pressures than later flow, and hence is undersaturated with respect to silica. As the pressure dropped saturation point was passed and silica reacted to form enstatite. The sequence may have been repeated over and over again in the same place.

The numerous examples of a dunite dyke traversed down the centre by a pyroxenite vein (e.g. fig. 85, and an example shown by Lipman, 1964, plate 2b) demonstrate this mechanism. The pyroxenite veins probably mark the site of the fracture along which the water flowed, forming first the dunite and then, with falling pressure, the pyroxenite.

Summary of the Role of Replacement Reactions.

Most veins and metamorphic layering involving orthopyroxene and olivine are probably of replacement origin. Such veins could be formed throughout the cooling history of the ultramafic rocks if water can circulate through the rocks at varying pressures. Also as the rocks cooled silica dissolved in the water phase may be deposited and orthopyroxene formed by replacement of olivine. The formation of dunite veins on the other hand appears to demand locally induced over-pressures of the water phase such as may be induced during active deformation of the ultramafic rocks.

Many veins, however, may be fissure fillings, deposited by circulating waters in tensional cracks during a drop in pressure or temperature. Veins of anorthosite, for instance, could be expected to be of such origin for it is difficult to imagine replacement of olivine by feldspar due to their total dissimilarity in lattice structure and content.

Diopside, on the other hand, may possibly be formed by replacement either of olivine or orthopyroxene.

ROLE OF DEFORMATION IN THE ORIGIN OF METAMORPHIC LAYERING.

Orientation of foliation and metamorphic layering.

Foliation and metamorphic layering are generally parallel whereever they occur together and neither change orientation much over distances of several hundreds of yards. For instance, near Porters Knob metamorphic layering and an associated foliation (with rare lamination) have parallel orientation (dip of about 25° to 30° to the south east) for more than two miles along the ridge top. Such consistency of orientation implies that the development of layering is induced by a major defomational episode. Also in many cases the metamorphic layering occupies surfaces of shear which show considerable displacement of earlier structures across the layering. Presumably parallel foliation also developed along such surfaces. Lineation which lies in the plane of foliation presumably indicates the direction of displacement but lineation is not sufficiently common to be of much use in regional mapping.

Recrystallisation. The development of some xenomorphic-granular texture has been ascribed to recrystallisation of original protoclastic texture (pages 140 - 145). The driving force of the recrystallisation has been suggested to be the instability of the fine grained matrix with a tendency to form equi-dimensional grains. Fabric studies and field observations, however, suggest that recrystallisation is even more extensive causing the reorientation of fabric elements within a rock and the partial or total obliteration of earlier structures such as veins (p. 153). The force inducing recrystallisation in these cases is suggested to be tectonic; under conditions of long sustained stress, permanent strain may occur by continuing recrystallisation of the rock.

Flow structure illustrates the latter process. Flow structures are common in Alpine-type peridotites (Thayer, 1963) and several have been previously described from the Red Hill Complex (p. 159). These are attributed to viscous flow occurring during emplacement of the ultramafic rocks. The mechanism of flow probably involved extensive recrystallisation and plastic deformation. It tended to disrupt the veins formed by replacement; lamination forming foliation (fig. 90); and layering forming schlieren (fig. 58). Less intense deformation tended to destroy the sharp contacts of veins and to redistribute minerals over a wider zone. Continued flow over a long period may possibly form a homogeneous peridotite. In this respect deformation is a complementary process to hydrothermal activity; one a differentiation process and the other an homogenising process. Well developed layering probably involves both processes.

During deformation both fractures and plastic flow occurred, e.g. in figure 56 a pyroxenite vein (S_1) presumably formed by replacement along a fracture zone can be seen to have been plastically deformed by flow parallel to metamorphic layering (S_2). The vein is flow folded, thinning where subparallel to, and thickening where at right angles to, the S_2 layering. This is an illustration of the concept of 'rheidity' (Carey, 1953). During short-term severe stress rocks behave as brittle solids but during sustained low intensity stress they may behave as a viscous fluid.

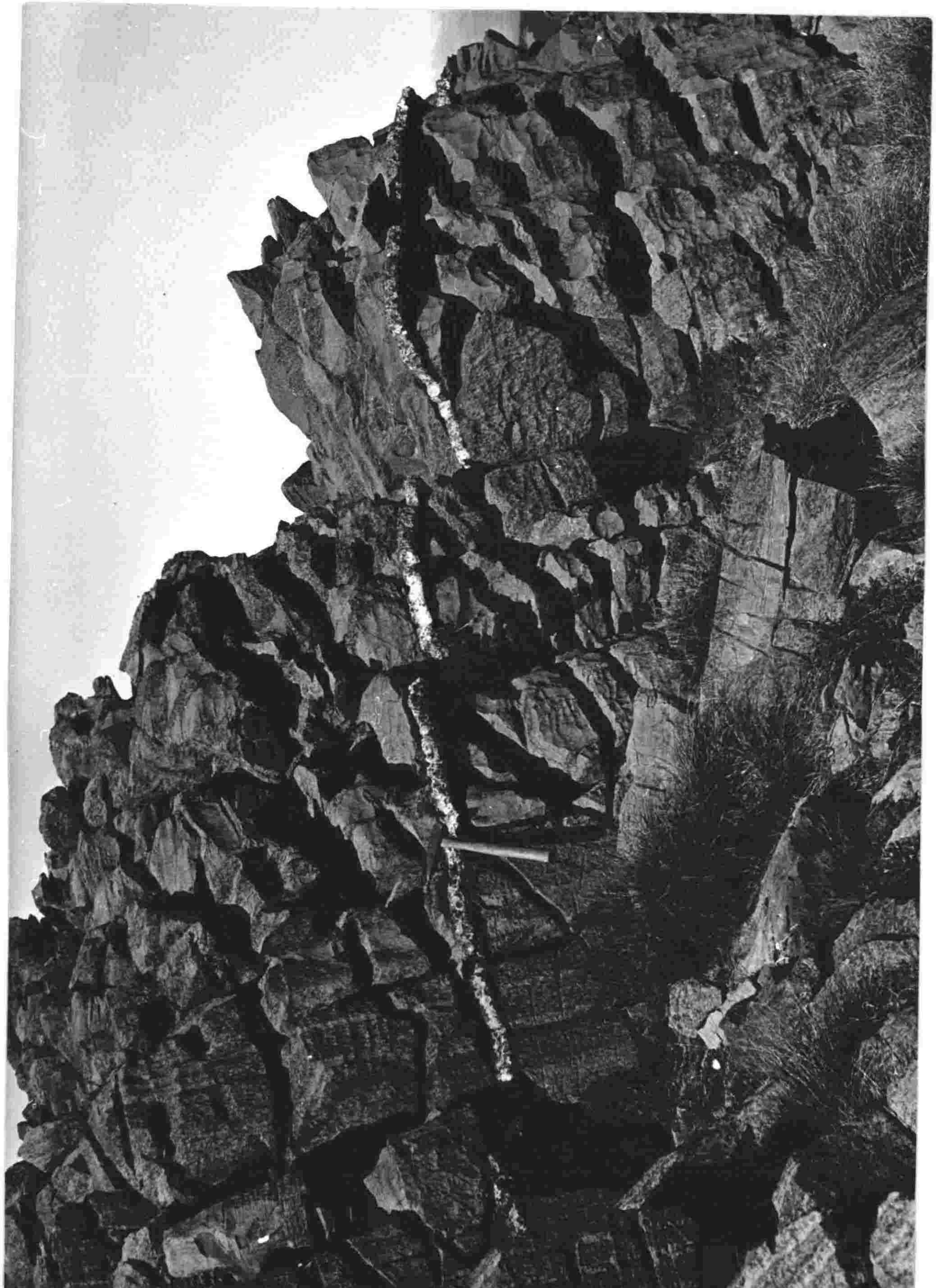


Fig. 87 Anorthosite vein cutting dunite-harzburgite layering.

Fig. 90. Disruption of pyroxene lamination (upper right) into foliation of pyroxene aggregates (see (ii) fig 41) in dunite.

PORTERS KNOB OUTCROP

The dual roles of replacement and deformation is illustrated in the Porters Knob outcrop (figs. 91, 92, 93). Two sets of layering are present, a plastically deformed early layering (S_1) composed of dunite and laminated harzburgite cut by metamorphic layering (S_2) which dips at about 30° to the right of the photograph.

The S_2 layering is compound, composed of subparallel dunite, orthopyroxene and feldspar-diopside veins. Dunite occurs as a thick tabular body (fig. 92 by hammer) and a few thin veins (fig. 92). The S_2 dunite veins are cut by veins of orthopyroxene forming a laminated structure (fig. 91) and a 1 inch thick vein (fig. 92, by hammer). Feldspar-diopside lamination is shown in lower left of figures 92 and 93.

The S_2 dunite veins occupy surfaces of shear, and the S_1 layers have been shear folded. The central dunite vein has been formed by replacement as relict S_1 structures can be traced across the vein showing that neither dilation nor offset has occurred and it is reasonable to expect that other S_2 structures have formed similarly. The effect of flow parallel to S_2 is shown in figure 93. Original S_2 pyroxene lamination has been disrupted and the pyroxene grains dispersed homogeneously through the olivine matrix to form a foliated harzburgite, and thick pyroxenite veins have lost their sharp boundaries and grade into the adjacent layers.

The sequence of development of metamorphic layering appears to be as follows:

(i) Original layering subjected to shear with the formation of dunite veins along, in most cases, planes of movement.

(ii) Orthopyroxene veins developed. These were formed subparallel to earlier dunite veins giving a laminated structure or comparatively thick individual veins. No displacement apparently occurred across these veins.

(iii) Feldspar-diopside lamination may have formed contemporaneously or later.

(iv) Further shear was generally concentrated along the thick tabular body near the top of the outcrop and also caused development of the shear fold in the original layering. Continued movement, probably the result of recrystallisation, tended to disrupt an initial (S_2) lamination to form foliation and redistribute the orthopyroxene (and olivine?) grains throughout the peridotite.

Three features of this outcrop deserve special emphasis.

(i) The metamorphic layering converges to the right of the outcrop obliterating the original folded layering that is so conspicuous on the left of the outcrop (compare figs. 91 and 93).

(ii) The metamorphic layering in this outcrop is parallel to and can be traced laterally into single layering developed further to the east, i.e. on the east limb of the Porters Knob Antiform.

(iii) The metamorphic layering exhibits the same general characteristics as the single layering developed elsewhere in the Red Hill Complex.

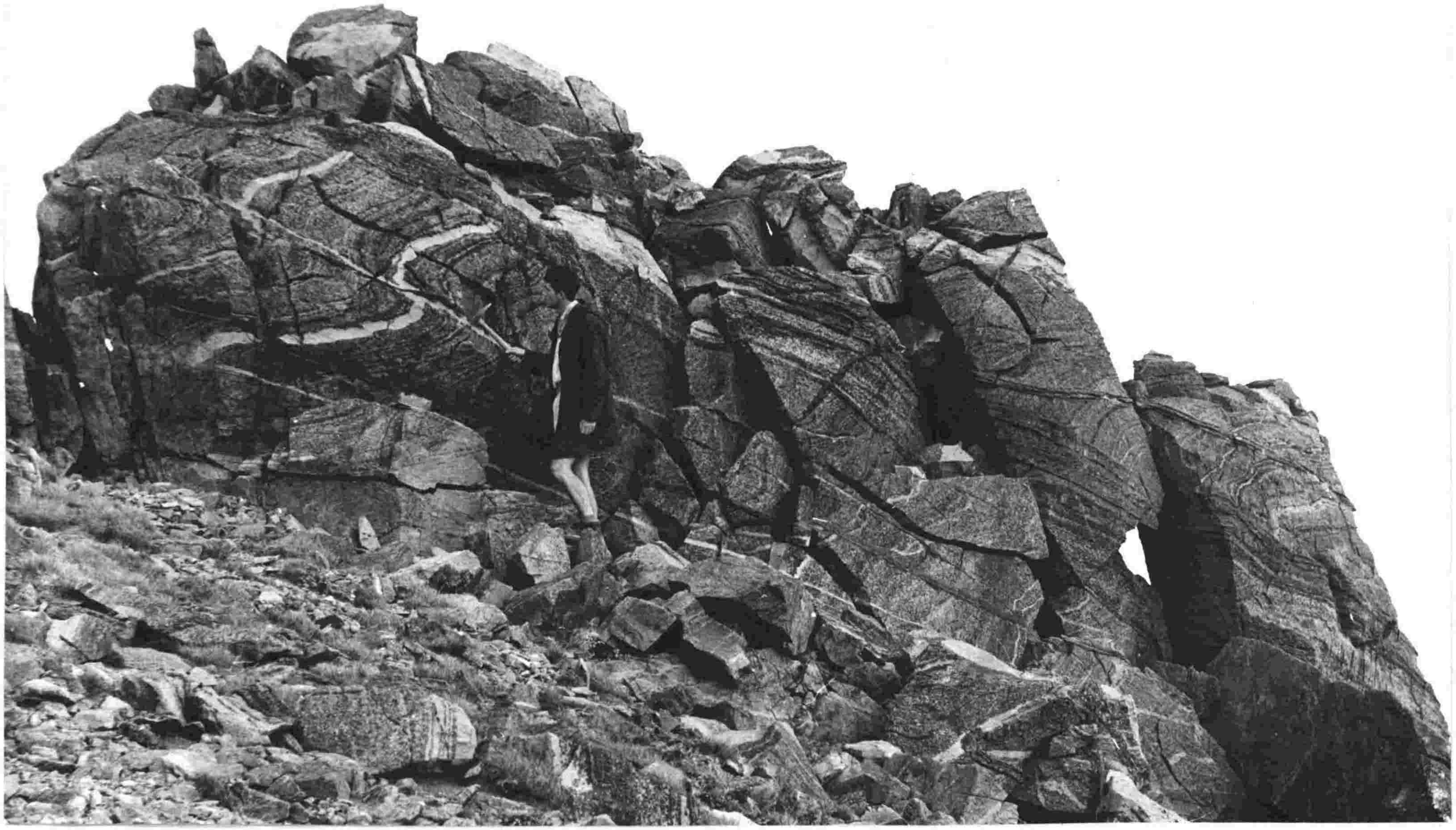


Fig. 91. Porters Knob outcrop at 384816. 'Early' layering (S_1) dips to the left of the photograph and is composed of dunite (light) and laminated harzburgite (dark). 'Late' layering (S_2) dips to the right of the photograph and is also composed of dunite (e.g. by hammer) and laminated harzburgite (above figure's head). 'Late' layering converges to the right of the outcrop obliterating 'early' layering.

Fig. 92. Detail of figure 91.
Note: (i) Pyroxene vein (black) cutting 'late' dunite vein in centre of the photograph.

(ii) Diopside-feldspar lamination parallel to and below the central dunite vein.

(iii) Very thin dunite veins between lamination and central dunite vein. S_1 layering has been offset.

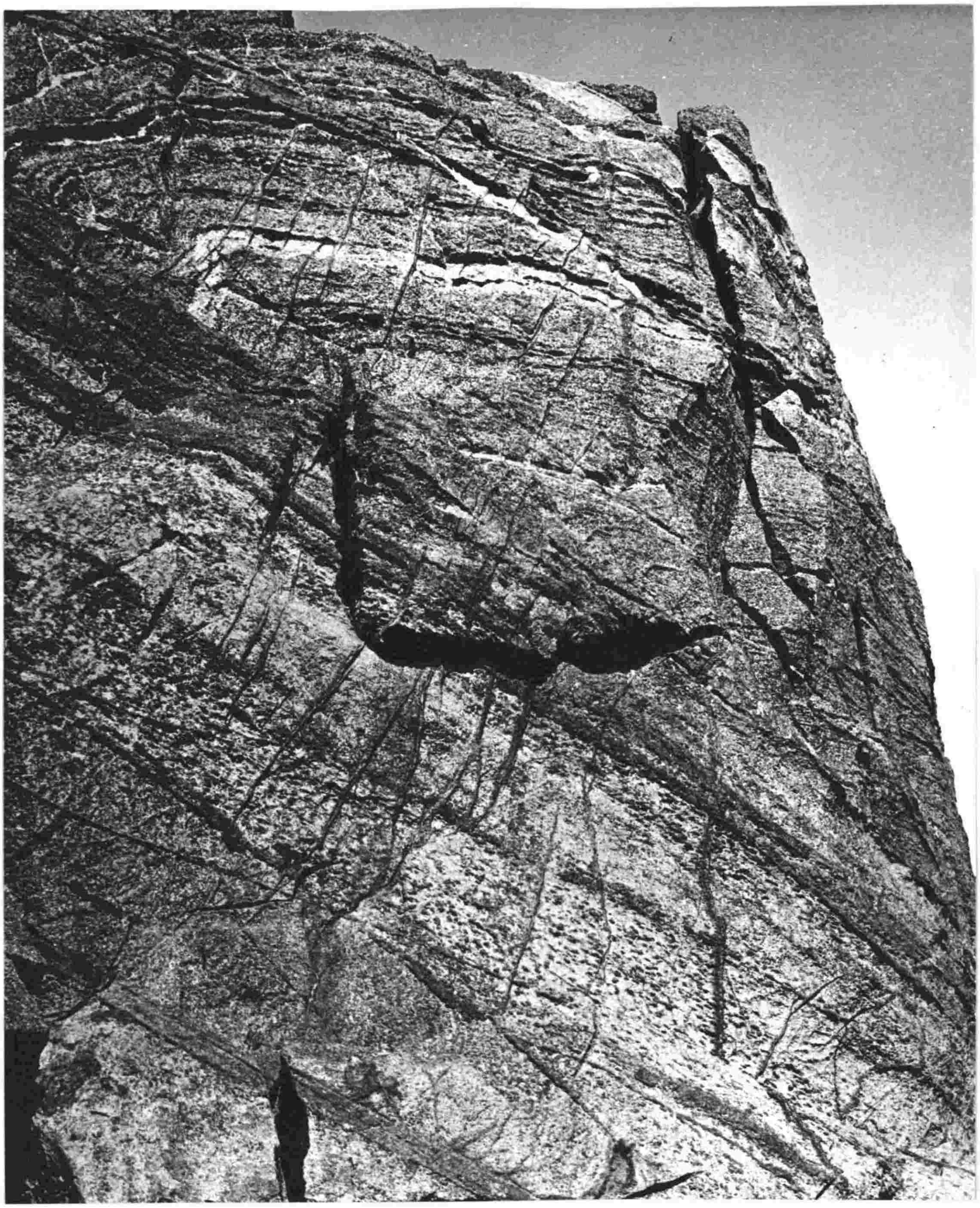


Fig. 93. Detail of figure 91. Laminated layer in lower left is the continuation of the one shown in figure 92.
Note: (i) Almost complete obliteration of 'early' layering in lower left of photograph.

(ii) Disruption of laminated harzburgite giving foliated harzburgite.

(iii) Diffuse boundaries to pyroxene veins lower centre of photograph.

EXTENT OF METAMORPHIC LAYERING.

It is evident that replacement and deformational processes may give rise to a metamorphic layering of the same general characteristics as layering developed throughout the Red Hill Complex. How much of this layering then is of metamorphic origin? Challis (1965a) concludes that some layers are formed by crystal sedimentation and so the question becomes how much layering can be definietly ascribed to one or other of these processes? It is important to emphasise that different criteria have been used to establish the differing proposed origins. Challis's identification rests largely on petrographic features and argues by analogy that similar textures in the Red Hill and the Stillwater Complexes suggest similar origin. The identification of metamorphic layering on the other hand rests on the existence of cross-cutting mesoscopic structures. Both types of layering may be present.

Metamorphic layering, however, is considered to be of very wide extent throughout the Complex. This view is based on the following observations:

(1) Metamorphic processes can give rise to a layered structure in rocks that is at least broadly similar to that present elsewhere.

(2) In the Porters Knob outcrop almost complete obliteration of earlier layering has occurred. Therefore the single layering which has occurred elsewhere could be of metamorphic origin.

(3) The single layering on the eastern limb of the Porters Knob Antiform is probably all of the same origin since it is of closely similar

orientation over the whole of the limb, dipping approximately 30° to the S.E. (although the strike swings from N.S. to E. W. in different parts of the limb). The metamorphic layering developed on the hinge of the Antiform is parallel to and in many cases (such as that of the Porters Knob outcrop) can be traced into the single layering of the eastern hinge. It is very probable therefore that all layering on the eastern limb of the Porters Knob Antiform is of metamorphic origin.

(4) The layering on the western limb of the Antiform is indistinguishable ~~in~~ in kind from that of the eastern limb. Also although not traced step by step it apparently continues around the nose of the Antiform (see Map II) into the layering of the eastern limb.

(5) Although crosscutting structures in which the S_2 planes (foliation and layering) are parallel to the single layering developed elsewhere on the western limb are rare (see for example Map II, and figure which shows feldspathic foliation parallel to the layering of the eastern limb), these examples strongly suggest that layering too is of metamorphic origin.

(6) The structures developed on the northern end of the Plateau are identical to layering developed in the Porters Knob Antiform. Crosscutting structures however, although common are not so well developed as, say, the Porters Knob Outcrop.

(7) It was earlier noted that lineation near the top of the Basal Zone lies in the plane of foliation developed nearby which is in turn parallel to a weakly developed layering present slightly higher in the Complex (p.152). This also suggests that that layering is of metamorphic origin.

(8) The dunite veins which cut protoclastic harzburgite at 412885 are parallel to layering developed nearby ~~and~~ which is therefore probably metamorphic (p. 135).

(9) Finally, the wide development of metamorphic layering throughout the Upper Zone is suggested by the general requirements of compositional development of the rocks. It was earlier suggested that apart from a trace of feldspar the bulk composition of the Upper Zone was similar to that of the massive and uniform Basal Zone. Although evidence of this kind is of uncertain reliability due to the extreme difficulty in determining the average composition of layered rocks it suggests that the Upper Zone has formed by metamorphic differentiation of rocks of the same composition as the Basal Zone.

MODE OF DEVELOPMENT OF METAMORPHIC LAYERING

Observations.

(1) The ultramafic rocks were very probably intruded as a sill into flat-lying sediments. This implies considerable horizontal movement of the peridotite 'magma'.*

(2) Layering is formed along, or parallel to, planes of shear.

(3) Layering is only developed widely in the upper parts of the sill.

(4) Layering is parallel over considerable distances.

(5) Some layering, notably that on the western limb of the Porters Knob Antiform is subparallel to overlying sediments and therefore, before Upper Mesozoic deformation, was probably approximately horizontal.

(6) Metamorphic layering developed when the rocks were hot.

(7) Challis (1965b) has shown that the ultramafic rocks were intruded when very hot.

(8) Folding of the Porters Knob Antiform and allied structures probably occurred during or shortly after emplacement of the sill.

It is concluded that

(a) Layering developed during emplacement of the sill because (4) makes it unlikely layering could be an inherited structure: deformation accompanying emplacement would be expected to completely disrupt and contort earlier structures. The 'early layering' which is cut by the metamorphic layering however, is possibly such an inherited structure (p.194).

(b) Layering probably developed along subhorizontal planes of shear.

* The term peridotite "magma" is used in the same sense as in Turner and Verhoogen (1960,p.310) "Ultramafic igneous material injected at high temperature without prejudice as to whether it was largely liquid or largely crystalline at the time of intrusion."

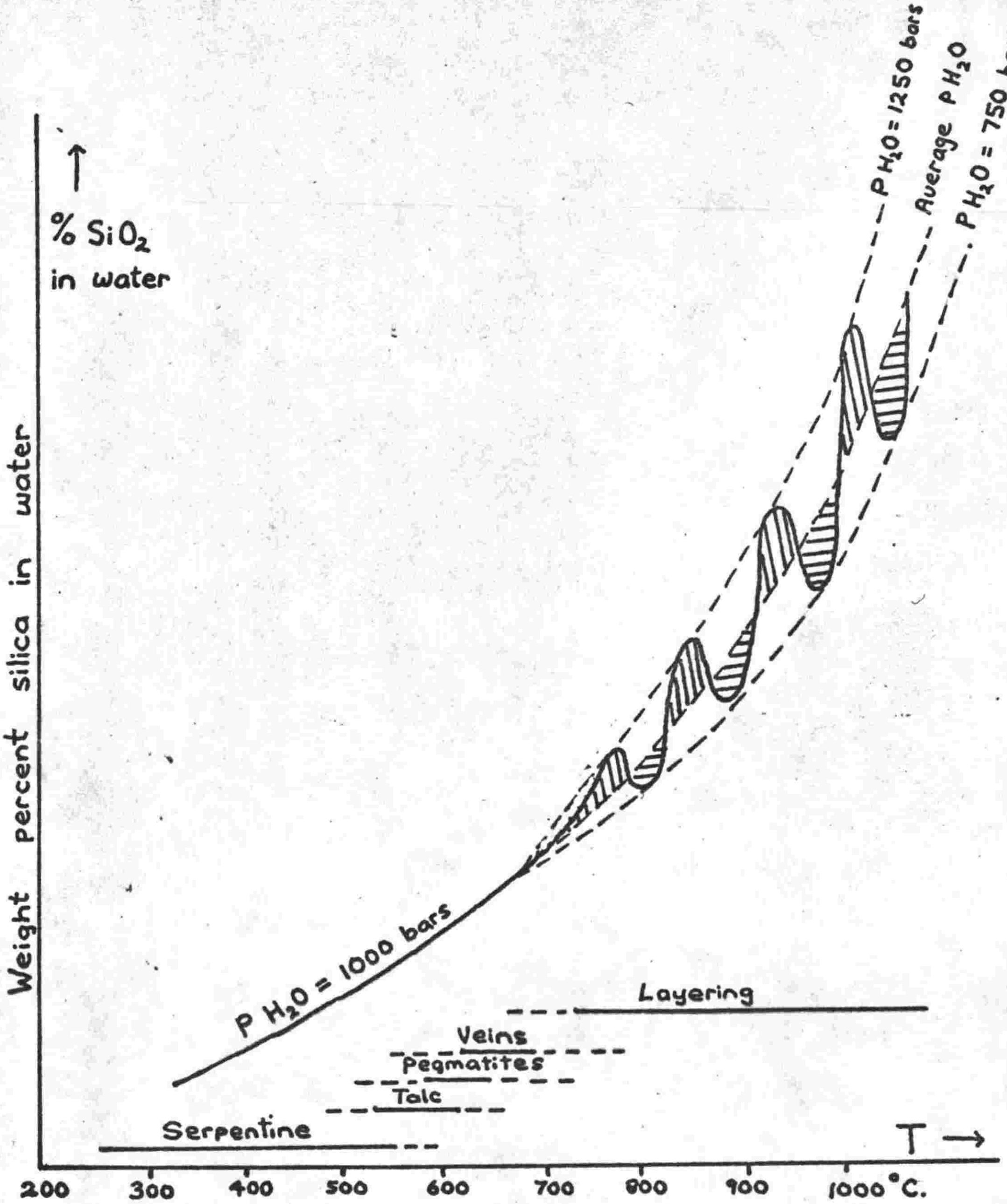


Fig. 88. Relationship between pegmatites, veins and layering.

Layering formed during emplacement at high temperatures and under active deformation. Oscillating pressures resulted in alternate dunite and pyroxenite vein formation. Areas shaded with horizontal lines indicate olivine + silica → enstatite; vertical lines enstatite + olivine → silica. Pegmatites formed after emplacement during falling temperature. (see text)

SUMMARY

Much layering in the Red Hill Complex is considered to have developed by a combination of plastic deformation and replacement reactions occurring during emplacement of the ultramafic rocks. No clear distinction can be drawn between veins and metamorphic layering - both may have developed by the same processes, but while pegmatites and some orthopyroxene veins probably developed following emplacement of the ultramafic mass layering is considered to have developed during active deformation when the rocks were still very hot. The postulated relationship is illustrated in figure 94.

This figure shows weight per cent silica dissolved in water plotted against temperature. The figure is diagrammatic but the curve follows in a general way the isobaric curves of Kennedy (1950). Temperature was probably falling constantly during and following emplacement and thus the abscissa can be regarded as indicating passage of time from right to left. During emplacement the water pressure at any particular point in the deforming sheet would have been oscillating about an average value. Consequently the water vapour would have become alternately supersaturated and undersaturated,and asit moved from place to place along planes of shear it transported silica and effected replacement reactions. This could have continued throughout emplacement of the ultramafic sheet.

After emplacement,during cooling, pyroxene pegmatites may have developed along tensional cracks. The observation that coarse-grained pegmatites are invariably younger than finer-grained veins of the same composition (p.172) is considered to indicate that the finer grained veins developed during the period of oscillating pressures. At this stage the degree of supersaturation of the water vapour is likely to be large

while the coarse-grained pegmatites formed after active movement, when the degree of supersaturation was probably small.

Mode of emplacement of the Mafic Dykes.

Hypabyssal rocks of the mineral composition of the hornblende and pyroxene microgabbros are most unusual yet chemically they are similar to dolerites, the most common of dyke rocks. Johannsen (1937, p. 305) lists only one hypabyssal rock of similar composition to the pyroxene microgabbros. This was described by Coleman (1914) from the Sudbury district, Ontario and named Sudburite (35 per cent bytownite, 25 per cent hypersthene, 25 per cent augite, 15 per cent magnetite) but was later considered to be a metamorphic rock (Thomson, 1935). Also dyke rocks of the composition and texture of the hornblende microgabbros (and of undoubted igneous origin) are exceedingly rare. But the assemblage hornblende-labradorite-quartz is common in metamorphosed dolerite dykes (e.g. Sutton and Watson, 1951). Also, large bodies of plutonic rocks of these compositions are common.

The differences in mineralogy between plutonic and hypabyssal rocks of basaltic composition are generally explicable in terms of different physical conditions of crystallisation; pigeonite persists in the more quickly cooled hypabyssal rocks but inverts to hypersthene during slow colling of the plutonic rocks; hornblende develops under high water pressures and slow cooling of the plutonic rocks but may only be partly developed in the more quickly cooled hypabyssal rocks. Thus while of undoubted hypabyssal nature, (they are usually only 6 inches to 2 feet wide) the mineralogy of the pyroxene microgabbros is not characteristic of rapid cooling but rather slow cooling of plutonic or metamorphic rocks.

It is considered that these features are due to intrusion of the dykes into ultramafic rocks which were at temperatures close to the crystallisation temperature of the dykes, perhaps about 500-800°C. It is considered most unlikely that the thin pyroxene microgabbros (containing hypersthene) could have been intruded into cold (about 200°C. or less) peridotite which is one of the best conductors of heat, for rapid chilling would result and hypersthene would not be expected to crystallise.

The banded and leucocratic rocks are also readily explained by intrusion of dykes into hot country rock. The almost complete segregation of felsic and femic components in some dykes is unlikely to have been effected during quick cooling and certainly the leucocratic veins extending up to 100 yards from the parent dyke would not be expected. If, however, the temperature of the dyke was maintained at about 600°C. it is conceivable that at 1000 bars water pressure a mixture of quartz and andesine may be sufficiently mobile to flow readily during deformation and intrude the crystalline hornblende in the segregated dykes or intrude joints or fissures in the peridotite. If the dykes were intruded into the hot ultramafic rocks as suggested, such deformation would be expected to occur in the waning stages of emplacement and folding of the Complex.

Although many dykes were probably intruded into hot country rock others were emplaced at lower temperatures. This is suggested by the hornblende microgabbro dyke near the base of the Complex and described on p. 99. The relatively wide chilled zone and the contrasted mineralogy between the contact and the more slowly cooled centre of the intrusion suggest intrusion into comparatively cold peridotite.

Dyke intrusion probably occurred over a long period therefore commencing shortly after the emplacement and folding and continuing throughout the cooling of the ultramafic rocks. Their orientation suggests that they were emplaced as gently inclined sills. West of Porters Knob the orientation of the dykes and layering is approximately the same although the dykes appear to cut across layering at an angle of a few degress. The layering there is subparallel to the strata of the Maitai Group and like the strata was probably tilted during the Upper Mesozoic. Thus the dykes and layering were probably subhorizontal before tilting. As the dykes generally lie in a zone extending from the stratigraphic top of the Complex to near the Basal Zone, the dykes were probably gently inclined. The dykes may have been intruded along planes of shear similar to those developed during flow of the mobile but virtually crystalline 'peridotite magma'. These planes, it must be postulated, formed during the waning stages of deformation following emplacement of the ultramafics.

MODE OF EMPLACEMENT OF THE RED HILL COMPLEX

Challis (1965a) suggests that the Red Hill Complex represents a deep level magma chamber of a Permian volcano. The layering is considered to be primary layering and the ultramafic rocks the accumulates formed at the base of the chamber. Certain evidence given by Challis, however, appears to the present writer to indicate that the ultramafic rocks, even if formed by gravity sedimentation underwent some intrusion into their present position as a peridotite 'magma'.

(i) The primary layering described by Challis (1965a) dips at 10° to the east and is situated about 300 yards or less from the contact which dips at 80° to the west. It is most unlikely that this layering could have escaped the regional tilting movements which caused the dip of the contact. Therefore it is probable that in the Permian the primary layering was standing almost vertical.

This suggests that the ultramafic rocks were considerably deformed, probably during their emplacement.

(ii) The metamorphic contact described by Challis (1965a) and (1965b) occurs near the western boundary of the Red Hill Complex. This is at the stratigraphic top of the Complex. At this part of the contact the ultramafic rocks are in direct, unfaulted contact (page 81 this thesis, and Challis, 1965a) with the metamorphosed volcanics. This strongly suggests that the ultramafic rocks were emplaced into their present position as hot rocks composed, at least in the top parts of the sill, of peridotite. This suggests emplacement as a peridotite 'magma'.

The argument for intrusion of the Complex as a peridotite 'magma' is also suggested by the foregoing argument on the origin and mode of development of the metamorphic layering.

Although emplaced as a sill (because of the metamorphosed rocks on its upper contact) the Red Hill Complex may have been emplaced at considerable depth or under only a thin cover of volcanic material. In Maitland Stream the ultramafic sheet is covered by only about 1000 feet of volcanics and metamorphic rocks. At one end of this section the contact is intrusive and unfaulted and at the other it is probably a disconformable contact with the Wooded Peak Limestone. Faulting in this section is probably not important. Elsewhere the Glennie Formation is very much thicker; up to 4800 feet in Porters Stream. This suggests that some erosion may have occurred before deposition of the Maitai Group and the volcanics were at one stage thicker than at present. Thus if the ultramafics were intruded after deposition of the Glennie Volcanics they may have been intruded at considerable depth, say a mile or so overburden.

However, the mafic dykes indicate that after intrusion of the ultramafics there was a period of basaltic activity during which, it is possible, much of the Glennie Volcanics may have been erupted. Therefore the Complex may have been emplaced under only a thin cover of volcanic material.

There is no decisive evidence to indicate which of these hypotheses is correct but two lines of evidence seem to the writer to indicate shallow rather than deep emplacement of the ultramafics.

(1) The Glennie Volcanics differ from the Goat Volcanics in form - sheets as against pillows. They also contain Atomodesma prisms suggesting relatively shallow water deposition whereas the abundant volcanic breccias and pillow lavas of the Goat Formation and the interbedded clastic sediments suggests deep water deposition. Also the Glennie Formation contains a distinctive red and green breccia consisting of green volcanic fragments set in a hematized matrix. This may be expected to form under subaerial conditions. Although none of this evidence is conclusive it suggests that the Glennie volcanics were deposited at or near sea level while the Goat volcanics were deposited in deep water. If the ultramafics pre-date the Glennie volcanics these points are readily explained as due to the lifting of the sea floor during intrusion. As the Red Hill Complex is at least 3,000 and as much as 12,000 feet thick the sea floor must have been lifted by a proportional amount. It would be expected that considerable differences would occur between those volcanics deposited before, and those after, emplacement.

If, however, the ultramafics were intruded after deposition of the Glennie volcanics these points must be explained by some separate and unrelated postulate. If the Glennie volcanics are shallow-water deposits and pre-date the ultramafics, it is difficult to see how they could have been preserved following intrusion for most probably they would have been lifted above wave base and consequently rapidly eroded.

(2) The Glennie volcanics are, at the few places where it is possible

to measure dips, parallel to the overlying Maitai Group. Yet intrusion of the ultramafics apparently involved considerable deformation causing folding of the metamorphic layering. It would be expected that superincumbent strata would be similarly deformed. If the Glennie volcanics were deposited after intrusion their concordance is readily explained. If deposited before intrusion it must be maintained that folding was intraformational as did not affect the overlying strata. This, though possible, is difficult to imagine.

For these reasons shallow intrusion, before a substantial part of the Glennie volcanics were erupted appears to be the most reasonable postulate.

The ultimate origin of the ultramafic rocks may have been as accumulates at the base of a Permian volcano, as suggested by Challis (1965a), and later intruded into the present site, or possibly formed from a liquid peridotite magma from which crystal settling took place to form a primary layering. Such a peridotite magma may exist at geologically possible temperatures (Clark and Fyfe, 1961).

The Red Hill Complex is therefore considered to have been emplaced as a near surface submarine intrusion of crystalline peridotite'magma' which metamorphosed superficial deposits. The cover including intruded clastic sediments was probably no more than 1000 feet thick and possibly much less in places. The ultramafics were later overlain and intruded by basaltic magma. Limestone was deposited as reefs on those parts of the volcanics near sea level and later in turn overlain by sandstone and argillites.

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